Sinkholes and the Engineering and Environmental Impacts of Karst

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Cover Photo:
A first look inside the National Corvette Museum sinkhole, which consumed eight rare Corvettes upon its collapse. The museum Skydome was built over a large cave passage which suffered a catastrophic roof collapse on the morning of February 12, 2015, causing more than $3 million in damage (see paper by Polk et al., p. 477-482). Photos provided courtesy of Jason Polk and Western Kentucky University.
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FOREWORD

Welcome to the Fourteenth Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst in lovely Rochester, Minnesota! This venue is the farthest north that the Sinkhole Conference—as it is better known—has met since its inception in 1984. The setting will provide conference participants with a unique opportunity to view karst phenomena within the glaciated and driftless terrains of southeastern Minnesota, and a significant number of papers presented in this volume address the hydrological and geological characteristics of karst within the upper Mississippi Valley region.

The National Cave and Karst Research Institute (NCKRI) is a non-profit organization dedicated to pure and applied research on caves, karst phenomena, and karst hydrology. In 2011, NCKRI assumed a leadership role in organizing and hosting the Sinkhole Conference, and the 2013 Sinkhole Conference that was held at NCKRI headquarters in Carlsbad, New Mexico was an unqualified success.

This year, NCKRI is pleased to partner with the Minnesota Ground Water Association (MGWA) for hosting the Sinkhole Conference. The MGWA is a non-profit, volunteer organization dedicated to the following primary objectives: 1) promotion and encouragement of the scientific and public policy aspects of ground water; 2) establishing a common forum for scientists, engineers, planners, educators, attorneys, and other persons concerned with ground water; 3) education of the general public regarding ground water resources; 4) dissemination of information on ground water through meetings of the membership. I can think of no better way to meet these objectives than through the excellent research and information presented at the conference and published within these Proceedings.

I am exceptionally pleased that the papers and abstracts within this volume aptly represent the current state of the science, as well as cover notable recent occurrences of sinkholes and other karst phenomena. At the time of this writing (August 19, 2015), news reports are circulating about yet another sinkhole in Florida, but not just any sinkhole: this particular sinkhole opened up in the town of Seffner, in the very same place where Jeffrey Bush tragically lost his life on the 28th of February, 2013 when the ground beneath the room where he lay caved in and swallowed him. That event captured the attention of the nation and was reported across the world, quickly elevating karst geohazards within the public eye. Once again, all sinkholes had become newsworthy, no matter how small, or whether they occurred in actual karst areas.

A number of truly spectacular sinkhole collapses followed the 2013 tragedy in Seffner, Florida, including the event at the National Corvette Museum in Bowling Green, Kentucky on the 12th of February, 2014 that destroyed a number of vintage automobiles (see the paper by Polk et al., p. 479-484), as well as the continued saga of salt dome collapse threatening an entire community within Bayou Corne, Louisiana (see the paper by Jones and Blom, p.415-422).

Sinkholes have also generated news outside of the United States in recent years. For example, periods of unusually heavy rainfall in the winter of 2013-2014 triggered numerous sinkholes in the United Kingdom, sparking a similar media frenzy of sinkhole coverage (see the paper by Banks et al., p. 223-230). Such events, and the public interest they have generated, demonstrate the current relevance of the study of karst and karst geohazards. I am confident that the information contained within these Proceedings will serve as a reference for many future studies.

Daniel Doctor
U.S. Geological Survey
Reston, Virginia
INVITED SPEAKER

HALES BAR AND THE PITFALLS OF CONSTRUCTING DAMS ON KARST

J. David Rogers
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Hales Bar Dam was built on the Tennessee River 33 miles downstream of Chattanooga by a private company to generate power in 1905-1913. The dam site was selected because it was the narrowest reach in the downstream end of the Walden Ridge Gorge. The site is underlain by Mississippian Bangor Limestone on the southeast flank of the Sequatchie Anticline.

Three different contracts failed to complete the dam because of difficult foundation conditions. From 1910-1913 diamond drill core holes were used to explore the site and a series of reinforced concrete caissons 40x45 ft on upstream side and 30x32 ft on the downstream side were installed. Excessive leakage soon appeared near the eastern abutment, and gradually increased. Soundings were made in 1914 to ascertain the areas of gross leakage. Shortly thereafter, rags were placed over suction holes on the river bed and concrete pumped over these. Once a leak was stemmed, leakage would resume at other, adjacent locations. The owners tried to stem the leaks by inserting hay bales, old mattresses, chicken wire, and even corsets! In 1919 the owners began drilling grout holes from the inspection gallery within the dam and pumping hot asphalt into the voids. This was followed by the injection of 78,324 cubic feet of hot asphalt grout into the dam foundation, using 6,266 lineal feet of boreholes with average hole depth of 92 ft. By 1922 the problem appeared solved, but leakage gradually resumed between 1922-1929, rising to the same level as had been observed previously.

In 1930-1931 a new program of exploration was undertaken, using dyes and oils to identify conduits under the dam. Leakage was found to vary between 100 and 1200 cubic feet per second (cfs). When the dam was acquired by the Tennessee Valley Authority (TVA) in 1939 they employed fluorescein dyes to track the underseepage. Dye tests revealed that the leakage varied between 1720 and 1650 cfs; about 10% of the river’s normal flow. They also noted seepage boils forming in the gravel bars, which increased each year. The TVA began constructing the most expensive cutoff wall ever built, drilling 750 18-inch diameter holes along the dam’s centerline and backfilling this with concrete to a maximum depth of 163 feet, extending 25 to 103 feet below the river bed. In April 1963 the TVA announced it was abandoning Hales Bar Dam, due to increasing leakage.

Biography
Dr. J. David Rogers holds the Karl F. Hasselmann Chair in Geological Engineering at the Missouri University of Science & Technology in Rolla, Missouri. He is presently representing the geological and geotechnical engineering professions on the National Academies panel that has been charged with examining “Levees and the National Flood Insurance Program: Improving policies and practices,” being funded by FEMA. Dr. Rogers has served as principal investigator for research funded by the NSF, U.S. Geological Survey, National Geospatial Intelligence Agency, Federal Highway Administration, Department of Defense, and several state departments of transportation. He has served on numerous panels, including the Mississippi Delta Science & Engineering Special Team, the Coastal Louisiana Recovery Panel, the NSF Independent Levee Investigation Team and USGS Investigation Teams evaluating the impacts of Hurricanes Katrina and Rita, the NSF team evaluating the 2008 and 2011 Mississippi River floods, and the Resilient and Sustainable Infrastructure Networks team funded by NSF to make a five year examination of the California Bay Delta flood protection systems.

Dr. Rogers received his B.S. degree in geology from California Polytechnic University at Pomona, his M.S. degree in civil engineering from the University of California, Berkeley, and his Ph.D. in geological and geotechnical engineering at the University of California, Berkeley. He served on the Berkeley faculty in civil engineering for seven years prior to accepting his current position in 2001.
Abstract

Trout Brook in the Miesville Ravine County Park of Dakota County Minnesota is the trout stream with the highest nitrate concentration in the karst region of southeastern Minnesota. Water quality data from 1985 and 1995 (Spong, 1995) and from 2001, 2002, 2006, 2010, and 2014, collected by the Dakota County Soil and Water Conservation District (Dakota SWCD, 2014) document an increasing level of nitrate in Trout Brook. A karst hydrogeologic investigation was designed to measure nitrate levels at sampling points along the stream and to increase our understanding of the source and movement of nitrates throughout the length of Trout Brook. Eighteen springs and seeps have been located in the Main Branch and tributaries of Trout Brook. A previously unreported flowing section and stream sieve, Weber Sieve, were found above what had been thought to be the head of perennial flow in the East Branch of Trout Brook. Two new sinkholes developed after the 14-15 June 2012 flood in a field northeast of the East Branch of Trout Brook. This investigation included regular monitoring of major anions in the streams and springs, synoptic stream flow measurements, and a dye trace of a sinking stream in the Trout Brook drainage.

The initial assumption was that the majority of the baseflow of Trout Brook was from discrete springs. However, synoptic baseflow and nitrate measurements show that only 30-40 percent of the total flow in Trout Brook is from discrete springs, and the rest appears to be from distributed groundwater discharge directly into the stream. Both the discrete springs and the distributed recharge occur along reaches of Trout Brook that drain the significant high transmissivity zone near the bottom of the regionally important Shakopee aquifer. Dye traces have confirmed flow-paths from Weber Sieve to LeDuc and Bridgestone Springs and have begun to define springsheds for these head water springs. Nitrate concentrations and chloride/bromide ratios decreased systematically from the upstream springs to the downstream springs.

The nitrate concentrations have been increasing at four springs from 1985 to 2014 and at two surface sampling points from 2001 to 2014. The nitrate concentration of another surface sampling point increased from 2001 to 2006, decreased from 2006 to 2012, and increased from 2012 to 2014. Snowmelt and rainfall runoff was sampled on 2 March 2012 and showed no detectable nitrate in the runoff from a watershed with no row-crop agriculture, but elevated nitrate was detected in an adjacent watershed with row-crop agriculture. All of these trends illustrate the dominance of agricultural sources of nitrate in Trout Brook.

Introduction

Karst

Complex surface and groundwater interactions are dominated by karst processes in southeastern Minnesota. Karst features often include caves, sinkholes, springs, stream sieves, and sinking streams. A springshed is the subsurface and surface areas that provide the discharge to a spring. A stream sieve describes a losing reach of a surface stream where specific water sinking points, stream sinks, are not evident. Karst features result from water containing carbonic acid which dissolves the carbonate in soluble bedrock. Water quality is a concern because karst features allow rapid groundwater velocities and short residence times. Karst springs provide the source water to premier trout streams in southeastern Minnesota. Trout Brook in the Miesville Ravine Park Reserve in southeast Dakota County and south of Miesville, Minnesota is one of these trout streams.

Location, Geology, and Topography

Trout Brook (Figure2) is located in the southeastern part of Douglas Township (T113N, R17W) of Dakota County, south of Miesville, Minnesota in the Miesville...
The Shakopee Formation unconformably overlies the Oneota Dolomite.

The Shakopee and Oneota Dolomite collectively form the Prairie du Chien aquifer, which is one of the most heavily used aquifers in Dakota County. An unconformity between these two formations represents a 10 million year subaerial erosion episode that left a high transmissivity zone (HTZ) of significantly enhanced porosity and permeability in the top of the Oneota. During and after the deposition of the Shakopee, karst solution processes expanded the HTZ up into the lower Shakopee. Runkel et al., (2003) and Tipping et al., (2006) report that this mid-Prairie du Chien HTZ is one of the primary features of the hydrostratigraphy of southeastern Minnesota. The source water for Trout Brook drains directly from this high transmissivity zone along bedding plane fractures, and through solutionally enlarged porosity and permeability and anastomosing karst conduits.

The Shakopee Formation unconformably overlies the Oneota Dolomite.

The Trout Brook surface watershed is largely an intensively cultivated gently rolling upland. The main land use type in the Trout Brook watershed is row-crop agriculture, and many farmers irrigate their crops due to the sandy soil. Livestock feedlots in the watershed are also potentially significant sources of nitrate. Very little surface water flows on the upland except during spring snowmelt and after the largest and most intense precipitation events. Most routine precipitation not consumed by evapotranspiration rapidly infiltrates to groundwater. South of Meisville the surface drainage abruptly incises steep-sided valleys to form the West and East Branches of Trout Brook. The East and West Branches join to form the Main Branch of Trout Brook in the Miesville Ravine Reserve Dakota County Park. The lower reaches of both branches of Trout Brook and the Main Branch widen downstream. All branches of Trout Brook meander across steep-sided, flat-bottomed valleys.

**Historical**

Spong studied Trout Brook, Dakota County, in 1985 and 1995. He analyzed water chemistry at four springs (Beaver, LeDuc, Fox, and Swede Springs) in 1985 and two springs (LeDuc and Swede Springs) in 1995. The two sampling events in 1985 and 1995 were collected during baseflow periods (with no significant stormflow) in years with normal precipitation (Ron C. Spong, written communication, 2012). Flow rates were measured at Trout Brook’s springs and streams in 1985
(Spong, 1995). These data are important because they document water quality of the springs 27 and 17 years ago; data points that are critical in defining water quality time trends.

The Dakota County SWCD measured baseflow and obtained grab samples during storm events at Trout Brook during 2001, 2002, 2006, and 2010. Flow measurements were taken to characterize low flow and stormflow. The water samples were analyzed for typical water quality parameters and were reported to the State of Minnesota. Automated stage monitoring was in place, but the data is suspect due to the flashy nature of this stream (Dakota Co. SWCD, 2010).

**Row-Crop Agriculture Versus Nitrate**

Reactive nitrogen is an environmental concern. In the hydrosphere it can lead to eutrophication, toxic algae blooms, and hypoxia. In the atmosphere it contributes to acid rain, deposition of nitrogen in forests leading to nitrogen saturation which can alter the soil, and global warming. Reactive nitrogen can also be harmful to humans due to air pollution and contamination of drinking water. The U.S. Environmental Protection Agency (EPA) reports that greater than 10 parts per million (ppm) of nitrate-nitrogen in drinking water can have adverse health effects (U.S. EPA, 1990). (In the rest of this work, the word “nitrate” is synonymous with “nitrate-nitrogen”.)

The Hastings Area Nitrate Study (HANS, 2003) was carried out in Dakota County, MN near the study area. That study highlights the nitrate contamination problem in groundwater and considered three possible sources: row-crop agriculture, feedlots, and septic systems. Groundwater samples collected in 2000 showed nitrate levels varied among the aquifers. The Shakopee aquifer had the highest concentration of nitrate at 15 ppm, the Quaternary aquifer was next at 8.7 ppm, and the Jordan aquifer had the lowest at 1.85 ppm.

Figure 1 from Watkins (2011) shows the percent of row-crop agriculture plotted against nitrate plus nitrite in streams at baseflow in the karst region of southeastern Minnesota. Groundwater discharge supports the baseflow of streams and rivers in this karst region, with discrete springs typically providing a substantial portion. A linear relationship is present with $R^2$ value of 0.70. This indicates a strong relationship between the percentages of row-crop agriculture in southeast Minnesota watersheds versus the nitrate concentrations in streams at baseflow. Trout Brook was selected for study because it has the highest nitrate concentration at baseflow of monitored streams in southeastern Minnesota. Trout Brook is the uppermost data point, below 18 mg/L, on the above graph.

**Monitoring**

An initial survey of Trout Brook occurred in February 2011, and a second round of sampling occurred in July 2011. A systematic water sampling campaign began in October 2011 and ended in October 2012.

Eighteen discrete springs documented along Trout Brook are shown in Figure 2. All of the springs emerge where the steep valley walls meet the flat valley floor. All but two of the springs, Beaver and Swede, emerge where Trout Brook has meandered up against the base of the valley walls. Beaver and Swede are buffered from Trout Brook by beaver dam induced wetlands.

**Geochemistry**

The field and analytical methods used in this work are described in Alexander and Alexander (2011). The nitrate concentrations at Fox, LeDuc, Beaver, and Swede Spring have been increasing with time (Figure 3). The nitrate concentration of Fox Spring is increasing at the greatest rate of 0.42 ppm/year while the concentrations at Beaver, Swede, and LeDuc Springs are increasing at rates of 0.25, 0.18, and 0.11 ppm/year, respectively. These rates were calculated over a 27 year span. The rate of increase for Fox Spring is almost twice as high as the next highest rate, that of Beaver Spring. These increases are likely due to changes in farming practices over time and the intensity of farming on the contributing springsheds.

Figure 2 shows the concentrations of these four springs compared to the other springs in Trout Brook as color-coded dots. The springs discharging into the West Branch of Upper Trout Brook have the highest nitrate levels. The springs discharging into the East Branch of Upper Trout Brook have low to moderate nitrate levels. The three yellow dots on the Main Branch indicate moderate nitrate levels and the springs further downstream have the lowest nitrate levels.

The nitrate concentrations in spring resurgence appear to be controlled by location. This relationship is likely
The chloride/bromide ratios of Trout Brook’s springs and streams on Figure 5 are averaged values from samples collected from 2011 to 2012. The figure also shows the discrete chloride/bromide ratios of the samples collected in the Main Branch of Trout Brook during the 28 October 2011 synoptic stream flow measurement campaign. The West Branch Springs have the highest ratios indicating the greatest anthropogenic impact. The East Branch Springs have the second highest chloride/bromide ratios. The Main Branch Springs have the lowest ratios. 

Figure 5 shows that the chloride/bromide ratios are a function of distance downstream. The chloride/bromide ratios at springs seem to be decreasing towards the southeast. This reduction is occurring from the headwater springs of the West Branch towards Swede Spring. The 28 October 2011 chloride/bromide ratios show that there was not a significant decrease from the upstream to the downstream end of the Main Branch. The total of all the averaged ratios of the springs are shown as the purple “X.” This total averaged value is very similar to the averaged ratios of all samples from TB3 (denoted by the dark green circle), as well as the discrete chloride/bromide ratio of the sample from TB3 on 28 October 2011. This is evidence that the majority of flow from distributed discharge into the stream channel has similar chloride/bromide ratios as the mixed water in the stream channel. The lower chloride/bromide ratios in the Main Branch springs do not significantly lower the chloride/bromide ratios of the mixed water.

determined by the springshed of each spring. The springs further downstream probably involve longer flowpaths draining deeper parts of the aquifers. The contributing springsheds probably vary substantially by different types and percentages of land use.

Figure 4 shows the nitrate concentrations at baseflow in the East, West, and Main Branches of Trout Brook between 2001-2012. Nitrate levels in the West Branch [as represented by the results of sample TB2] increased at the greatest rate, 0.35 ppm/year. From 2001-2006 nitrate in the East Branch [represented by the sample TB1] increased at a rate of 0.11 ppm/year; however, nitrate concentrations decreased from 2006-2012. The scope of this study did not include further analysis of this phenomenon. The Main Branch [as represented by sample TB3] was collected the furthest downstream and had a nitrate increase of 0.09 ppm/year. The TB3 sample near the end of the Main Branch is a collection of all the water mixing with surface water and groundwater in the streamshed, derived from baseflow and runoff events. At baseflow that water is almost entirely a variable mixture of discrete spring flow and distributed groundwater discharge, as there is no significant surface runoff to Trout Brook during baseflow.

Chloride/bromide ratios are useful in groundwater studies because chloride and bromide are conservative anions that travel with the groundwater and can be used to identify the source water that is recharging the aquifer. Chloride/bromide ratios provide indications of anthropogenic impacts on waters and are explained further by Anger and Alexander (2010).
A 28-29 October, 2011 synoptic study, of the flow and nitrate levels of Trout Brook at baseflow conditions was conducted. The data were used to understand the interaction between flow and nitrate and to determine if the majority of the flow came from discrete springs or from distributed discharge into the stream channels.

Flow measurements were taken at 32 locations in the Main Branch, tributaries, and springs of Trout Brook to estimate the water contribution from the springs to the overall flow of this trout stream. Measurements were taken upstream and downstream of where the identified springs discharge into East, West, and Main Branches of Trout Brook.

The approximate spring flow was calculated by subtracting the flow of the stream measured upstream of the spring, from the flow measured downstream of the spring. If a spring had formed a channel with enough water, then direct measurements in the spring run were also taken to calculate the flow. All direct flow measurements of the springs were averaged with the flow measurements calculated from the upstream/downstream subtraction. Unfortunately, the flow of some of the springs (with no measurable separate channel) were within the uncertainty of the up and downstream stream flow measurements. In those remaining cases, limits on flow in those spring was visually estimated.

Flow was also measured at particular sites in the stream where a spring was not in proximity. The objective was to distribute the measurements, in order to interpret reaches that may be gaining or losing flow.

Every location where a flow measurement was taken, a water sample was also obtained. Water samples were also retrieved at spring orifices. This was done to understand the mixing concentrations of nitrate and chloride/bromide ratios.

The initial hypothesis that the source of baseflow in the Main Branch of Trout Brook and its tributaries would be primarily from discrete springs is falsified. Data from the two-day synoptic flow measurements show that the majority of the flow is not from discrete springs. Approximately 30-40 percent of the total flow at the sampling point TB3, close to the outlet of Trout Brook, is from spring water. The remaining flow is apparently

Swede Spring has an average chloride/bromide ratio of 348. This is the lowest ratio detected in the springs and is closest to the ratio of rainwater, which varies from 200-250 in Minnesota. The West Branch Springs have the highest ratios, ranging from 630-680. These ratios are comparable to those found in manure and fertilizer (Alexander, 2005).

Surface water runoff was sampled on 2 March 2012 from the recession of 1.67 inches of rain that occurred on 29 February 2012. Two samples were obtained from two different ravines. Each ravine has its own small watershed, which were delineated using the watershed tool in ArcGIS: ArcMap 10.1. Land use polygons were created using aerial photographs.

In 2011-2012, the smaller watershed (darker green on Figure 2, next to fluorescent green) did not contain any row-crop agriculture while the larger watershed (fluorescent green on Figure 2) contained 40 percent row-crop agriculture. The nitrate concentration was below the detection limit in the samples collected from the watershed with no row-crop agriculture. The samples from the watershed with row-crop agriculture had a nitrate concentration of 10.6 ppm.

The 10.6 ppm nitrate in its runoff event from the watershed with row-crop agriculture is consistent with the results in Figure 1. The rapid groundwater velocities and short residence times in karst does not allow enough time for significant nitrate reduction. The nitrate concentrations at springs may be a key indicator of the percent row crop agriculture of their springsheds due to the rapid, direct water flow in karst.

**Synoptic Flow and Nitrate Assessment**

**Runoff Versus Land Use**

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**Figure 5.** Chloride/bromide ratios are a function of distance from the confluence of East and West Branches of Upper Trout Brook.

**Synoptic Flow and Nitrate Assessment**

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Flow measurements were also taken on 31 March 2012. This was done to see how much flow was gained from the confluence of the East and West Branch to the stream section between Beaver and Hill Springs. The majority of this area was not measured on the 28-29 October 2011 study. It was found that approximately 4.5 cfs of water was gained along this portion of the Main Branch. The only discrete spring in this area is Beaver Spring, which was discharging approximately 0.7 cfs of water into the Main Branch. Approximately 3.8 cfs of water was from distributed discharge into the stream channel.

The majority of flow of Trout Brook is from distributed discharge into the stream channel. However, the flow inputs have changed in Trout Brook. In 1985 the East Branch was contributing more flow to Trout Brook than the West Branch (Spong, 1995) and in 2011 the West Branch was contributing more flow than the East Branch. This illustrates the changes that occur with flow regimes over time. These changes could be from climate, anthropogenic activities such as irrigation, changes in land use, or from changes in the steam channel itself by major floods.

**Weber Run Dye Traces**

Weber Run is located in the E½ of the SE¼ of sec 22 of Douglas Township on private land on the northwest corner of the County Park. Weber Run is shown on Figure 2.

Surface flow in Weber Run is fed by several source springs and then, under all but high flow conditions, sinks completely in a stream sieve (indicated by the

Figure 6 displays nitrate concentration and measured flow versus distance from the confluence of the East and West Branches of Trout Brook. Flow is displayed on the left vertical axis. Nitrate concentration is on the right vertical axis. The horizontal axis displays distance (in meters) along the streams from the confluence of the East and West Branches to form the Main Branch of Trout Brook.

The nitrate concentration decreases only slightly downstream, from ~ 16 ppm below the confluence to ~ 13 ppm at TB3. In contrast, flow in the Main Branch increases by about a factor of 3 over the same reach, from about 4 cfs to over 12 cfs. This relationship can be seen in Figure 6. Distributed groundwater inflow dominates the Main Branch of Trout Brook and the nitrate content of that water is apparently in the 13 to 15 ppm range.

**Figure 6. Flow and nitrate concentrations are a function of distance from the confluence of the East and West Branches of Trout Brook.**
blue-dashed line). The source springs (also shown on Figure 2) are a series of small springs and seeps on the southwest stream bank.

On 28 December 2011 Rhodamine WT was introduced in Weber Run. A second dye trace was introduced in Weber Run on 12 April 2012 using eosine dye.

Four positive detections of Rhodamine WT and three positive detections of eosine at LeDuc Spring and one detection of Rhodamine WT confirm that at least a portion of the water of LeDuc and Bridgestone Springs comes from the Weber Sieve area of Weber Run. The 29-day travel time between Weber Sieve and LeDuc Spring measured in the first dye trace corresponds to a groundwater flow velocity of ~27 meters/day in the south-southeast direction. The first dye trace also estimated groundwater flow from Weber Sieve to Bridgestone Springs to be 15-20 meters/day. The 79-day travel time between Weber Sieve and LeDuc Spring corresponds to a ~10 meters/day groundwater velocity during the second dye trace. As the 28 December 2011 and 12 April 2012 traces took place under low flow conditions in an exceptionally dry winter, spring, summer, and fall, the flow velocities under normal, higher flow conditions may be faster.

**Conclusion**

Trout Brook is a trout stream in Dakota County’s Miesville Ravine Park Reserve. An MPCA survey found Trout Brook had the highest baseflow nitrate concentrations measured in southeastern Minnesota’s karst region. This project investigated the karst hydrogeology and water quality in Trout Brook’s water to gain information on the source and movement of nitrates through the landscape. This investigation located springs, stream sinks, sinkholes, and other karst phenomena in the Trout Brook watershed. Synoptic surveys of the stream and spring flows. Periodic water samples were collected and analyzed to document nitrate and chloride/bromide ratio time trends. Two dye traces were conducted initiating springshed mapping for the springs. This study combined existing, historical data from 1985 and 1995 with our 2011-2012 results to quantify nitrate time trends for four springs. Data from the 2001, 2002, 2006, and 2010 Dakota SWCD surveys, combined with our 2011-2012 results, permitted documentation of shorter-term time trends for three points in the surface streams. The study period 2011-2012 was abnormally dry but significant floods occurred on 6 May 2012 and 14-15 June 2012.

Our stream and spring flow measurements indicate that 30-40 percent of the water in Trout Brook is from identified discrete springs. As surface runoff contributes to Trout Brook steam flow only during and immediately after significant precipitation events, distributed groundwater discharge into the stream channel, therefore makes up 60-70 percent of the baseflow.

Nitrate concentrations ranged from about 9 ppm in springs near the downstream end of Trout Brook to over 25 ppm at the upstream springs in the West Branch of Trout Brook. Chloride/bromide ratios decreased systematically from the upstream springs to the downstream springs. The nitrate concentrations in four of the springs increased at rates ranging from 0.42 to 0.11 ppm/year from 1985 to 2012. The nitrate concentrations at one upstream site, and the downstream surface water sampling points, have increased at similar rates from 2001 to 2012. The nitrate concentration at another upstream surface water sampling point increased from 2001 to 2006 but decreased from 2006 to 2012.

Runoff from a 29 February 2012 winter rain event was sampled on 2 March in two small sub-watersheds along the East Branch of Trout Brook. Runoff from a sub-watershed containing only forest and CRP land contained no detectable nitrate. Runoff from a sub-watershed that was 40 percent row-crop land contained 10.6 ppm nitrate.

Dye tracing documented karst aquifer flow-paths from the Weber Run stream sieve to LeDuc and Bridgestone Springs with flow velocities in the range of 15 to 27 meters per day.

The results of this study indicate that row-crop agriculture in the surface and subsurface drainage basins of Trout Brook is the primary cause of the water’s elevated concentrations of nitrate. This conclusion is supported by the MPCA’s correlation between the percentages of row-crop agriculture and the nitrate concentrations in run-off and stream samples.
References


Miesville Ravine Park Reserve – Dakota County.


HUMAN IMPACTS ON WATER QUALITY IN COLDWATER SPRING, MINNEAPOLIS, MINNESOTA

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Abstract
Coldwater Spring in Minneapolis, Minnesota was the water supply for Fort Snelling from the 1840s to 1920. The spring site has been declared a sacred site by some federally recognized Native American tribes. The site is managed by the National Park Service. This project has monitored the water chemistry of Coldwater Spring to document human impacts on the spring’s water quality. Temperature, dissolved oxygen, conductivity, pH and anions were monitored weekly and cations and alkalinity monitored monthly at Coldwater Spring and the adjacent Wetland A from 15 February 2013 through 18 January 2015. Coldwater Spring’s water flows through fractures in Platteville Limestone of Ordovician age. The basic chemistry of Coldwater Spring should be the calcium magnesium bicarbonate water typical of carbonate springs. However, on an equivalent basis, Coldwater Spring’s water currently contains almost as much sodium as calcium + magnesium and more chloride than bicarbonate.

The chloride concentrations are about 100 times the levels from 1880. Maguire (1880) reported the chloride levels of Coldwater Spring were about 4.5 ppm. During the current study the chloride content in the spring increased from about 320 ppm from March 2013 to about 410 ppm in December 2014. In April, May, and June of 2013 and 2014, the chloride rose about 100 ppm in three-month-long pulses. The chloride concentration of the water in Wetland A ranges from about 400 ppm to over 600 ppm with a pattern that is a mirror image of the Coldwater Spring pattern. This major anthropogenic chloride component has a chloride to bromide ratio of 2,500 ± 300, well within the range of chloride to bromide ratios of road salt, 1,000 to 10,000. Road salt is applied to two major multi-lane highways close to the spring and is used extensively in this heavily urbanized area throughout the winter.

The temperature of the spring is variable and higher than its pre-settlement temperature. Nicollet (1841) recorded the temperature of Coldwater Spring multiple times in summer of 1836 as 46°F (7.8°C) and multiple times in winter of 1837 as 45.5°F (7.5°C). More recently the temperature of Coldwater Spring fluctuates smoothly between 10.7 and 13.1°C. The higher temperature of the springs’ discharge also indicates an anthropogenic source of heat within the spring-shed or spring recharge area. The spring water is coldest in May and June and warmest in October and November. The temperature of the water in Wetland A fluctuates from 6.4 to 13.8°C – in a pattern that is opposite of that in Coldwater Spring.

Coldwater Spring also contained a significant, increasing nitrate-nitrogen component which ranged from 2.5 to 5.2 ppm – with dips at the same times as the chloride pulses. Wetland A’s nitrate-nitrogen level varied between 0.2 to almost 6 ppm with large pulses at the same time as Coldwater Spring’s dips. A 2014 study performed by the US Geological Survey came to the conclusion that increasing chloride levels in lakes and streams are likely driven by increasing road salt application, rising baseline concentrations, as well as an increase in snowfall in the Midwestern area of the US during the time of the study (Corsi 2014). The significant chloride, temperature and nitrate levels are likely to be driven by anthropogenic sources.
**Introduction**

This project monitored the water quality in Coldwater Spring, Minneapolis, Minnesota, to evaluate the impact of anthropogenic pollutants on the spring and an adjacent wetland (“Wetland A”). This report expands and adds a second year of data to Kasahara’s (2014) report on the first year’s results of this project.

In the Twin Cities Metropolitan area (TCMA), roughly 350,000 tons of de-icing road salt is applied to the TCMA roads every year (Sander et al., 2007). The majority of this salt dissolves in snowmelt that either flows overland to surface water bodies or infiltrates to recharge the water table. According to a recent study about 70% of the road salt applied in the Twin Cities area stays in the regions’ watershed (Rastogi, 2010). The road salt used is about 60% chloride and 40% a positive ion, usually sodium, and is generically referred to as NaCl (Keseley, 2007). This road salt runoff can lead to high levels of salinity in freshwater areas due to the dissolution of NaCl in the water. This is harmful to freshwater aquatic life, regional mammals and birds, and plants native to Minnesota.

**Salinity Effects on Wildlife**

Ten percent of freshwater species can die off after just 30 days at salinity concentrations of 220-240 ppm, trout behavior is affected at levels as low as 250 ppm, and the overall diversity of aquatic species decreases as the salinity concentration rises (Keseley, 2007). According to the Minnesota Pollution Control Agency’s (MPCA) 2010 draft report, 11 metro-area streams have levels of chloride concentration above 230 ppm (Homstad, 2010). Road salt particles may attract moose and deer to roadsides, where cars or trucks may strike them, and birds, such as sparrows, can die after eating only two salt particles. Plants as far as 200 feet away from the roadside can still be affected by the rise in salinity, and just a 30 ppm concentration can lead to damage to coniferous species such as the pine tree (Keseley, 2007). Because groundwater is the source for drinking wells, high levels of salinity can also affect humans on restricted-sodium diets (Rastogi, 2010).

**The Sample Sites**

Coldwater Spring is located at 44° 53’ 57.61” N; 93° 11’ 47.81” W. Coldwater Spring is an important Native American cultural site (NPS, 2012) and is an integral part of the history of Fort Snelling and Minneapolis. The spring site is now a part of the National Park Service’s Mississippi River National River and Recreation Area (NPS, 2012) and is open to the public.

The Minnesota Department of Health test results of the spring water in 2005 revealed the presence of bacteriological contamination of total coliform indicating organisms, but did not detect E. coli. The NPS has posted signage at the spring indicating the water is not suitable for drinking.

Figure 1 is an aerial view of Coldwater Spring area. Coldwater Spring is in a highly urbanized area in the southeast corner of Minneapolis, Minnesota. A major trunk highway/light rail interchange is immediately west of the spring. The large white squares visible northwest of the spring are buildings of a VA Medical center. Runway 22-4 of the MSP International Airport extends into the southwest corner of Figure 1. St. Paul, Minnesota is across the Mississippi River to the east-northeast.

The groundwater in the Coldwater Spring region typically flows from west to east toward the Mississippi River, and recharges from precipitation and lakes such as Lake Nokomis and Mother Lake. The bedrock geology of this area is of ± 26 feet of Platteville Limestone and a variety of soils, which overlays ± 3 feet of Glenwood Shale (Howe 2014). Both these layers overlie a highly permeable layer of St. Peter Sandstone. The Platteville Limestone is often described as being two different layers, Upper and Lower Platteville, because of the differing amounts of shale content. In the Lower Platteville Limestone, the shale comprises greater than 30% of the Platteville’s composition, while the Upper Platteville has a much lower shale contents. The Lower Platteville Limestone thus has a lower conductivity and acts as a

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**Figure 1.** An aerial view of the Coldwater Spring site.
confining layer for the Upper Platteville Limestone. Water in the Upper Platteville Limestone has easy horizontal flow paths (Howe 2014). Coldwater Spring’s water flows through fractures in the Upper layers of Platteville Limestone of Ordovician age. Studies of this formation have shown that it is ineffective at filtering out many of the contaminants from the recharge areas, in this case runoff from residential housing and major highways (NPS, 2012).

The Wetland A sample site is a reconstructed wetland. The Wetland A previously had a building from the former Bureau of Mines, Twin Cities Research Center (TCRC) situated adjacent to it. The TCRC was closed in 1996 and the building removed in 2012. The adjacent historic springhouse and reservoir were left intact as the restoration of the Coldwater Spring site was completed by the National Park (NPS, 2012). Wetland A is located at 44° 53’ 55.14” N; 93° 11’46.91” W.

The “Big Culvert” is a 6-foot diameter storm water outlet that drains the area southwest of the Coldwater Spring site. It flows perennially and was sampled twice late in this project. The Big Culvert is located at 44° 53’ 47.91” N; 93° 11’44.81” W.

**Methodology**

Weekly visits were conducted to Coldwater Spring and Wetland A to record the temperature, to measure the conductivity, dissolved oxygen and pH levels of the spring using a Thermo Orion multi-meter, and to collect weekly anion samples. The Coldwater Spring samples were collected at the outfall pipe on the southwest corner of the springhouse structure. The anion samples were analyzed using ion chromatography [EPA 300.0]. Samples were collected monthly for alkalinity analyses by digital titration [ASTM 1067 02] and cation analyses by ICP/OES [EPA 200.7]. Water samples were collected from the two sites from the beginning of February 2013 to the middle of January 2015. The timing of this data collection was intended to study any impact to the springs’ salinity levels during snow melt, when deicing salt would presumably be washing into and through groundwater supplies. A temperature data logger was installed in Coldwater Spring. The data logger was inserted into a feed pipe along the north side of the springhouse about 3 meters from the outfall pipe.

**Data and Interpretation**

Although Coldwater Spring and Wetland A are less than 100 meters apart, their water chemistries and flow rates show differences. Both show significant seasonal fluctuations in many parameters which, in several cases, are mirror images of each other.

**Chloride**

Figure 2 displays the chloride concentrations in Coldwater Spring, Wetland A and the Big Culvert during the monitoring period.

**Historical Data**

During the 19th Century Coldwater Spring was the water source of the U.S. Army’s Fort Snelling in Minneapolis. Army Captain Maguire (1880) reported the chloride level in Coldwater Spring to be 0.26 grains per gallon which is equivalent to 4.5 ppm in modern units. [Early analyses of chloride were done by titration with a clear end-point and are quite reliable.]

**Coldwater Spring**

Figure 2 shows that the chloride content of Coldwater Spring’s water increased from 320 ppm in March 2013 to 410 ppm in January 2015. Superimposed on the increase of about 45 ppm/year were two pulses between roughly April and July of 2013 and 2014. The first pulse reached 432 ppm on 10 June 2013. The second pulse rose to 455 ppm on 20 May 2014. These chloride levels are about 100 times the levels from 1880.

Deicing salt is applied to two major multi-lane highways close to the springs throughout the late fall, winter and early spring. If the pulses are attributable to infiltration...
of road salt runoff, these data suggest it takes about five months for the winter road salt input to reach Coldwater Spring. The magnitude of each year’s pulse should depend on the severity of the winter – which fluctuates significantly from year to year depending on the severity of each winter. Ignoring the chloride pulses, chloride levels in Coldwater Spring are very high and generally increased over the period of this study.

**Wetland A**

Wetland A chloride levels (Figure 2) were higher than the chloride levels at Coldwater Spring, and displayed an inverse relationship to Coldwater Spring. Wetland A’s chloride levels were low when Coldwater Spring’s chloride pulses were high – and vice versa. The dips in Wetland A chloride concentrations were greater than the pulses in Coldwater Spring’s concentrations, but occurred at essentially the same time. Wetland A’s chloride levels ranged from 313 ppm to 632 ppm. This mirror image behavior of Coldwater Spring and Wetland A’s water quality is repeated in several of the other parameters discussed below. This behavior is cryptic and surprising. We are not aware of any other system that displays this mirror image behavior. Chloride is the most common anion in both Coldwater Spring and Wetland A’s water.

**Big Culvert**

Only one sample was collected from the Big Culvert. Big Culvert’s value is shown in Figure 2 and subsequent graphs. This result is similar to the data seen at Wetland A.

**Chloride to Bromide Mass Ratios**

Salts from different sources have different, characteristic chloride to bromide ratios (Davis et al., 1998). In addition to the sewage and rainfall Cl/Br ratios, Panno et al. (2006) found Cl/Br ratios for softened water ranging between 175 and 1,122 and for agrichemical-affected water with tile drains, from 108 to 1974. The road salt ratio minimum was 1,164 and ranged up to 4,225. Measurements of the chloride to bromide ratios in Coldwater Spring, Wetland A and Big Culvert help to identify the source of the salt. Figure 3 shows the chloride to bromide mass ratios from the sample sites. It also shows the chloride to bromide mass ratios of three road salt samples taken from different sites on the University of Minnesota campus in winter 2014.

**Coldwater Spring, Wetland A and Big Culvert**

The chloride to bromide ratios in all three sample sites scatter from about 2,000 to 4,500 - well within the range of reported values for road salt. The ratios from the three samples of road salt taken at three different sites around the University of Minnesota campus are in the same range. These data suggest that road salt is the primary contributor to the high chloride levels in Coldwater Spring, Wetland A and the Big Culvert.

**Temperature**

**Historical Temperature Data for Coldwater Spring**

In 1836 and 1837, a French explorer named Joseph Nicollet measured the temperature of Coldwater Spring in both the summer and the winter months (Nicollet, 1845). He described Coldwater Spring’s temperature as:

“Of the numerous springs that issue from the foot of the...bluffs [adjointing Fort Snelling] there is one particularly deserving of notice. It is very abundant and perfectly shaded. It goes by the name of Baker’s spring. Having taken its temperature three times a day during twenty days of the month of July, 1836, and then again during the following winter months, I never found it to vary more than 46°F in July, and 45.5°F in January.”

Converting the temperatures from Fahrenheit to degrees centigrade, the measured temperature average of Coldwater Spring was about 7.8°C in July and 7.5°C in the winter. This indicates that in the summer of 1836 and
The winter of 1837, Coldwater Spring’s temperature was essentially constant. The temperature of shallow springs is typically the average annual air temperature of that region. The shallow ambient groundwater temperature in this part of Minnesota is about 8°C.

**Temperature Records**

Figure 4 displays the weekly temperatures of Coldwater Spring and Wetland A. The temperature was measured using an ASTM calibrated glass thermometer.

Figure 5 contrasts the weekly temperature records for Coldwater Spring and Wetland A with the average air temperatures on the days that samples were collected. The air temperature data, measured at the Minneapolis/St. Paul Airport and which is two kilometers from Coldwater Spring, are from the National Weather Service website, as follows: [www.crh.noaa.gov/mpx/Climate/MSPClimate.php](http://www.crh.noaa.gov/mpx/Climate/MSPClimate.php).

In Figure 6 the 15 minute interval data logger temperature record from Coldwater Spring is shown in blue. The gaps in the data logger plot are due to equipment problems. The weekly thermometer data from Coldwater Spring (from Figures 4 and 5) is shown in red.

**Temperature in Coldwater Spring**

The field thermometer temperature of Coldwater Spring (Figure 4) fluctuated seasonally between 10.7°C and 13.1°C in a sinusoidal pattern. This is a 3- to 5-degree increase from the 1836 and 1837 temperature levels. Since the temperature fluctuates over a 2.4 degree range, it is clear that Coldwater Spring is no longer a constant temperature spring. The increasing range of spring temperatures may reflect increased urbanization where snow is cleared from roads and parking areas allowing greater frost depths and much higher temperatures under summer sun.

The lowest point in the temperature fluctuations occurs in early June. The highest temperatures occur in early November. In comparison the air temperatures (Figure 5) peak in late July or early August and the minimum temperatures occur in late January to early February. The temperature fluctuations in Coldwater Spring lag air temperature fluctuations by about four to five months.

The data logger record (Figure 6) has relatively smooth, linear temperature profiles with minimal short term variation. The absence of hours to days temperature excursions indicate that the flow path to this sampling point is long enough to even out variations due to storm events but not long enough to average out seasonal effects.

The data logger temperature record is surprisingly different from the thermometer readings record. The shape is closer to a linear zig zag than a sine wave. The maximum and minimum temperatures of the data logger record are 0.3 to 1°C higher than those in hand measured temperature record using the thermometer.

Discrepancies in the two temperature plots may be due to the location of the temperature samples taken. The data logger was located in one of the entrance pipes on the north side of the springhouse, and the thermometer samples were taken from the south side of the springhouse in the exit pipe, about ten feet from the data logger. Alternatively, the discrepancies may...
The concentration of nitrate-nitrogen rose by about 2 ppm over two years, and also displayed a sinusoidal pattern. The annual nitrate minima occur almost simultaneously with the highest chlorides. Seasonal application of lawn fertilizers occurs roughly opposite the use of road salt.

Nitrate-Nitrogen Levels in Wetland A
Wetland A had a wider nitrate-nitrogen level range than Coldwater Spring, varying between about 0.3 ppm and 6.0 ppm. Wetland A’s nitrate-nitrogen levels vary much more than do Coldwater Spring’s but are rising when Coldwater’s levels are dropping and vice versa.

Major Ion Chemistry
Piper Diagrams are a 2D graphical technique for displaying and interpreting the major ion chemistries of water samples (Piper, 1944). The lower left triangle is a ternary diagram of the major dissolved cations. The lower right triangle is a ternary diagram of the major anions. The two triangles are projected into the middle square. Figure 8 is a 3D Piper Diagram where the total concentrations of the dissolved ions are shown by the vertical bars. Figure 8 illustrates that the waters from Coldwater Spring and the waters from Wetland A have roughly similar major ion chemistries, but differ enough to be able to be distinguished. Wetland A’s waters contain higher levels anthropogenic ions of sodium, chloride, and nitrate, than Coldwater Spring’s waters.

Most springs from carbonate rocks in Minnesota fall closer to the bicarbonate corner of the anion triangle than do the waters Coldwater Spring and Wetland A. Solution of the Platteville Limestone produces calcium, magnesium and bicarbonate ions. Coldwater Spring’s anions are about 50% chloride and Wetland A’s anions provide evidence for multiple, different water inputs to the Coldwater Springhouse. Such different sources (fed by different flow paths) would be the start of an explanation for the observed chemical and temperature variations.

Temperature in Wetland A
Wetland A thermometer temperatures ranged between 6.4 and 13.8°C, with the temperature peaking at the end of August. Wetland A is also not feed by a constant temperature spring. The Wetland A’s water source temperature fluctuation is out of phase with the air temperature by about two months. Wetland A temperatures range from well above to below the average groundwater temperatures of roughly 8°C. Wetland A’s temperatures can be explained as being driven by the air temperatures in a very shallow flow system.

Nitrate-Nitrogen
The 1974 Safe Drinking Water Act set a national maximum contaminant level (MCL) for nitrate of 10 ppm as an enforceable standard (Water, 2013). Although the nitrate levels in each sample site are below this drinking standard, maximum contaminant levels are purposely set as close as possible to the health goals. Nitrate-nitrogen is an indicator parameter of contamination, and is often accompanied with other pollutants such as pathogenic viruses, bacteria and synthetic organic compounds. Therefore, the elevated levels found in Coldwater Spring and Wetland A are significant and, in this urban setting, suggest contamination from human sources such as lawn fertilizer runoff or leaking sanitary sewage lines.

Nitrate-Nitrogen in Coldwater Spring
The nitrate-nitrogen levels in Coldwater Spring (Figure 7) varied between 2.6 ppm to 5.2 ppm during this study.

Figure 6. Comparison of the data logger temperatures with the thermometer temperatures in Coldwater Spring.
3. Coldwater Spring is no longer a constant temperature spring. Coldwater Spring’s temperature fluctuates smoothly by 2.4°C. The temperatures are 3 to 5°C warmer than the temperatures of the spring in 1837. The temperature fluctuations are not in phase with seasonal air temperatures. Coldwater Spring’s waters are coolest in June and warmest in November. Dense urbanization of the region around Coldwater Spring has driven larger seasonal variation in soil temperature and higher temperatures overall.

4. Wetland A is fed by a shallow and more local ground water recharge area than Coldwater Spring. Wetland A has larger temperature fluctuations than Coldwater Spring and is only slightly out of phase with the seasonal air temperature.

5. Coldwater Spring’s nitrate-nitrogen levels rose from about 3 ppm in February 2013 to over 5 ppm in January 2015. The nitrate-nitrogen values in Coldwater Spring vary inversely with chloride. One potential source of nitrate-nitrogen is leaking sanitary sewage lines, but one would expect that input would be a more constant nitrate source year-round. A second source is larger pulses of nitrate from seasonal lawn fertilizer applications which occur during the spring and summer – in different seasons than the road salt applications.

6. Wetland A's nitrate levels peak in the spring at 6 ppm but are generally lower than Coldwater Spring, around 1 ppm through most of the summer and winter. This may reflect the more locally dominated recharge and lack of fertilizer application on the Coldwater Spring property and adjacent roadways.

Summary and Conclusions
1. The temperature and water chemistry of the groundwater flowing from Coldwater Spring and Wetland A appear to have been significantly impacted by human activities.

2. The chloride concentrations in Coldwater Spring’s waters have increased 100-fold between 1880 and the present. The chloride contents in Wetland A’s waters are even higher than those in Coldwater Spring. The chloride/bromide ratios in both springs suggest that road salt is likely the primary source of the chloride. The chloride contents of Coldwater Spring and Wetland A fluctuate in mirror image fashion: one is high when the other is low and vice versa.

3. The large concentrations of chloride in the waters from both sources are not balanced by the sodium concentrations. Cation exchange has occurred along the groundwater pathways in the time between the applications of the road salt, and when the dissolved road salt reaches the springs.

4. Both Coldwater Spring and Wetland A’s waters have relatively constant elevated sulfate ion concentrations of about 80 to 90 ppm. These values are probably produced naturally from the oxidation of sulfides present in the Platteville Limestone.

Summary and Conclusions

1. The temperature and water chemistry of the groundwater flowing from Coldwater Spring and Wetland A appear to have been significantly impacted by human activities.

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Future Work
In addition to the potential for multiple water inputs to the Coldwater Springhouse, NPS staff and the Howe and Lauzon (2014) report identified three additional artificial drains of the Coldwater Spring area – two old drain pipes to the east and one additional drain pipe to the south. All of these places where ground water returns to the surface need to be monitored and their chemical and physical properties documented.

Dye tracing will be a significant tool to determine the possible recharge areas for the Coldwater Spring site in this complex urban environment. Alexander et al. (2001)
documented connections to Coldwater Spring from two, nearby points, but did not monitor any of the other potential resurgences.

A dye trace is currently being conducted at Coldwater Spring. Rhodamine and uranine/fluorescein dyes were poured into two separate rain gardens in the Veteran’s Hospital, located across Highway 55 from Coldwater Spring on 6 June 2015. Eight dye bugs were placed around the Coldwater Spring site to track the movement of the dyes through the groundwater system at Coldwater Spring. Figure 9 shows the various input and monitoring points. We hope to have dye trace results to report by the 14th Sinkhole Conference.

Acknowledgments

A special thanks goes to the National Park Service for partial funding of this project and for NPS staff assistance and encouragement. Another special thanks to the University of Minnesota’s Undergraduate Research

**Figure 9.** Map showing the relationships between the TH62/TH55 traffic interchange, the Minneapolis VA Medical Center and the Coldwater Spring Mississippi National River & Recreation Area. The two blue circles show Coldwater Spring (upper) and Wetland A Spring (lower). The orange and green arrows show the results of 2001 dye traces (Alexander et al., 2001). The two red triangles show the 2015 dye input points in the northern parking lot rain garden (upper) and southeastern parking lot rain garden (lower). The green triangles show the outfall pipes of drainage and storm water drains. The two black crosses are sampling locations under the foot bridges across the Coldwater and Wetland A spring runs. Charcoal detectors were placed and are being periodically changed at the eight locations shown by the blue circles, black crosses and green triangles.
Oppunities Program for partially funding this project and for the opportunity to present the preliminary results at the National Conference on Undergraduate Research in April 2014. SMK would also like to thank her father Hisanoa for taking her sampling every week throughout the school year and the cold Minnesota winter, and to her mother Susan for driving her to and from the University of Minnesota campus during the summer. We thank Betty Wheeler for her editing of this paper.

References


HYDROLOGIC AND GEOCHEMICAL DYNAMICS OF VADOSE ZONE RECHARGE IN A MANTLED KARST AQUIFER: RESULTS OF MONITORING DRIP WATERS IN MYSTERY CAVE, MINNESOTA

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Abstract
Caves provide direct access to flows through the vadose zone that recharge karst aquifers. Although many recent studies have documented the highly dynamic processes associated with vadose zone flows in karst settings, few have been conducted in mantled karst settings, such as that of southeastern Minnesota. Here we present some results of a long-term program of cave drip monitoring conducted within Mystery Cave, Minnesota. In this study, two perennial ceiling drip sites were monitored between 1997 and 2001. The sites were located about 90 m (300 ft) apart along the same cave passage approximately 18 m (60 ft) below the surface; 7 to 9 m (20 to 30 ft) of loess and 12 m (40 ft) of flat-lying carbonate bedrock strata overlie the cave. Records of drip rate, electrical conductivity, and water temperature were obtained at 15 minute intervals, and supplemented with periodic sampling for major ion chemistry and water stable isotopes. Patterns in flow and geochemistry emerged at each of the two drip sites that were repeated year after year. Although one site responded relatively quickly (within 2-7 hours) to surface recharge events while the other responded more slowly (within 2-5 days), thresholds of antecedent moisture needed to be overcome in order to produce a discharge response at both sites. The greatest amount of flow was observed at both sites during the spring snowmelt period. Rainfall events less than 10 mm (0.4 in) during the summer months generally did not produce a drip discharge response, yet rapid drip responses were observed following intense storm events after periods of prolonged rainfall. The chemical data from both sites indicate that reservoirs of vadose zone water with distinct chemical signatures mixed during recharge events, and drip chemistry returned to a baseline composition during low flow periods. A reservoir with elevated chloride and sulfate concentrations impacts the slow-response drip site with each recharge event, but does not similarly affect the fast-response drip site. Nitrate concentrations in drip waters were generally less than 4.0 mg/L as \( \text{NO}_3^- \) (or less than 1 mg/L as N). Nitrate was either stable or slightly increased with drip rate at the fast-response drip site; in contrast, nitrate concentrations decreased with drip rate at the slow-response drip site.

Introduction
The vadose zone is the principal pathway for autogenic recharge of karst aquifers. Many recent studies have been conducted on characterizing vadose zone flow in karst (see Fairchild et al., 2006 and Fairchild and Baker, 2012 for reviews); however, the processes controlling flow through the vadose zone in karst settings are still not very well understood. Due to the potential for high flow velocities and rapid transmission of infiltrating water through the vadose zone in karst, these aquifers may be highly susceptible to pollutants applied to the land surface. In a karst system where a mantle of sediment overlying bedrock is thick, the aquifer might be considered to be more ‘protected’ from the influence of surface pollution (e.g., Döerfliger et al, 1999); however, relatively few studies of cave drip water chemistry have been conducted in mantled karst settings.
Here we present some results of a long-term program of cave drip monitoring conducted between 1997-2001 within Forestville/Mystery Cave State Park, Fillmore County, Minnesota. In southeastern Minnesota, the karst is mantled by a layer of post-glacial loess measuring up to 9 m (30 ft) thick in some areas. Although preferential flow through macropores may allow water to infiltrate rapidly, the fine sediment matrix also retards flow and allows for water retention. The degree to which water is transmitted or retained is thus highly dependent upon antecedent hydrologic conditions.

Fertilizer and pesticide pollutants such as nitrate and atrazine are commonly detected in shallow carbonate aquifers of southeastern Minnesota due to agricultural practices on the surface. These chemicals migrate down through the vadose zone, and may accumulate in the epikarst and/or saturated zone (Kamas et al., 2015; Huebsch et al., 2014). For example, Alexander et al. (1999) reported that an anthropogenic Ca+Mg/Cl+NO₃ chemical component of up to 6 meq/kg is detectable in some Minnesota groundwaters. Early results from the present study indicated that these chemicals are also retained within the vadose zone, in similar concentrations (Doctor and Alexander, 1998). This strong anthropogenic chemical component acts as a marker of young groundwater ages, since this chemical signature is not seen in aquifers that have ages older than the onset of intensive agricultural practices in Minnesota (Alexander et al., 1999).

**Study Area**

This study took place within the eastern portion of Mystery Cave (Figure 1). Mystery Cave is managed by the Minnesota Department of Natural Resources, whose cooperation greatly facilitated this study. The cave is developed in flat-lying limestone and dolomite rocks of Paleozoic age. Two bedrock units contain the majority of the cave passages: the Dubuque Formation, and the dolomitic Stewartville Formation of the Galena Group. The Dubuque is composed of intercalated limestone and calcareous shales, and overlies the more massive Stewartville dolomite.

![Figure 1](image-url)
Two perennially dripping sites within the cave were monitored. Both of these sites are located along the tourist route in the portion of the cave called Mystery II, where the passages are formed primarily within the Stewartville dolomite. In many places the passage ceiling marks the contact between the Dubuque and the Stewartville units, labeled BP1 by Palmer and Palmer (1993), as shown in Figure 2.

The first drip site is named Coon Lake Drips (CLD). Here, water enters the cave as it flows out of a vertical joint fracture in the ceiling, and then runs down the wall of the passage, depositing a rind of calcite flowstone. The water drips onto the floor, and runs across the footpath in the main passage into a small ephemeral pool on the other side called Coon Lake. Coon Lake Drips and Coon Lake have essentially identical water chemistries during periods of relatively high discharge (>2 L/hr) at the drips (Jameson and Alexander, 1994).

The second site is located further down the passage within a room called Garden of the Gods (GG). This room is one of the more highly decorated parts of the cave, exhibiting numerous stalactites and stalagmites and an abundance of flowstone. The dampness and associated speleothem growth at Garden of the Gods indicates that this area of the cave is a significant focal point for discharge of vadose zone waters held in storage above the cave. The soda straw stalactites number in the hundreds, however only three drip sites have been observed to show measurably variable discharge on hourly time scales; these drips have been designated GG1, GG2 and GG3. Only GG1 discharges water perennially and thus was chosen as the site to be instrumented for this study. Since only data from the GG1 drip site is presented in this paper, the site will hereafter be referred to as GG. Water at GG drips off of five distinct stalactites; however, all five extend off of a singular flowstone deposit originating as seepage from a bedding plane parting near to the ceiling.

The land surface above the two drip sites slopes gently to the northeast, and is partially natural grassland, and partially agricultural land. A driveway leading to the entrance of the cave runs along the base of the hillslope, and the passage along which the drip sites are located lies beneath the hillslope, roughly parallel to the driveway. From the chemical data obtained, it is apparent that the drip waters in the cave are affected by the agricultural activity overlying or in the immediate vicinity of the cave.

Figure 2. Map and extended profile of the section of Mystery Cave studied. Cave map provided courtesy of Warren Netherton, Forestville/Mystery Cave State Park; extended cave profile modified from Palmer and Palmer, 1993. Surface profile obtained from LiDAR elevation data provided by the Minnesota Dept. of Natural Resources. BP1, SX1, SX2 indicate bedding plane horizons.
Shallow seismic surveys conducted directly over the two sites indicate a depth to bedrock of 6.7 m (22 ft) over GG, and 8.5 m (28 ft) over CLD. A fairly uniform deposit of loess, laid down by wind during the last glacial retreat, mantles the bedrock. Detailed surveys from the drip sites through the cave and onto the surface indicate that the bedrock strata above the two drip sites are approximately 9 to 12 m (30-40 ft) thick. Water which falls as precipitation onto the surface must infiltrate through the soil, loess, and rock before entering the cave at the drips. Vertical flow paths are generally assumed to control the movement of water from the surface to the drips given the prominent vertical joints in the rock; however, significant horizontal flow paths along bedding plane partings cannot be ruled out, and may play a critical role in determining the hydrologic dynamics of the drips.

**Research Approach and Methods**

In order to characterize temporal hydrologic and chemical changes at the drips, a system was set up whereby the drip waters were funneled into a reservoir that spilled out into a rain gauge, and the flow was continuously monitored for drip rate, water temperature, and specific electrical conductance. The instrumentation setup is illustrated in Figure 4.

The drip water fell into a funnel which rests in a reservoir of approximately 100 ml. The reservoir bottle has holes drilled around the top, below the base of the funnel, to allow water to freely flow out of the reservoir as fresh drip water is added. The water coming out of the top of the reservoir bottle then flows into a tipping-bucket rain gauge, which measures the discharge. Each tip of the rain gauge bucket equals 4.73 mL. Inside the reservoir, within the funnel, rests a conductivity/water temperature probe. The probe and the rain gauge are connected to a data logger. The logger is programmed to take a conductivity/temperature measurement every minute, then to record the average of those measurements every 15 minutes. The logger also sums the number of tips from the rain gauge, and records that sum every 15 minutes.

Water samples were collected approximately twice per month through the first half of 1997, and more frequently during a snowmelt event in 2001. The pH and temperature of the samples were measured in the field, and splits collected for analysis of major cations and anions (15 ml for each) and stable isotopes of water (about 5 ml). The alkalinity of each sample was determined by manual titration within 24 hrs of sample collection. Cations were measured using an inductively-coupled plasma mass

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**Figure 3.** Schematic sketch of rock and sediment above the Garden of the Gods drip site (GG). Only GG1 was instrumented, and is referred to as GG in the text.

**Figure 4.** Drip monitoring setup within Mystery Cave.
spectrometer (ICP-MS), and anions were measured by liquid chromatography at the University of Minnesota. Samples obtained for O and H stable isotope analysis of drip water were analyzed at the USGS in Denver, Colorado, and at the University of Missouri, St Louis. The stable isotope samples were stored in glass vials with polyseal caps to prevent evaporation.

A weather station was set up on the surface, approximately 100 m from the visitor center and near entrance of the main cave. The weather station recorded total rainfall, relative humidity, and air temperature every 15 minutes, and was synchronous with the dataloggers at the drip sites. This enabled records of drip discharge to be directly compared with surface precipitation. Snow depth data were obtained from a weather station at Preston, Minnesota, approximately 18 km from the research site.

Results
The continuous monitoring of the drip waters has illustrated distinct differences between the two sites. The faster-flowing drip site, CLD, shows rapid response to rain and snowmelt recharge. Drip rate at CLD can increase in a matter of minutes after rainfall, depending upon antecedent moisture conditions. Drip rates over 14 L/hr have been measured. In contrast, the drip rate at GG rarely rises above 2.0 L/hr (Figure 5).

The electrical conductivity and water chemistry data at GG has revealed a consistent increase in dissolved ion concentrations with an increase in drip rate, irrespective of the amount of the increase in drip rate. For example, a snowmelt event in early March of 1997 caused the drip rate to increase from less than 0.5 to 2.0 L/hr at GG, and the electrical conductance increased from 0.27 to 0.40 mS/cm. In October the same year, rainfall caused the drip rate to increase to roughly 1.0 L/hr, yet the conductivity increased again to nearly 0.40 mS/cm. More stable water chemistry irrespective of discharge is seen at CLD. The discharge at CLD is consistently greater than that at GG year round (Figure 5). In addition, CLD responds much faster to large precipitation events on the surface (lag time on the order of 15 minutes) than at GG (lag time on the order of two days).

Chemical constituents such as nitrate (NO$_3^-$), chloride (Cl$^-$), and sulfate (SO$_4^{2-}$) have been detected in the drip waters. Measured concentrations range up to 30 mg/L SO$_4^{2-}$, 23 mg/L Cl$^-$, and up to 4.0 mg/L NO$_3^-$ at the GG site. Between 1997 and 2001, concentrations of these constituents were observed to increase at the GG drip site (Figure 6 and Figure 7). The concentrations vary according to the flow conditions, with apparent storage and buildup of these chemicals within a reservoir above the cave. Chloride and sulfate concentrations tend to co-vary, and increase with increases in discharge; in contrast, nitrate concentrations appear to have decreased with increases in discharge at GG in 2001. Although some data on drip rate and conductance was lost in 2001 due to instrument failure, it is evident from the patterns in earlier years (Jameson and Alexander, 1994) that these chemical changes are related to recharge events, coincident with the recharge events recorded at CLD.

Conductivity and Discharge Relations
CLD shows very little change in conductivity with increasing drip rate during snowmelt events or even during large summer storms, except at peak flows. Only during the highest drip rates (> ~10 L/hr) does the conductivity decrease in response to drip rate at this site, exhibiting only a modest change of 0.015-0.030 mS, although that change can occur rapidly. For example, during the snowmelt of 1997, the conductivity decreased from 0.260 to 0.245 mS over the course of 6 hours, from March 27-28. Similarly, during a large summer rain storm event on June 27-28, 1998, the conductivity decreased only from 0.288 to 0.267 mS over the course of 4 hours (Figure 8).

At CLD, the seasonal difference in drip water conductivity (between winter and summer) is greater than the change in conductivity across any single event, and the greatest change occurs during the transition between the end of spring snowmelt and the onset of the growing season. In both 1997 and 1998 the greatest conductivity increase took place from June through August, and after early September the conductivity began to level off and then decrease once again. This behavior is likely the result of the increased partial pressure of CO$_2$ due to biological respiration in the soil zone during the growing season, which allows for greater carbonate mineral solution and/or a reduction in prior calcite precipitation (PCP) in the bedrock above the cave (Wong et al., 2011). Once the growing season ends, plants and microorganisms gradually reduce the production of CO2 due to respiration, and the total dissolved solids in the drip water gradually decreases. In contrast, the changes in conductivity at GG are much greater than at CLD, and...
there is little seasonal influence on the conductivity at GG. The changes in conductivity at this site are a direct function of the drip discharge. Every discernible flow event, irrespective of its magnitude, produces a change in conductivity at the drips that is nearly equivalent in its maximum value (Figure 5).

**Hysteresis**

Hysteresis plots of discharge versus conductivity show distinct patterns between the two drip sites, and among different seasons. A modest increase in drip rate at GG causes a large increase in conductivity, and the conductivity peaks prior to drip rate. At CLD, a rapid
and large increase in drip rate causes little change in conductivity; only at peak drip rates above ~10 L/hr does the conductivity decrease, and during the recession may return to a higher value than that preceding the event (Figure 9).

**Drip Water Oxygen Isotope Data**

As seen in Figure 5, the drip waters show relatively little seasonal variation in δ¹⁸O composition measured with respect to VSMOW (Vienna Standard Mean Ocean Water) over the period studied; however, large individual recharge events can cause rapid changes in the isotopic composition of the drip water. For example, during the spring of 2001, large changes in drip rate at CLD caused changes in the δ¹⁸O of drip water of over 3.5‰ in the span of three months, in spite of minimal change in conductivity (Figure 10). From late March to mid-April, the δ¹⁸O of the CLD water decreased markedly with increased drip rate; however, during a rainstorm in early May, the δ¹⁸O of the water increased with increased drip rate. In contrast, the δ¹⁸O of the drip water at GG showed less variation during the same 3-month time period, spanning a range of less than 0.5‰ (Figure 10).

It is interesting to note that the long-term mean δ¹⁸O isotopic composition values are significantly different between the two sites, with the CLD drips about 0.6‰ lower than the GG drips across the sampled time period.

![Figure 6. Chemical data obtained at both drip sites in the year 1997.](image)

![Figure 7. Chemical data obtained at both drip sites in the year 2001. Loss of data from the GG site between April and May resulted from equipment failure.](image)
Discussion

The drips at Garden of the Gods (GG) show two distinctly different types of water that discharge into the cave: one has a high Cl-SO$_4$ component in addition to the background Ca-Mg-HCO$_3$ chemical signal. A significant increase in discharge at GG is invariably accompanied by the Cl-SO$_4$ chemical signal, however this is not the case at CLD. In fact, when large increases in discharge occur at CLD (from 2 to 14 L/hr), they are often accompanied by decreases in specific conductance, and do not show high chloride-sulfate chemistry.

The water that has an elevated Cl-SO$_4$ component most likely obtains this signature from the spring and fall seasonal application of fertilizers to the agricultural fields above the cave in the form of KCl and CaSO$_4$, from the waste of livestock, or from some other anthropogenic source. It is possible that the cattle pond located just east of the cave entrance across the road may be a reservoir that contributes to this chemical signal (Figure 1). The elevation of the pond surface is approximately 1250 ft, while that of the bedding plane feeding the drips at GG is ~1230 ft (Figure 2). Whatever the source, the elevated Cl component is unlikely to be a natural chemical signal; natural waters collected within several other areas within Mystery Cave have also been analyzed for major ion chemistry, and show low chloride concentrations, generally less than 5 mg/L (Jameson and Alexander, 1994). SO$_4$ could be derived by oxidation of pyrite in the bedrock (Palmer and Palmer, 1993), but would not be correlated with Cl. This means that water with this chemical signal must have recharged after the onset of intensive agricultural activities above the cave, and is relatively modern water.

There is a strong correlation between the rate of flow at the drips at GG, and the chemistry of the drip waters. Only under recharge conditions is the elevated Cl-SO$_4$ component observed. Thus, the water that comes out of vadose storage under recharge conditions may not be very old, while that which contributes to the baseflow of the drips may be older.

The stable oxygen isotope data indicate that the vadose zone above the cave effectively dampens out the annual precipitation variability, which spans a range of approximately 15‰. Factors which may affect the isotopic composition of infiltrating recharge water include direct evaporation, exchange with the water vapor within voids in the vadose zone, and mixing with water already in storage. Transpiration will have a great

![Figure 8](image-url)

**Figure 8.** Detailed time series of conductivity response to drip rate at CLD. Top panel: 5-day record of snowmelt in March of 1997. Bottom panel: 4-day record of a summer rain storm event in 1998 (precipitation and long-term records shown in Figure 5).
effect on the amount of water which actually infiltrates into the soil, but has been shown not to significantly fractionate the isotopic signature of the infiltrating water (Zimmerman et al., 1967).

The isotopic data seem to indicate that neither of the drips show much variation in their stable isotopic composition when sampled on a monthly basis during the study period; however, when sampled with greater frequency during a storm or snowmelt event, the data show that the isotopic composition of the drip waters are indeed subject to rapid change during large recharge events. During the spring snowmelt period of 1997, the monthly sampling was too infrequent to characterize changes.
in the isotopic composition that may have occurred at the drips; however, the snowmelt period of 2001 was sampled more frequently (weekly to sub-weekly) and provided a better record (Figure 10).

Conclusions
At least two reservoirs of water exist in the unsaturated zone above the cave. One has higher chloride and sulfate concentrations than the other. The higher chloride-sulfate water is flushed out of the vadose zone under elevated flow conditions, and is stored as a reservoir within or above the bedrock. The higher chloride-sulfate water is only observed at the slow-response, lower-flow drip site (GG) during recharge events, and is not similarly observed at fast-response, higher-flow drip site (CLD).

One hypothesis for the observed chemical and isotopic behavior is that the water at GG is dominated by piston-flow behavior in the vadose zone, while in contrast, CLD is dominated by better-integrated preferential flow paths leading from the surface to the cave, and thus exhibits more rapid changes in chemistry and discharge. In addition, the water feeding the drips at CLD is not likely to be sourced within the same recharge area as that feeding the drips at GG. The chemistry at GG reflects a simple 2-component mixture, with the higher chloride-sulfate water expressed at the drips only at higher drip rates. Thus, a storage compartment for this water must exist above the cave. It is a possibility that this water comes from the nearby cattle pond, but this hypothesis has not been tested to date. Nevertheless, these chemical dynamics help shed light on the processes that serve to both rapidly transmit and store chemical constituents within the vadose zone in karst systems.

References

Figure 10. Oxygen isotope variability of drip waters compared to drip rate and conductivity during snowmelt of 2001.


CONDUIT FLOW IN THE CAMBRIAN LONE ROCK FORMATION, SOUTHEAST MINNESOTA, U.S.A.

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Abstract
The karst lands of southeast Minnesota contain more than one hundred trout streams that receive perennial discharge from Paleozoic bedrock springs. Several of the Paleozoic bedrock units that provide discharge are karst aquifers. Field investigations into the flow characteristics of these formations have been conducted using fluorescent dyes to map groundwater springsheds and characterize groundwater flow velocities for use in water resource protection.

Recent field work has focused on the Cambrian Lone Rock Formation, a siliciclastic unit consisting of very fine-grained sandstone and siltstone with minor beds of shale and dolostone. The formation is mapped within tributary valleys of the Mississippi River throughout southeastern Minnesota and southwestern Wisconsin. Overlying the Lone Rock is the Cambrian St. Lawrence Formation. Over a dozen streams have been observed to disappear into stream sinks where the upper St. Lawrence is the bedrock unit closest to the land surface. At three of these sinking stream locations, dye was recovered emanating from springs located in the basal St. Lawrence or from springs located in two distinct zones in the Lone Rock. Dye-breakthrough velocities calculated using passive charcoal detectors ranged between 21-214 meters/day at one location and 88-153 meters/day at another. At a third site, automatic water samplers were placed at a spring that had been previously demonstrated to be connected to a St. Lawrence stream sink through dye tracing. In that trace, an eight-hour sampling frequency determined the dye-breakthrough velocity was 314 meters/day.

Based on outcrop and borehole observations in Minnesota, secondary pore networks in siliciclastic-dominated units generally have bedding-parallel and vertically oriented apertures less than a few centimeters. The process by which the bedding-parallel secondary pore networks form remains obscure; some appear to be mechanically developed. However, interstitial carbonate cement within these units leads to the possibility of dissolution being a minor factor in the formation’s groundwater flow characteristics. These dye traces were conducted at three different sites across a twenty-three kilometer distance and are evidence that the siliciclastic Lone Rock Formation has a conduit-flow component similar to that found in carbonate karst aquifers.

Introduction
Southeastern Minnesota is underlain by a sequence of Cambrian to Devonian sedimentary bedrock layers that were deposited in a broad structural depression known as the Hollandale Embayment. In general, the older rocks are dominated by siliciclastic materials and the younger rocks dominated by carbonates (Mossler, 2008). A hydrogeologic framework that describes four prominent karst systems for southeastern Minnesota (Runkel et al., 2013) is based largely on the work of Alexander and Lively (1995), Alexander et al. (1996), and Green et al. (1997, 2002). These include the Devonian Cedar Valley, the Upper Ordovician Galena-Spillville, the Upper Ordovician Platteville Formation, and the Lower Ordovician Prairie du Chien Group.

The karst systems defined in the framework meet the traditional criteria associated with karst, “an integrated mass-transfer system in soluble rocks with a permeability structure dominated by conduits dissolved from the rock and organized to facilitate the circulation of fluid” (Klimchouk and Ford, 2000).
In the last seven years, investigators in southeastern Minnesota have been characterizing sinking streams and rapid flow in two of the siliciclastic dominated formations of the Cambrian system: the St. Lawrence and Lone Rock Formation (Green et al., 2008, 2012). The St. Lawrence consists of well-cemented, thin-to-medium beds of siltstone, dolomitic siltstone, very fine-grained sandstone and shale. The Lone Rock, a formation within the Tunnel City Group, is mostly composed of a fine-grained sandstone and siltstone with interbedded shale and dolostone (Mossler, 2008). Groundwater flow velocities of 35–750 meters/day (115–2460 feet/day) recorded in many dye traces through these units are consistent with conduit flow. However, the lack of large conduit networks in outcrop and borehole observations makes classifying these formations as karst tenuous. Characterization of flow through tracing and breakthrough curves for these units is more consistent with the definition of pseudokarst, “landscapes with morphologies resembling karst, and/or may have a predominance of subsurface drainage through conduit type voids, but lack the element of long-term evolution by solution and physical erosion” (Kempe and Halliday, 1997). Recent work has shown that the millimeter-to-centimeter-sized, bedding-parallel apertures in siliciclastic bedrock of southeastern Minnesota are connected through an anastomosing network of apertures, clustered along discrete (<2m) stratigraphic intervals and found at depths exceeding 200 meters (Runkel et al., 2015). The apertures are more limited in size than in karstic carbonate rock and are commonly associated with distorted bedding interpreted to reflect dewatering features that occurred shortly after burial when the rock was only partially lithified. Therefore, these voids are unlikely to be primarily the result of dissolution as karst is traditionally defined (Stewart et al., 2012 and Runkel et al., 2015).

This paper focuses on the geologic and hydrogeologic setting and results of three recent traces in these siliciclastic units conducted in Houston and Winona counties (Figure 1).

Geologic and Hydrogeologic Setting of Dye Traces

In Houston and Winona counties, bedrock units from the Upper Cambrian through the Upper Ordovician are generally within 15 meters of the land surface and are capped by unconsolidated sediments such as loess, sand, and colluvium (Steenberg, 2014ab; Lusardi et al., 2014). The topography is dominated by a broad plateau of resistant dolostone of the Ordovician Prairie du Chien Group (OPDC). The OPDC is one of the four karst systems identified in southeastern Minnesota and contains many solution-enhanced fractures and cavities such as sinkholes, large conduits, and caves. A region-wide high permeability zone is located at an unconformity between the Shakopee and Oneota Formations of the OPDC (Runkel et al., 2003; Tipping et al., 2006). Catastrophic failures of sewage treatment ponds have occurred where this high permeability zone is close to the land surface (Alexander et al., 2013). The resistant dolostone layer that dominates the plateau is dissected by numerous narrow valleys, especially in the eastern region of the counties near the Mississippi River. In general, the stratigraphy underlying the OPDC is dominated by more easily weathered sandstone, siltstone, and shale layers that are prevalent on slopes and valley floors. It is in these non-carbonate layers that the focus of this report occurs.

Recent work characterizing fracture characteristics at outcrops of the lower Jordan Sandstone through upper Tunnel City Group in the Twin Cities metropolitan area shows that vertical to subvertical fractures tend to preferentially terminate at specific bedding contacts (Runkel et al., 2014). Changes in lithology at these interfaces are also likely responsible for large head
differences noted at a recently installed multilevel groundwater monitoring multiporl system that intersects the same layers that were characterized in outcrop (Runkel et al., 2014).

In deep bedrock settings, the Cambrian St. Lawrence Formation (CSTL) has a low bulk vertical conductivity and a high bulk horizontal conductivity, demonstrating a marked anisotropy (Runkel et al., 2003). The low bulk vertical hydraulic conductivity for the CSTL in deep settings has influenced the general understanding of the hydrostratigraphic properties of the unit and it is generally thought of as a regional aquitard. Characterization of the CSTL as an aquitard is included in the Minnesota Well Rules handbook where it specifically states that “a stratum at least 10 feet [3.05 m] in vertical thickness of the St. Lawrence” is a confining layer (Minn. Dept. of Health, 2011). In southeastern Minnesota where bedrock is shallow and is cut by deeply incised valleys, the CSTL can have highly variable bulk vertical conductivity and high bulk horizontal conductivity (Runkel et al., 2003; Green et al., 2008; Green et al., 2012). In proximity to deeply incised valleys, the ability of the CSTL to behave as an aquitard is tenuous.

A geologic column for Houston County (Figure 2) shows lithostratigraphic and generalized hydrostratigraphic properties for each of the units (Steenberg, 2014a). Hydrostratigraphic attributes have been generalized into either aquifer or aquitard based on their relative permeability. Layers assigned as aquifers are permeable and easily transmit water through porous media, fractures or conduits. Layers assigned aquitard have lower permeability that vertically retards flow, effectively hydraulically separating aquifer layers. However, layers designated as aquitards may contain high permeability bedding plane fractures conductive enough to yield large quantities of water.

**Methods**

The dye traces and geochemical data presented in this report were conducted to further delineate springsheds in the area, characterize sinking streams in the Cambrian St. Lawrence Formation, and describe surface water to groundwater interactions in the counties. Traces focused on locations where surface water was known to be sinking and where local landowner permission was granted. Fluorescent dye was poured into a sinking stream or sinkhole. From there its flow through a conduit system was timed and mapped based on when and where it re-emerged at a spring. Dye was recovered through water grab samples, passive dye receptors (packets of coconut charcoal informally referred to as “bugs”) and, for the Bridge Creek trace, an automatic water sampling device.
Analyses of water samples and charcoal detectors were performed at the University of Minnesota Department of Earth Sciences Hydrochemistry Laboratory using a Shimadzu RF5000 scanning spectrophotometer and PeakFit software. Anions were analyzed for additional direct water samples from springs and streams for two of the traces using a Dionex ICS-2000 Ion Chromatography System. These samples were collected under base flow conditions to further characterize the hydrologic system. Stratigraphic interpretations of spring positions are based on a combination of field outcrop examination and correlations to water well records from the County Well Index (Bauer and Chandler, 2014).

**Dye Tracing Results**

**Bridge Creek**

Bridge Creek in Yucatan Township, Minnesota was the site of dye traces in 2012 and 2013 (Figure 3). It receives discharge from a number of springs that emanate from the basal Jordan Sandstone. The 2012 trace consisted of the introduction of 1.104 kilograms (kg.) of Eosin dye into a discrete sinking stream point on Bridge Creek (MN28:B00006) located in the CSTL. Stream discharge at the time was estimated to be 0.008-0.014 cubic meters per second (0.3-0.5 cubic feet per second).

Eosin dye was detected at levels high enough for positive quantification at a number of sampling sites. Dye was detected at Rostvold Spring which emanates from the Lone Rock (CTLR) 20 – 28 days later. Assuming a straight-line distance from the stream sink (MN28:B00006) to Rostvold Spring, this translates to a minimum peak groundwater velocity ranging from roughly 146 to 205 meters/day (480 ft/day to 670 ft/day). Dye was also detected in charcoal receptors at the Bridge Creek Outlet, the Frauenkron Crossing, the Frauenkron well, and the Bolster spring pond, all within the CTLR. Dye detection for the Frauenkron well was determined using passive detectors placed in their toilet tank reservoir. The well owners had previously stopped using the well for potable water after having multiple gastrointestinal issues. The well has since

![Figure 3. Bridge Creek site sampling locations and dye flow vectors for 2012-2013. Dye flow vectors for the 2012 trace are shown in black. Dye flow vectors for the 2013 trace are shown in yellow. See Figure 2 for definition of map symbols. Note: trace vectors not drawn to dye receptor locations that integrate upstream waters. Geologic map from Runkel et al., 2013.](image-url)
been abandoned and replaced. Borehole geophysics conducted during well abandonment found evidence of fracture flow in the middle to upper Tunnel City from the caliper and fluid resistivity logs.

The year following the 2012 Bridge Creek trace, 24.1 centimeters (9.5 inches) of rain fell in the Bridge Creek watershed between June 21 and 25, 2013. Overland stream flow during this event heavily altered the geomorphology of Bridge Creek, causing large shifts in the stream’s thalweg. Following the precipitation event, surface water was no longer making it down the valley and reconnaissance was conducted to determine where the stream was losing water. Surface water was found to be sinking into a location roughly 610 meters (2000 feet) upstream from MN28:B00006 in the CSTL. A stream sink had previously been speculated to be in this general vicinity based on black and white aerial imagery from 1991. An additional trace was conducted on September 12, 2013 because of this dramatic shift in the sinking stream location. The second dye trace introduced 1.073 kg of Uranine dye into a sinking pool and discrete sinking stream point on Bridge Creek (MN28:B00005) (Figure 3).

Stream discharge at the time was estimated to be 0.003 cubic meters per second (0.1 cubic feet per second). Passive receptors and three automatic samplers programmed to sample at eight hour intervals were deployed during the first seventy days of the 2013 trace. Automatic samplers are ideal for characterizing dye-breakthrough curves (Figure 4). Uranine dye was detected at levels high enough for positive identification at both the passive and active sampling sites. Uranine dye was detected in the automatic samplers at Rostvold Spring (CTRL) 15 days later. Assuming a straight line distance from the stream sink (MN28:B00005) to Rostvold Spring, this translates to a minimum peak groundwater velocity of roughly 314 meters/day (1,031 ft/day). This velocity is consistent with previous traces in the St. Lawrence (Green et al., 2012). Dye was also detected in charcoal detectors at Frauenkron Crossing, Jerry Lee Spring, and Bolster Pond Spring Outlet. Eosin dye used in the 2012 trace was additionally detected a year later during the 2013 trace at Rostvold Spring, Bridge Creek Outlet, Jerry Lee Spring, and Frauenkron Crossing. The long tail on dye recovery observed in this trace is consistent with dye recovery in previous St. Lawrence traces (Green et al., 2008, 2012).

Rhodamine WT dye was detected at levels high enough for positive identification at multiple sampling sites including the Whispering Hills Spring (CTRL) within 15 days and at the Peterson Spring (CTRL) within 14 days (Figure 5). An early breakthrough time for the St. Lawrence “nose” spring was not recorded due to mammals destroying passive detectors early in the trace. In southeastern Minnesota, numerous CSTL and CTRL springs resurge at promontory points at the toe of the slope where incised valleys meet; they are informally referred to as “nose” springs in this paper. Dye was recovered at the passive detectors at the CSTL “nose” spring between September 27 and October 17.

Assuming a straight line distance from stream sink (MN28:B00004) to Whispering Hills Spring, this translates to a minimum peak groundwater velocity of roughly 88 meters/day (290 ft/day) and to Peterson Spring, a minimum peak groundwater velocity of roughly 153 meters/day (503 ft/day).
The 2012 trace consisted of the introduction of 1.166 kg of Uranine dye into a discrete sinking stream point on upper Campbell Valley Creek (MN85:B0020) located in the CSTL. Stream discharge at the time was estimated to be 0.00006–0.0003 cubic meters per second (0.002–0.011 cubic feet per second). Uranine dye was detected at levels high enough for positive identification at multiple sampling sites including the Barnhardt 1 and 2 springs (CSTL) and Pagel’s Big Spring and Power Spring (CTLR). Assuming a straight line distance from the stream sink (MN85:B0020) to the Barnhardt springs, the minimum peak groundwater velocity is roughly 21 meters/day (69 ft/day). Assuming a straight line distance from the stream sink (MN85:B0020) to Pagel’s springs, the minimum peak groundwater velocity ranges between 55 and 214 meters/day (180 ft/day-702 ft/day). Despite being located within the lowermost Lone Rock Formation, these velocities are still consistent with previous traces in the St. Lawrence and uppermost Lone Rock Formation.

The locations of the Strand nose spring and bank spring were not known during the active monitoring stage of this trace and they were not visited until March 4, 2014. At that point passive dye receptors were deployed and water chemistry was collected. Passive detectors showed no evidence of dye at the Strand springs.

**Campbell Valley Creek**

Dye traces were conducted on Campbell Valley Creek in Pleasant Hill Township, Minnesota in 2012 and 2013. Campbell Valley Creek starts at a spring that emanates from the basal Jordan Sandstone and sinks within 50 meters (164 feet) into the upper St. Lawrence Formation (Figure 6). Farther down the valley, water resurges at two small perennial St. Lawrence Formation springs. The Campbell Valley Creek also sinks into the Lone Rock Formation within 10 meters (32 feet) downstream and resurges at the basal Lone Rock Formation farther down the valley. The 2012 trace consisted of the introduction of 1.166 kg of Uranine dye into a discrete sinking stream point on upper Campbell Valley Creek (MN85:B0020) located in the CSTL. Stream discharge at the time was estimated to be 0.00006–0.0003 cubic meters per second (0.002–0.011 cubic feet per second). Uranine dye was detected at levels high enough for positive identification at multiple sampling sites including the Barnhardt 1 and 2 springs (CSTL) and Pagel’s Big Spring and Power Spring (CTLR). Assuming a straight line distance from the stream sink (MN85:B0020) to the Barnhardt springs, the minimum peak groundwater velocity is roughly 21 meters/day (69 ft/day). Assuming a straight line distance from the stream sink (MN85:B0020) to Pagel’s springs, the minimum peak groundwater velocity ranges between 55 and 214 meters/day (180 ft/day-702 ft/day). Despite being located within the lowermost Lone Rock Formation, these velocities are still consistent with previous traces in the St. Lawrence and uppermost Lone Rock Formation.
Nitrate concentrations greater than 1 part per million (ppm) are greater than background conditions and possibly indicate that an aquifer has been impacted by activities on the land surface (Minn. Dept. of Health, 1998 and Wilson, 2012). Nitrate concentrations greater than 3 ppm indicates that an aquifer has been impacted by activities on the land surface (Minn. Dept. of Health, 1998). Chloride concentrations of greater than 5 ppm can also be used to indicate that an aquifer has been impacted by activities on the land surface. Multiple investigators have used Cl/Br ratios to identify chloride sources to groundwater (Davis et al., 1998; Panno et al., 2006). In general, samples with chloride-to-bromide ratios above 300 are waters that have been elevated by human activity.

Anion chemistry collected during the Bridge Creek and Girl Scout Creek dye traces show elevated levels of nitrate, chloride, and chloride-to-bromide ratios in the groundwater of upper stratigraphic units and surface waters of these watersheds (Table 1). These elevated levels of nitrate and chloride can be used as geochemical indicators of recent human influence on groundwater. They can be attributed to the application of road salts, the use of water softeners, and fertilizer application. We used the following classification scheme to assign the term elevated to these anion species.

The 2013 trace consisted of the introduction of 1.843 kg of Rhodamine WT dye into the creek upstream from the stratigraphically lower sinking stream reach located in the Lone Rock Formation (MN85:X0037). Stream discharge at the time was estimated to be 0.008–0.014 cubic meters per second (0.3–0.5 cubic feet per second). The dye input points and receptor locations for both the 2012 and 2013 traces are shown in Figure 6. Dyes were detected at levels high enough for positive quantification at Pagel’s Big Spring and Pagel’s Power Spring. Minimum peak groundwater velocity to the Pagel Springs is roughly 33 meters/day (108 ft/day), assuming a straight line distance from the sinking stream reach (MN85:X0037) to the springs.

**Figure 6.** 2012-2013 Campbell Valley Creek site sampling locations and dye flow vectors. Dye flow vectors for the 2012 trace are shown in yellow. Dye flow vectors for the 2013 trace are shown in black. See Figure 2 for definition of map symbols. Note: trace vectors not drawn to dye receptor locations that integrate upstream waters. Geologic map from Runkel et al., 2013.
moves rapidly downward through enhanced fractures in the Prairie du Chien until it encounters the basal Jordan Sandstone (Figure 7). From there, lateral flow across the basal Jordan resurges as springs in incised valleys that form the headwaters of many creeks. The nitrate concentrations of these headwater streams are elevated above background conditions.

Analytical results of water collected for these investigations verify the presence of anthropogenic signatures from land use in groundwater in this geologic setting. In southeastern Minnesota, wells open to the Prairie du Chien Group are more than twice as likely to yield water with a nitrate concentration above 2 ppm (30.3 percent) than are wells open only to the Jordan Sandstone (12.3 percent) (Runkel et al., 2013).

The marked difference in nitrate concentrations of the Prairie du Chien and the Jordan is attributed to the ability of the lower Oneota to retard nitrate-enriched water downward into the Jordan. However, near the edges of the Prairie du Chien plateau, groundwater moves rapidly downward through enhanced fractures in the Prairie du Chien until it encounters the basal Jordan Sandstone (Figure 7). From there, lateral flow across the basal Jordan resurges as springs in incised valleys that form the headwaters of many creeks. The nitrate concentrations of these headwater streams are elevated above background conditions.

In incised valley settings, rapid vertical flow in a “stair step” like pattern appears to continue through formations underlying the Jordan, with sinking streams identified in both the St. Lawrence and Lone Rock Formations (Figure 7). Groundwater emerging from springs progressively deeper in the geologic section show mixing of nitrate poor water from distant sources up-gradient of the incised valleys. In general, nitrate concentrations of these springs show moderately high mixed signatures or low signatures that are consistent with un-impacted groundwater.

Tritium values of groundwater collected in nearby Wabasha and Fillmore counties show that vintage water, elevated species are likely due to land use activities on the agricultural landscape in these and surrounding watersheds and are similar in concentration to levels found in a regional nitrate investigation of southeastern Minnesota (Runkel et al., 2013). No water samples were collected for anion analysis for the Campbell Valley trace.

### Table 1. Anion chemistry of grab samples collected in the Bridge Creek and Girl Scout Creek Watersheds. Note: ug/g = mg/L = ppm

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Strat</th>
<th>Date</th>
<th>Concentration Units</th>
<th>Detection Limits for 1x Dilution (ug/g)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Bridge Creek</strong></td>
<td></td>
<td></td>
<td>Fluoride ug/g</td>
<td>0.005</td>
</tr>
<tr>
<td>Bridge Creek (surface water)</td>
<td>Collected near lower Jordan subcrop</td>
<td>9/12/2013</td>
<td>0.089</td>
<td>7.594</td>
</tr>
<tr>
<td>Roland Rogers Well</td>
<td>Upper CTGC (15' below top)</td>
<td>9/12/2013</td>
<td>0.075</td>
<td>7.218</td>
</tr>
<tr>
<td>Frauenton Well (Un: # 27859)</td>
<td>Open to uppermost CTGC (40-75' below top)</td>
<td>3/4/2014</td>
<td>0.193</td>
<td>7.140</td>
</tr>
<tr>
<td>Frauenton Spring</td>
<td>Upper CTGC (20' below top)</td>
<td>3/4/2014</td>
<td>0.190</td>
<td>7.520</td>
</tr>
<tr>
<td>Fairy Lee Spring</td>
<td>Upper CTGC (30' below top)</td>
<td>3/4/2014</td>
<td>0.192</td>
<td>7.477</td>
</tr>
<tr>
<td>Restored Spring</td>
<td>Mkt CTGC (75' below top)</td>
<td>3/4/2014</td>
<td>0.087</td>
<td>4.475</td>
</tr>
<tr>
<td>Restored Spring</td>
<td>Mkt CTGC (75' below top)</td>
<td>9/12/2013</td>
<td>0.079</td>
<td>4.278</td>
</tr>
<tr>
<td>Girl Scout Creek (surface water)</td>
<td>Collected near lower Jordan subcrop</td>
<td>9/12/2013</td>
<td>0.104</td>
<td>6.765</td>
</tr>
<tr>
<td>OSC CSTL Norse Spring</td>
<td>Lower CSTL</td>
<td>9/12/2013</td>
<td>0.084</td>
<td>9.585</td>
</tr>
<tr>
<td>Whispering Hill Spring</td>
<td>Upper CTGC (20' below top)</td>
<td>9/12/2013</td>
<td>0.080</td>
<td>9.530</td>
</tr>
<tr>
<td>Peterson Spring Head</td>
<td>Upper CTGC (15' below top)</td>
<td>3/4/2014</td>
<td>0.083</td>
<td>4.554</td>
</tr>
<tr>
<td>Peterson Spring Head</td>
<td>Upper CTGC (15' below top)</td>
<td>3/17/2014</td>
<td>0.097</td>
<td>4.994</td>
</tr>
<tr>
<td>Strand Spring</td>
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<td>3/4/2014</td>
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<td>6.660</td>
</tr>
<tr>
<td>Strand Spring</td>
<td>Upper CTGC (100' below top)</td>
<td>3/17/2014</td>
<td>0.109</td>
<td>6.734</td>
</tr>
<tr>
<td>Strand Bank</td>
<td>Upper CTGC (100' below top)</td>
<td>3/17/2014</td>
<td>0.097</td>
<td>8.266</td>
</tr>
</tbody>
</table>

Table 1. Anion chemistry of grab samples collected in the Bridge Creek and Girl Scout Creek Watersheds. Note: ug/g = mg/L = ppm
in excess of 50 years old, generally is present underlying the St. Lawrence Formation (Petersen, 2005, Zhang and Kanivetsky, 1996). Outside of incised valley settings; the St. Lawrence impedes downward flow and acts as an aquitard.

Dye traces completed from sinking streams in the St. Lawrence and Lone Rock all show similar breakthrough curves. The curves exhibit a rapid breakthrough, rapid rise to a peak, followed by very long (months to years) tails. In February 2015, passive dye receptors were placed at springs where positive dye detection occurred for each of the traces detailed above. The receptors were in place for a two week period and recovered Uranine dye from the 2013 Bridge Creek trace at the Jerry Lee Spring and recovered dye Rhodamine WT dye from the 2013 Girl Scout Camp Creek trace at the Peterson Spring. These dyes were recovered roughly 510 days following their introduction. No dye was recovered from the Campbell Valley trace at Pagel’s Power Spring.

Breakthrough curves from traces in the St. Lawrence and Lone Rock formations are fundamentally different from curves of the Galena and Prairie du Chien formations. Breakthrough curves in these karst units are generally asymmetric with rapid peaks and tails that last hours to days for the Galena and days to weeks for the Prairie du Chien (Alexander, E.C., Jr., oral commun., 2015). The groundwater velocities calculated for the Bridge Creek, Girl Scout Camp Creek, and Campbell Valley Creek traces are consistent with previous traces in the St. Lawrence and Lone Rock where peak rates range from 35 to 600 meters/day (Green et al., 2012).

The connection between surface water and groundwater found in these dye traces has human health and contaminant transport implications that are related to aquitard integrity and aquifer susceptibility in this unique setting. In incised valley settings in southeastern Minnesota, both the St. Lawrence and Lone Rock may be compromised by connectivity to surface water and the ability of water to travel at rapid rates.

Conclusions
Dye trace results, outcrop observations, water chemistry, and data gleaned from borehole geophysics demonstrate...
that the St. Lawrence and the Lone Rock have attributes of both an aquifer and an aquitard dependent upon depth of burial and landscape setting. Dye traces have shown that conduit networks throughout the St. Lawrence in incised valley settings are connected to conduit networks in the underlying Lone Rock Formation and that these networks exhibit high to very high horizontal and vertical hydraulic conductivity within the relatively low permeability rock matrix.

The conduit networks investigated here allow for rapid subsurface drainage but do not exhibit the behavior of solutionally enlarged networks common in karst (rapid rising and losing limbs on breakthrough curves). Instead, breakthrough curves for traces in the St. Lawrence and Lone Rock formations had very long losing limbs that last from months to years. This demonstrates the lack of large scale karst conduit networks and supports recent work suggesting that these features are mechanical in origin and formed very early on in the rock’s history.

Hydrogeologic characterization of Houston and Winona counties will be furthered through the Department of Natural Resources portion of the County Atlas Program. Roughly 100 samples from wells and 20 from springs in each county will be collected and analyzed for cations, anions, trace metals, tritium and stable isotopes. These data will allow us to better resolve the interaction of recent and older regional groundwater, especially in the incised valleys of these counties.

**Acknowledgments**

The work presented in this report could not have occurred without the permission of landowners who graciously allowed access to their property. This effort was conducted as part of the Innovative Springshed Mapping for Trout Stream Management-Phase II as funded by the Minnesota Environment and Natural Resources Trust Fund as recommended by the Legislative and Citizen Commission on Minnesota Resources (LCCMR). Calvin Alexander, Jr. of the University of Minnesota Earth Sciences Department performed sample analysis and interpretation. Special thanks are given to Holly Johnson for her graphic editing assistance.

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Abstract
Regression analysis is used to identify monotonic trends to assign water age using ion data from two large well water databases from southeast Minnesota (SE MN). Nitrate (NO$_3$-N), chloride (Cl), sodium (Na), and sulfate (SO$_4$) ions in the commonly used aquifers in SE MN can be used as groundwater tracers since they are either entirely or partly anthropogenic in their sources, their loading occurs on a regional scale, and they are almost entirely conserved.

Ion concentrations over time are used to establish six trend patterns. Two patterns are unchanging (background and stable above background), and four are changing (linear up, exponential up, peaking, and down). These patterns are then used to assign specific age values or age ranges to the well based upon that ion. For ions with linear upward trends, specific ages are derived from the “Intercept Year” representing a time when water extracted from the well first infiltrated from the land surface containing the ion at detectable concentrations and the “Marker-Year” representing the beginning of large scale trend changes for that ion source.

Introduction
Public agencies have collected large amounts of environmental data in recent times as part of numerous programs some have limited purposes and time frames, others are broader and open ended. Data analysis for groundwater quality has often been limited to defining what is in the drinking water today.

This work relates trends in water ion concentrations over time to decadal changes in ion loading sources to estimate the “age of water” now extracted from wells. It varies from traditional groundwater age dating in that it relies on changing patterns in the regional or national scale related to continuous loadings. This contrasts with introduced chemical or isotopic markers, which generally have specific release events or time periods. In addition, this method relies on commonly monitored analytes from already available sources.

Data from two databases is used to look at statistical trends in groundwater quality from residential wells over time. First is the Southeast Minnesota Water Analysis Lab (SEMWAL) database that provides data from Olmsted County from 1970-2014 (SEMWAL, 2015). Much of the data in the SEMWAL database results from required testing during property transactions. Over 50,000 samples have been processed by the lab, 19,337 of which are for private wells in Olmsted County including 7,048 distinct addresses. An attempt is made to assign addresses to specific wells. Where five or more years of data is available, trend analysis is done for that address -- 590 for NO$_3$, 446 for Cl, and 430 for SO$_4$. The monitoring period record varies significantly by well.

The second database used for this analysis is the Dakota County Ambient Groundwater Quality Study (AGQS), which has sampled from 24-80 wells annually since 1999 (AGQS, 2015). Wells were chosen to reflect both geographic and aquifer differences in the county. The goal was to determine baseline water chemistry, characterize the occurrence of anthropogenic compounds, conduct trend analyses, and identify land use factors influencing the three most heavily used drinking water aquifers in the County. The database includes seventy-six (76) wells that have been sampled up to 13 times each since 1999.

Anthropogenic Sources and Background Concentrations
Nitrate (NO$_3$-N), chloride (Cl), and sulfate (SO$_4$) are highly mobile in groundwater in SE MN. The first two are mainly attributable to anthropogenic sources while SO$_4$ has a natural component. A 1994 tritium study of 67 private wells in Olmsted County found that NO$_3$-N, Cl, and SO$_4$ were lowest in vintage water (<1953) and highest in recent water (>1953). Nearly all vintage age water had no detectable NO$_3$-N or Cl.

Nitrate and Chloride
Background or natural concentrations for NO$_3$-N and Cl in SE MN groundwater used for domestic drinking wa-
Fertilizer use in Minnesota increased substantially from 1940 to 1955 and then about doubled in the following five years (Dahl, 1970). By 1964, nitrogen use was 20% of today’s rate and potash about 30% (USDA NASS).

Regression analysis was used to establish Marker-Years used in this study to represent the first large scale change in fertilizer source loads. Minnesota use statistics yielded a Marker-Year for nitrogen of 1962 and potash 1960. National use statistics yielded Marker-Years of 1955 for nitrogen and 1948 for potash (USDA NASS). Not enough county specific data is available to calculate Marker-Years.

Salt use increased dramatically between 1960 and 1970, declined during the 1970s, and has been generally increasing since the 1980s. There is substantial annual variation in road salt use. The Marker-Year for salt based on MNDOT data is 1959 (MNDOT, 2014, UMN, 2008), and the Marker-Year based on national statistics for total salt use is 1955 (Kelly, 2013). Figure 1 illustrates the ion source use changes and Table 1 summarizes the Marker-Years used in calculating water age for those sources.

**Sodium**

Sodium from road salt and softener salt is partially retained by clay minerals through ion-exchange reactions and is not delivered to groundwater as completely as are chloride and nitrate. Sodium has a natural background in groundwater from geologic weathering processes. Sodium in the absence of chloride in well water is one basis for defining the natural background. Sodium is highly correlated with chloride, and regression of Cl and Na yields a background of about 4 mg/l when Cl is not detected. The correlation also indicates that sodium has a lower molar concentration than chloride due to its retention by soils and due to a portion of the Cl originating from potash fertilizer (KCl). Sodium to chloride (Na:Cl) ratios reflect the proportion of Cl from agriculture versus road salt or softener salt (AGQS, 2015).

**Sulfate**

Background or natural concentrations of sulfate may be derived from gypsum in limestone and sandstone and from the weathering of iron pyrite (FeS2) in shales (Roberson, 1989). Sulfate is also contributed by atmospheric deposition from the combustion of fossil fuels. By the mid 1970s, the implementation of the Clean Air Act reversed the rapid increase in Sulphur Dioxide (SO2) emissions that had begun in the late 1950s. The downward trend in SO2 emissions continues (Smith 2004, Smith, 2011). The National Atmospheric Deposition Program (NADP) monitoring data confirm the decline in deposition in the Midwest. A regression analysis of national SO2 emissions data identifies a peak in 1970.
Sulfate moves relatively freely with infiltrating water in the SE MN groundwater environment. Common salts of \( \text{SO}_4 \) are generally soluble in the range of concentrations found in SE MN, and redox conditions for converting \( \text{SO}_4 \) to sulfur or hydrogen sulfide exist only in very deep groundwater not ordinarily used for domestic purposes (USGS, 2009). Sulfate, like \( \text{NO}_3 \)-N, is involved in biological processes and can be temporarily retained in soils or biological materials. The many \( \text{SO}_4 \) trends found during this study illustrates the ion’s mobility.

### Method

Statistical analysis using Kendal-Thiel (KT) nonparametric regression is employed to identify monotonic trends - not necessarily linear (Helsel, 2002, Helsel, 2012). At least 5 samples in separate years are needed to detect trends at the minimum 90% statistical significance level accepted for this work. Of the approximately 7,000 wells in the Olmsted dataset approximately 10% meet the criterion for \( \text{NO}_3 \)-N while only 8% meet that criterion for \( \text{Cl} \) and \( \text{SO}_4 \). This reflects the longer period of record for \( \text{NO}_3 \)-N (1970-2014) than \( \text{Cl} \) and \( \text{SO}_4 \) (1988-2014).

The choice of significance level could, in principle, affect whether a well is identified as trending or not. For the wells in the two datasets analyzed, all but a few, which show 90% significance also show 95% or greater significance. Where the trend pattern is significant but

### Table 1. Marker-Years for Ion Source Loadings.

<table>
<thead>
<tr>
<th>Ion Loading Source</th>
<th>Marker-Year by Data Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nitrogen Fertilizer</td>
<td>Minnesota 1962, US 1953</td>
</tr>
<tr>
<td>Potash Fertilizer (KCl)</td>
<td>1951</td>
</tr>
<tr>
<td>Road Salt (NaCl)</td>
<td>1959</td>
</tr>
<tr>
<td>Total Salt Use (NaCl)</td>
<td>1955</td>
</tr>
</tbody>
</table>
not necessarily linear, quadratic regression is tested to see if the well is peaking or leveling off. Wells without a trend were categorized as background or stable above background based on NO$_3$-N and Cl values. This analysis produced the six types of statistically trending patterns discussed below.

For the upward trend pattern wells, the KT regression yields an “Intercept-Year”. The Intercept-Year is a time when the water now extracted from the well first infiltrated from the land surface. The Intercept-Year is a reference marker for each ion, which represents the lag time or age of the water relative to the Marker-Year for the specific ion source. This provides a specific numerical estimate of the water age.

The calculated Intercept-Year is valid only for the “c” portion of the logistic curve which is assumed to have a minimum slope of 3%. This rate reflects the large loading increases beginning in the 1950s. Lesser relative slopes are assumed to be in the “b” and “d” portions of the logistic curve and can’t be used to calculate an appropriate Intercept-Year.

The other five trend categories do provide a range of upper and/or lower limits to the age of the water in the well but no Intercept-Year can be calculated from those trends.

**Trend Patterns for Groundwater Ion Concentrations**

Figure 1 is an example of a long-term ion (NO$_3$-N) pattern in a well, which is represented mathematically as a logistic (sinusoidal) curve. We expect that any well experiencing a change in an ion loading from surface sources would have an upward trend curve similar to that shown in Figure 2 or its mirror image if a downward trend.

The logistic curve contains five distinct trend patterns (a-e in Figure 2). In the earliest period “a”, the ion is stable at background concentrations, followed by periods of rapid increase “c” and then a relative stable period at higher concentrations “e”. The transition at “b” represents an exponential increase which is modeled as a quadratic curve. The transition at “d” is also modeled as a quadratic curve and represents a peak or leveling off.

---

**Figure 2.** Logistic Curve Model Applied to NO$_3$ in Well 11323 from Olmsted County. a-Background Non-Detect (ND), b-Exponential Up (Quad+), c-Linear Up (UP), d-Peaking (Quad-), e-Stable Above Background – No Trend (NT)
The other pattern identified in this work is a decreasing pattern, which is a mirror image of that in Figure 2.

The five trend patterns in the logistic curve can be grouped in two categories: unchanging (background, and stable above background), and changing (exponential up, linear up, and peaking). The sixth trend pattern represented by the exponential downward pattern is also in the changing category (Figure 6).

All of the patterns and transitions represented by Figure 2 and its mirror image are seen in the well monitoring records. The statistical analysis is the basis for classifying well by trend pattern. In most cases the monitoring period contains only a portion of the logistic curve. Few wells in the dataset show the complete logistic curve primarily because of the short period of monitoring record and infrequency of monitoring.

The graph of Well 11323 illustrates a logistic curve for Cl and its mirror image for SO$_4$. Figures 3 through 5 show a side-by-side comparison of the respective ion sources and ion trends in the well. The well ion concentrations reflect the trend patterns in the respective ion sources. In this particular well, the lag time between the ion source loading (Marker-Year) and well ion concentrations (Intercept-Year) is approximately 40 years. The lag time is similar for all three ions.

Well 11323 illustrates that a long record is necessary to be able to see all or most of the logistic curve. In addition, the period of record would also have to reflect the lag time (water age) for the well.

The Intercept-Year from the linear portion of the logistic curve is not the exact asymptotic intercept year. As can be seen in the example in Figure 1, the linear intercept year can vary slightly from the asymptotic year. In the logistic model, the initial increase is exponential and not linear as is assumed with the linear trend regression.

Figure 6 illustrates a portion of the mirror image of Figure 2, the downward trend portion of the logistic curve. In well 695881, the trend would be statistically modeled as a quadratic or exponential. This downward trend correlates with a reduction in NO$_3$-N loading consistent with the area’s conversion from agriculture to mixed residential in the early 1990s.

**Ion Trends for Olmsted County Wells**

Table 2 summarizes the six trend categories for the subset of the Olmsted County data that meet the 5-year criterion. More non-detects are seen with NO$_3$-N (>50%)

![Figure 3. Minnesota Nitrogen Fertilizer Use Trends and the Nitrate Concentrations in Well 11323.](image-url)
Figure 4. Minnesota Potash Fertilizer Use Trends and the Chloride Concentrations in Well 11323.

Figure 5. United States SO$_2$ Emissions Trends and the Sulfate Concentrations in Well 11323.
than with Cl (19%). Even with the longer NO$_3$ record (44 years) there are fewer detections than Cl (26 years). The fraction of NO$_3$-N non-detects in the trend well subset [430 NO$_3$-N, Cl, SO$_4$ and 590 NO$_3$-N only] used in the trend analyses are consistent with the entire Olmsted County dataset of 7,048 records. Trends wells had 55% non-detects versus 62% in the full Olmsted County dataset. There is a wider range in the Cl non-detects (33% vs 19%). The subset of wells used for trend analysis is more representative of the larger dataset for NO$_3$-N than Cl. There are fewer NO$_3$-N trends, either up or down than is the case with either Cl or SO$_4$ (11% vs 27% & 29% respectively). Chloride and SO$_4$ show similar numbers of upward trends (25 vs 23).

We calculate that more than 1,400 wells with 6 or more years of ion data are needed to be able to find one statistical logistic curve. It is thus fortunate that a couple of such curves were found in the Olmsted County dataset. Portions of logistic curves are easier to identify and several of these were also found.

The year categories in Table 3 are based on the ion source trend patterns presented in Figure 5. These categories reflect the differing trend patterns during these general periods. This allows the placement of wells into age brackets by trends for each ion. This provides a test for the calculated Intercept-Year and the consistency of age ranges for the other five trend patterns.

<table>
<thead>
<tr>
<th>Trend Pattern</th>
<th>Nitrate</th>
<th>Chloride</th>
<th>Sulfate</th>
</tr>
</thead>
<tbody>
<tr>
<td>Background (ND)</td>
<td>322</td>
<td>85</td>
<td>0</td>
</tr>
<tr>
<td>Stable Above</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Background (NT)</td>
<td>204</td>
<td>239</td>
<td>304</td>
</tr>
<tr>
<td>Linear Up (UP)</td>
<td>28</td>
<td>98</td>
<td>75</td>
</tr>
<tr>
<td>Linear Down (DN)</td>
<td>16</td>
<td>9</td>
<td>19</td>
</tr>
<tr>
<td>Peaking (Q-)</td>
<td>18</td>
<td>14</td>
<td>24</td>
</tr>
<tr>
<td>Exponential Up or Logistic (Q+)</td>
<td>2</td>
<td>1</td>
<td>8</td>
</tr>
<tr>
<td>Total</td>
<td>590</td>
<td>446</td>
<td>430</td>
</tr>
</tbody>
</table>

Table 2. Olmsted County Wells by Trend Pattern.
Analysis of the assigned trend pattern for each well is shown in Table 3. Only wells with data for all three ions are represented. 164 of the wells (38%) during their monitoring period are associated with water sources preceding 1960. 211 of the wells (49%) are calculated to have water younger than 1970 based on their monitoring period.

For the 85 wells with upward trends for either NO$_3$-N and/or Cl, Intercept-Years were calculated. 75 of those wells (88%) had their Intercept-Years in the predicted category in Table 3. When there is more scatter in the data or the projected Intercept-Year is further from the monitoring period, there is greater uncertainty in the Intercept-Year.

The age classification in Table 4 is based on the best fit of all three ions. There is some overlap in the age periods for the 430 wells due to multiple combinations of age trend patterns in Table 3.

**Ion trends for Dakota County Wells**
The Dakota County well database consists of 76 wells monitored annually from 1999 to 2013. The monitoring period for individual wells in this database differs from that of the SEMWAL database. The SEMWAL database represents a larger number of wells and often with a longer monitoring record.

Table 5 summarizes the results of trend analysis for 76 Dakota County wells from 1999 to 2013. Twenty-two (29%) of the 76 Dakota County wells have changing NO$_3$-N trend patterns, of these 72% are upward or recently peaking. Thirty-three (43%) of the 76 have changing trend patterns for Cl with all but two upward or recently peaking. Thirty (39%) of the 76 have changing trend patterns for SO$_4$, 25 of these are up and five are down. Twenty-six (34%) of the wells have trends for Na and only one is down. Fifteen of these 26 also have Cl trends and again only one is down (the same as for Na).

**Comparison of Dakota and Olmsted Well Trends**
The Dakota dataset includes a smaller fraction of non-detections of both NO$_3$-N (46% vs 55%) and Cl (19% vs 22%) when compared to the Olmsted data. In addition, there is a larger fraction of NO$_3$-N (13% vs 5%) and Cl (37% vs 22%) linear up trends in Dakota County. The Dakota data also includes Na which has a similar linear up trend pattern as Cl (37% vs 32%). The proportion of SO$_4$ linear up and linear down trends vary less between the Dakota and Olmsted data sets – up trends (24% vs 17%) and down trends (5% vs 4%).

**Discussion**
This method has just two variables, ion source and the time lag between the source and ion level in extracted water over time. Unlike hydrologic models, it is not dependent on the hydrogeologic setting variables such as gradient, porosity, permeability, or any other flow variables. Knowledge of the ion source pattern at a regional scale allows an estimate of the time lag.

This method can be applied to a single or a large set of wells. The degree of consistency in predicted ages or age ranges based upon NO$_3$-N, Cl, Na, and SO$_4$ suggest that any one of these anions could be used alone to predict water ages. However, using a combination of anions increases confidence in the results. Public wells with longer records and more frequent testing for a greater number of analytes would be particularly suitable.

### Table 3.
**Expected Well Trends Based on Source Trend Patterns.**
<table>
<thead>
<tr>
<th>Period</th>
<th>Nitrate</th>
<th>Chloride</th>
<th>Sulfate</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt;1960</td>
<td>ND</td>
<td>ND</td>
<td>NT or Quad+</td>
</tr>
<tr>
<td>1960-1970</td>
<td>ND, NT or UP</td>
<td>ND or UP</td>
<td>UP</td>
</tr>
<tr>
<td>1970s</td>
<td>ND, NT or UP</td>
<td>ND, NT or UP</td>
<td>NT or Quad-</td>
</tr>
<tr>
<td>&gt;1980</td>
<td>ND, NT or UP</td>
<td>ND, NT or UP</td>
<td>Down</td>
</tr>
</tbody>
</table>

### Table 4.
**SEMWAL Wells Categorized by Matching Well Trend Pattern to Ion Source Trend Periods from Table 3.**
<table>
<thead>
<tr>
<th>Source Trend Pattern Periods</th>
<th># Wells</th>
</tr>
</thead>
<tbody>
<tr>
<td>Older than 1960</td>
<td>164</td>
</tr>
<tr>
<td>1960-1970</td>
<td>28</td>
</tr>
<tr>
<td>1960-1980</td>
<td>13</td>
</tr>
<tr>
<td>1970s</td>
<td>4</td>
</tr>
<tr>
<td>Younger than 1970</td>
<td>186</td>
</tr>
<tr>
<td>Younger than 1980</td>
<td>21</td>
</tr>
<tr>
<td>Likely Land Use Change</td>
<td>14</td>
</tr>
<tr>
<td>Total</td>
<td>430</td>
</tr>
</tbody>
</table>

### Table 5.
**Dakota County Wells by Trend Pattern.**

<table>
<thead>
<tr>
<th>Trend Pattern</th>
<th>Nitrate</th>
<th>Chloride</th>
<th>Sodium</th>
<th>Sulfate</th>
</tr>
</thead>
<tbody>
<tr>
<td>Background</td>
<td>35</td>
<td>17</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Stable Above Background</td>
<td>19</td>
<td>26</td>
<td>50</td>
<td>46</td>
</tr>
<tr>
<td>Linear Up</td>
<td>10</td>
<td>28</td>
<td>24</td>
<td>18</td>
</tr>
<tr>
<td>Linear Down</td>
<td>7</td>
<td>1</td>
<td>1</td>
<td>4</td>
</tr>
<tr>
<td>Peaking</td>
<td>4</td>
<td>3</td>
<td>1</td>
<td>7</td>
</tr>
<tr>
<td>Exponential Up or Logistic</td>
<td>1</td>
<td>1</td>
<td>0</td>
<td>1</td>
</tr>
<tr>
<td>Total</td>
<td>76</td>
<td>76</td>
<td>76</td>
<td>76</td>
</tr>
</tbody>
</table>
The SO₂ loading peak in the mid-1970s provides a second marker for calculating a specific age (Figure 2). In addition, the apparent correlation between SO₂ emissions and SO₂ in Figure 2, and the availability of a much longer SO₂ emissions record (1850-2014), suggests that SO₂ will be useful in differentiating the age of wells with non-detectable NO₃-N and Cl. Significant decreases in SO₂ loadings during historic recessions might be particularly useful age markers.

The SEMWAL database contains an additional 34,000 NO₃-N, Cl, and SO₂ records for wells in counties adjoining Olmsted. That data could be used to complete similar studies. Because these counties have historically applied a less rigorous NO₃-N standard for well development, they likely would have more wells with trends useful for calculating the Intercept-Year.

Using the well from Figures 1 – 4 as a case study illustrates the potential value of this analysis for individual wells. Well 11323 serves a bar and grill that is licensed by the Olmsted County Health Department. Based on the estimated 40-year age of the water, it is not likely that the well will present a health hazard based on bacteria or viruses since those organisms would no longer be viable. The decline in the rate of increase in NO₃-N and Cl concentrations in the well in recent years corresponds with the smaller rate increases in fertilizer and road salt use. The Cl: NO₃-N ratio is about 4:1 suggesting that over half of the Cl is from road salt or softener salt. The current NO₃ level is about 50% of the drinking water standard (10 mg/l) and Cl about 10% of the secondary standard (250 mg/l). Given the 40-year historical relationship between loading rates and NO₃-N and Cl concentrations in the well, the well is not likely to exceed either standard in the near future. Considering the water age, this would not be a useful monitoring well for best management practices being implemented today. If in the future monitoring is expanded beyond basic parameters they should include both agricultural as well as non-agricultural pollutants, particularly those used in earlier decades.

References
Minnesota Pollution Control Agency (MPCA), Minnesota Department of Agriculture (MDA).


Abstract
The Mitchell Aquifer averages 80 m in thickness and underdrains a karst region in the Crawford Upland and Mitchell Plateau region in south-central Indiana (110,000 km²). The Springville Escarpment is a transitional boundary between the upland and plateau. Cave stream linking between cave tiers in the aquifer and correlation of cave tier inception horizons to a base level decline surface is interpreted for the Kirby Watershed, encompassing the prekarst headland of Indian Creek (42 km²). The watershed was severed from lower Indian Creek at Eller Col by limestone cavern drainage on the ridge between White River and East Fork. Correlation of recharge basin topography and cave tiers is possible owing to the observation of 55 karst springs confined to lithostratigraphic contacts at three spring stratigraphic levels. Karst Spring Cutoffs are a specific type of vadose canyon diverting cave streams, bypassing around springs and passing into the laterally offset cave streams in the next lower cave tier. Cutoffs connect upper to middle tier cave streams and middle to lower tier cave streams as they enlarge below sinking stream basins and tributary spurs. Three speleogenic enlargement cycles characterize the eastern Leonard Springs Area, but only two cycles have enlarged in the western Garrison Chapel Area.

Introduction
Engineering projects use geotechnical borings in support of design and construction. This is not sufficient for karst regions where environmental and hydrogeological studies are necessary both before design phase and ongoing during construction phase. Dye tracing is a primary method to gain information about cave stream flow patterns and linking between multiple cave systems as it pertains to structural concerns and identifying base level decline surfaces.

Several local base levels may be present for a study area and it is necessary to identify the one that is relevant to a concern, and may be determined in part by examining the speleogenesis, rock unit lithology, and spring stratigraphic levels. Base level flow at a spring often becomes groundwater recharge through fissures channel underflow draining to the next lower base level spring in a region. Recharge is passed between multiple cave levels or cave tiers by karst spring cutoffs. Karst spring cutoff is a new term to describe a specific type of cave stream diversion associated with springs and linking between two cave tiers. A more detailed description and illustrated example follows in this paper.

Fluvial base level decline, lithology, and water chemistry are the primary agents controlling speleogenic enlargement of a vadose cave stream after speleogenic breakthrough. This study demonstrates a geologic description of a karst region by using the framework of stratigraphy and the cycle of speleo enlargement after breakthrough. The shallow karst spring cutoffs and the analog deeper vadose canyons are useful for identifying the cave stream linking across multiple cave tiers and have a geomorphic significance as points preserved from a former base level surface. The spring that was cutoff originated at grade with a former base level surface. The cutoff passage drains to a spring that enlarged at grade with a younger base level surface.

The topographic surface correlates with the present fluvial base level decline surface. Former base level decline surfaces are difficult to map because they are unconformable surfaces caused by differential erosion. This study projects an extended base level surface from a spring channel across a ridge with sinking stream basins where groundwater infiltrates into the karst aquifer, and across to the next spring channel.

Methods
This study is based on lithology and on detailed stratigraphy, measured rock unit profiles, and observation that many of the carbonate units are not entirely isotropic, but are very consistent in their matrix properties, joint patterns, and intensity of fractures over sufficient distances where a trunk cave stream enlarged in a lithostratigraphic interval. Trunk cave streams on cave maps were color indexed by rock units. The observation that 55 springs in the 80 m thick carbonate sequence are confined at speleogenic breakthrough to three discrete spring stratigraphic horizons provides the framework for delineating three cave tiers. Each trunk cave stream and
spring can be fitted into a cave tier and allows the observation of multiple cycles of speleogenic enlargement in a tier across a watershed.

**Concepts**
The cave stream linking involves diversions beyond the general hydro-geometry characteristics described by White and Deike (1989). Karst spring cutoffs at Kirby Watershed are associated with full flow springs and represent a different aspect of cave tier linking beyond the elementary classification of overflow and underflow spring diversions described by Worthington (1991), and karst spring cutoffs are a different phenomenon in addition to the limited basin underflow described by Ray (1997).

The terms cave level and cave tier are generally synonyms, but more precisely level applies to hydrologic interpretation and tier has a stratigraphic connotation. The speleogenic states of inception and enlargement were described in the context of piezometric limits or cave levels by Palmer (1987, 2003) and for cave tiers by White (1988). One or more inception horizons may enlarge within the presently defined cave tiers, as inception horizons were defined by Lowe (1992).

Cave levels and cave tiers are often described as stacked vertically and having local diversions out of and back into trunk cave streams. At Kirby cave tiers are stacked vertically, but additionally have a laterally offset cave stream link between tiers where a karst spring cutoff diverts from a position upstream from the spring and passes below a narrow upper slope tributary or below the width of a larger sinking stream basin. This overview of Kirby Watershed is supported by an atlas of individual cave descriptions and a developing karst geology database for south-central Indiana.

**Location**
The Kirby Watershed includes a 42 km² area of karst drainage severed from the former headland of Indian Creek located in south-central Indiana (Figure 1). Kirby Watershed is centered about the unincorporated village of Kirby at the intersection of Airport Road and Kirby Road: Latitude 39.14 degrees North, Longitude 86.61 degrees West; WGS 1984.

Kirby Watershed is underdrained by more than 34 km of mapped cave passages in three stratigraphically defined cave tiers and 55 springs confined at speleogenic breakthrough to three spring stratigraphic levels. The watershed is divided into the Leonard Springs and Garrison Chapel structural areas including a pattern of eleven karst valleys with perimeters bounded by topo-
graphic ridges and cols (Figure 2). The reconstructed branches of prekarst and early karst tributaries to Indian Creek are identified for Kirby Watershed (Figure 2). Each karst valley evolved around a pre-karst tributary branch that was later modified to the shape of one or more internal sinking stream basins and underdrained by the successive cave tiers (Figure 3).

**Earlier works**
The name Kirby Watershed is applied to the former headland branch of Indian Creek with subterranean stream piracy and karst geology described by Beede (1911). Malott (1922) illustrated thirty-six ponors (swallets) and nine storm water rises (springs) in Kirby Watershed on a township map with reconstruction of the pre-karst drainage channels. The western portion of the watershed near the village of Blanche includes a karst valley with pre-karst drainage illustrated by Wayne (1950). Powell (1965) described karst hydrology for four westward flowing cave streams near Blanche. DesMarais (1973) compiled cave maps and descriptions of caves in the area now described as Kirby Watershed. DesMarais (1981) used mercaptan air tracings to identify four cave streams and flow abandoned upper levels in the western portion of the former headland of Indian Creek referred to as the Garrison Chapel Area (1981) (Figure 2). Reconnaissance stratigraphy in the region was developed by Malott (1952). Precision cave stratigraphy was mapped by Conner (1987).

**Rock Units**
The Mississippian Period Blue River Group averages 80 m in thickness and comprises a major karst forming unit in south-central Indiana, western Kentucky, and southern Illinois. The upper half includes multiple and truncated shoaling oolite cycles. The lower half of the group includes dominantly organic stained micrite, pellet, and dolomitic beds with intercalated sparite and coarsely crystalline biocalcarenite beds. The shoaling cycle starts with pelletal muds, is followed by calcarenites and oo-

![Figure 3. Kirby Watershed former headland of Indian Creek, 11 karst valleys and interior sinking stream basins.](image3)

![Figure 4. Geologic Column for Mitchell Aquifer and Kirby Watershed, upper case named units, lower case informal units.](image4)
lites, and is capped by a thin shaley or sandy limestone. The pelletal mud interval observed throughout caves and core holes in south-central Indiana is populated by a high purity lithographic textured micrite which is very brittle, has closely spaced conchoidal fractures, and reveals a ductility contrast to the interbedded oolitic calcarenites. The micrite is often thinly bedded, 5 to 7 cm, in a 5 m interval. The oolitic calcarenite beds are massive up to 7 m thickness. Mississippian carbonates in the Kirby Watershed are shown on the geologic column (Figure 4).

**Mitchell Aquifer**

The Mitchell Aquifer is introduced as a hydrostratigraphic unit concurrently with description of three cave tiers and karst hydrology in the Mississippian Period Blue River Group in south-central Indiana. The name Mitchell is given as a geographic reference to the city in nearby Lawrence County, Indiana. The name Mitchell was also used in the stratigraphic name Mitchell Limestone with its original interval expanded and later modified, as it was reviewed by Smith (1986) and was replaced by the Blue River Group named by Gray et al. (1960). Karst conduits, springs, and cave streams in the Mitchell Aquifer extend across a ten county area in the Crawford Upland and Mitchell Plateau in south-central Indiana (110,000 km²).

**Physiography**

Kirby Watershed with sinking stream basins evolved athwart a headland mass of the Crawford Upland dividing the drainage between White River and East Fork. Local relief for the Crawford Upland is 90 m and for the Mitchell Plateau is 60 m north of East Fork River. The transitional boundary between the upland and plateau is characterized by the Springville Escarpment (Gray, 2000).

**Bedrock Structure and Flow Zones**

Measured rock unit profiles in the caves were used to map a structural datum on top of the Indian Creek Limestone Beds for comparison to the surveyed cave stream altitude trends (Conner, 1986). An apparent dip of 7 m/km to the southwest is interpreted from the structure contours (Figure 5).

Upper to middle cave tier links through karst spring cutoffs are shown by color indexed lines and illustrate reversal of flow direction or deflections of up to 60 degrees in plan view. Some lower tier cave streams drain down stratigraphic section, but up the structural bedrock slope. This occurs in a thick well jointed micrite interval where thin weathered beds are exposed at Leonard Steeplehead. Flow direction changes or reversals related to cave tier links are illustrated (Figure 5). Dye tracings are from Reeves and Goodes Cave system was level tube surveyed for comparison of passage ceilings and thalweg gradients to true bedrock dip. This system is enlarged in an anticlinal structure plunging 11 m/km southwest. Visual estimates of structure were made for other caves using apparent dip and rock unit thicknesses. Cave streams at Kirby flow down the bedrock slope in the upper and middle cave tiers with three distinct flow zones typically recognized by thalweg gradients within both tiers. The fissures and swallets zone is relatively steeper. The gradient through the inception horizon zone follows the dip or is slightly steeper. And the karst spring cutoffs zone is oriented down the dip with thalwegs graded between 12 and 21 m/km. Some cascades ranging 5 cm to 3 m in depth factor into the thalweg steepness. Graded flow zones are illustrated in (Figure 6). A detail to be noted is the karst spring cutoff zone at the top of the middle tier occurs in the same rock interval as the fissures and swallets zone for the middle tier. The difference is that the cutoff drains an upper tier cave and the fissures and swallets zone captures surface water as well.

**Spring Stratigraphic Levels**

Spring stratigraphic levels at Kirby are defined for a bedding plane parting on the base of a lithostratigraphic interval bearing evidence of the flow position at the time of speleogenic breakthrough. The definition is based on rock unit sections measured at springs where the level of initial flow is preserved and can be recognized, (Conner, 2011). The spring discharge point at the time of speleogenic breakthrough flows at the position of the spring stratigraphic level. After speleogenic enlargement commences the spring thalweg entrenches below the base of the inception horizon. Spring stratigraphic levels are shown (Figure 4).

**Cave Tiers**

The cave tiers are defined with their bases at a spring stratigraphic level and the upper bound is the next higher spring stratigraphic level in the carbonate sequence. The upper tier caves at Kirby generally have one cave level corresponding to one primary inception horizon, but the middle tier caves typically have multiple piezometric limits corresponding to transitions between multiple inception horizons within the cave tier. The spatial relation between cave streams, flow zones related to structure, and cave tiers is shown (Figure 6).

**Karst Spring Cutoffs**

Karst spring cutoffs are vadose canyons intercepting a cave stream above a spring and diverting it below, by-
Figure 5. Cave Tiers, Springs, Dye Traces, and Structure Contours for Indian Creek Limestone Beds.
passing the spring and linking with the next lower cave tier. This pattern also occurs for springs on the structurally high side of broad sinking stream basins where the intercepting cutoff canyon passes below the basin and drains down the bedrock slope through a fissures and swallets zone before draining into an inception horizon in the next lower cave tier.

Karst spring cutoffs are vadose canyons with a vertical rectangular outline and little upward enlargement above the initiating bedding plane. The cutoff ceiling declines downstream through successively lower bedding planes. The upstream segment of the cutoff thalweg is much steeper than bedrock dip, cascading through thin beds, and becomes closer to horizontal or follows the bedrock slope through the downstream channel. The system of Salamander, Shaft, and Grotto caves illustrates a karst spring cutoff. The cutoff canyons below the tributaries and basins are analogs of canyons below high sandstone capped ridges and both canyon types cross-cut the contact between two cave tiers. A karst spring cutoff example is illustrated in (Figure 6).

Figure 6. Coon Hollow Transects: Upper Cave Tier Enlargement (6A) and later in Middle Tier (6B) with linking through Karst Spring Cutoff.
Discussion

A study combining base level decline and trunk cave streams related to a phase or cycle of speleogenic enlargement involves the initial spring discharge elevations at the time of breakthrough and afterward during the general period of enlargement. The segment of the trunk cave stream passing from some distance upstream from the spring and discharging out of the spring through a short reach of the spring channel is observed at Kirby to become entrenched with a thalweg steeper than the bedrock slope below the base of the inception horizon. The vadose cave stream and its sinking stream basin source above do not decline at the same rate.

Accurate reconstruction of former base level decline surfaces correlating with the enlargement phase for the three cave tiers is difficult. However, the karst spring cutoffs and the spring profiles provide points on the former base level surface that coincide generally with the present topographic surface; and provides useful information for environmental and geotechnical evaluations. Two images show the same transect, one with an approximation of the pre-karst topographic surface, base level when the upper tier cave entered breakthrough phase (Figure 6A). The other shows the present topographic profile and relation to the features of a middle tier cave with a spring cutoff (Figure 6B).

The location for three transects of Kirby Watershed are shown (Figure 5). Transects show the present topographic surface and flowing trunk cave streams for the three cave tiers (Figure 7). Karst spring cutoffs identify points on former base level decline surfaces associated with enlargement in a cave tier.

The relation of an upper tier cave, Truitt’s Cave, and a lower tier cave, Shirley Springs cave is shown relative to the topographic surface. The transect (Figure 7A) shows the present topographic surface which has declined since the Truitt’s Cave stream enlarged in the upper cave tier and has entrenched into the upper Spar Mountain Beds in the cave and at the spring, but no middle tier enlargement related to a middle tier spring has been observed. The same topographic surface extending to the east shows a recharge basin for active cave streams in both the middle and lower cave tiers. There is reasonable evidence that there was westward flowing upper tier cave enlargement potential above the present surface in the east if the westward flowing upper portion of Reeves Cave is considered, shown (Figure 7C). The upper portion of Reeves Cave enlarged flowing to the west, at grade with a former eastern branch of Indian Creek; since captured by the present sinking stream basin flowing to Goodes Branch of Clear Creek. Another example is the upper tier portion of Leonards Spring Cave shown in plan view (Figure 5). Shirley Rockshelter Cave is a relict passage after a spring that became flow abandoned and mud filled in the middle cave tier.

The transect (Figure 7B) shows Saltpeter Cave enlarged in the upper cave tier with a former spring draining into the western branch of Indian Creek. The former spring was diverted by a karst spring cutoff that drained toward a middle tier enlarging cave, Queen Blair. The diversion was contemporary with the early formation of the sinking stream basin in karst valley 10. The active stream in Saltpeter Cave has entrenched with a thalweg into the Spar Mountain Member, but there is no known deeper cave enlargement in the lower tier below the mapped Saltpeter Cave.

The transect (Figure 7C) at the downstream end of Kirby Watershed shows an example of an active middle tier cave stream draining to Richland Creek to the west and an upper and middle cave tier example to the east draining toward Clear Creek. The stream in Reeves Cave upper segment remains active after the spring was abandoned for a karst spring cutoff which appears not to have passed through the main stream in Reeves Cave, but drained through Goodes Cave. Both Reeves and Goodes trunk cave streams are in the middle cave tier.

Conclusion

Identifying spring stratigraphic levels, delineating lithostratigraphic cave tiers, and correlating cave streams with present and former base level decline surfaces is a method of investigating karst flowpaths and their relation to speleogenic enlargement. Flow zones described in this paper and their gradients in addition to the bedrock slope orientation of flowpaths in a cave tier are helpful for interpreting gradients between piezometric limits. The concept is extensible to other areas of the Mississippi Valley Plateaus region for gently dipping and relatively unfaulted limestone strata. Identifying the structure oriented flow zones related to a cave tier improves understanding of flow patterns and changes in an area as base level declines. The changes in flow direction are associated with the headward expansion of a karst plateau and capture of groundwater from an eroding upland at Kirby Watershed and along the Springville Escarpment.

Acknowledgements

The author wishes to acknowledge the long term successes of Keith Dunlap in establishing the Indiana Karst Conservancy, Sam Frushour of the Indiana Geological Survey for efforts supporting the Ard Blenz Nature Conservancy and the Blooming...
Figure 7. Hydrogeologic Transects with Trunk Cave Streams in Cave Tiers and Karst Spring Cutoffs correlating to Base Level Decline Surfaces.
ton Parks Department for establishing the Leonard Springs Nature Park. Their dedicated stewardship to the karst and cave resources in Kirby Watershed and other areas in southern Indiana is greatly appreciated.

References


DRIFTLESS AREA KARST OF NORTHWESTERN ILLINOIS AND ITS EFFECTS ON GROUNDWATER QUALITY

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Abstract
The bedrock aquifer of the Driftless Area of northwestern Illinois is Ordovician Galena Dolomite. Previous work by the authors and others showed that the geology and hydrogeology of this area fall well within the definition of karst. Bedrock in the study area has been shown to be extensively fractured and creviced; karst features in the county are dominated by solution-enlarged crevices from 0.4 inches to 3 feet (1.0 cm to 0.9 m) wide within most road cuts and quarries examined. Other karst features include cover-collapse sinkholes ranging from 3 to 25 feet (0.9 to 7.6 m) in diameter overlying Galena Dolomite, karst springs and crevice caves.

A preliminary evaluation of the groundwater quality in Jo Daviess County in the Driftless Area of northwestern Illinois was conducted to assess the susceptibility of the Galena Dolomite aquifer to surface-borne contaminants. This was done by evaluating available groundwater quality data from published sources and the Illinois State Water Survey Groundwater Quality Database (i.e., wells and springs), and also by sampling 11 private wells across the county and analyzing for inorganic chemistry, dissolved organic carbon, stable isotopes and tritium. We found that groundwater in the study area is of a Ca-Mg-HCO$_3^-$ type as would be expected for an aquifer dominated by dolomite. In parts of the county, the upper part of the carbonate-hosted aquifer contains elevated concentrations of chloride, nitrate and potassium. Likely contamination sources are anthropogenic and include road salt, potash and nitrogen fertilizers, and livestock/human waste. The presence of these contaminants suggests movement of surface-borne contaminants into the aquifer and into wells screened at depths ranging from 65 to 150 feet (20 to 46 m).

Historic data reveal stratification of surface-borne contaminants (greatest concentrations nearest the surface) within the fractured and creviced aquifer to a depth of about 300 feet (91 m). Nitrate-N (NO$_3^-$-N) concentrations in karst springs are typically between 10 and 13 mg/L, but can exceed 30 mg/L. Because the predominant land use in the study area is row-crop agriculture, it is likely that much of the NO$_3^-$-N is derived from N-fertilizer. In the 11 water well samples, NO$_3^-$-N concentrations ranged from < 0.04 (detection limit) to 5.4 mg/L and concentrations were clearly related to tritium. Specifically, NO$_3^-$-N concentrations in groundwater containing less than 3 TU were below detection (0.04 mg/L), and above 3 TU, NO$_3^-$-N and tritium were positively correlated. This relationship suggests a nonpoint source of N and denitrification within the aquifer. Chloride (Cl$^-$) concentrations in karst springs were between 15 and 25 mg/L above background concentrations (1 to 15 mg/L). Water wells samples had lower Cl$^-$ concentrations with 7 of 11 wells below background (ca. 15 mg/L), although the concentration in the shallowest well was 110 mg/L and was probably derived from road salt. Overall, the groundwater quality of the Galena Dolomite aquifer in Jo Daviess County is what would be expected in an open, dolomite-dominated karst aquifer.

Introduction
Groundwater in karst regions of the Midwestern US is typically Ca-HCO$_3^-$ to Ca-Mg-HCO$_3^-$ type water that can have relatively high levels of surface-borne contaminants, especially at shallow depths. Typical contaminants in wells and springs include sodium (Na), potassium (K), chloride (Cl$^-$), nitrate-nitrogen (NO$_3^-$-N) and enteric bacteria (e.g., Panno et al. 1996; Hackley et al. 2007).

The Driftless Area of northwestern Illinois was mapped in the mid-1990s as karst by Weibel and Panno (1997) and Panno et al. (1997). Domestic wells in Jo Daviess County get their water primarily from the Galena Dolomite at depths less than 250 feet (76 m) (Frankie and Nelson 2002). While the degree of karstification has been well established (Panno et al. 2015a,b,c; Luman and Panno 2015), its effects on groundwater quality in the area are uncertain, although Panno and Luman (2008) explored this to a limited extent. They found that Cl$^-$ and NO$_3^-$-N concentrations in private and public wells (most are cased more than 100 feet below the surface) in the county were as high as 55 and 31 mg/L, respectively. These concentrations are well above the
upper background threshold concentrations (1 to 15 mg/L for Cl\textsuperscript{−} and 0 to 2.5 mg/L for NO\textsubscript{3}-N) for freshwater aquifers in the northern two-thirds of Illinois (Panno et al. 2006a, b). Potential sources include both urban and rural contaminants such as road salt (Cl\textsuperscript{−}), nitrogen fertilizer (NO\textsubscript{3}-N), and animal waste and septic effluent (both Cl\textsuperscript{−} and NO\textsubscript{3}-N). Stratification of concentrations of both NO\textsubscript{3}-N and Cl\textsuperscript{−} was observed.

In order to determine the susceptibility of the shallow karst aquifer to groundwater contamination in the study area, we gathered water quality data from historic sources (Illinois State Water Survey (ISWS) Groundwater Quality Database), a recent study of six karst springs in northeastern Jo Daviess County (Maas 2010), and from sampling 11 domestic shallow water wells across the county. The objectives of this investigation were to establish a preliminary range and background for major cations and anions, and conduct a preliminary evaluation of the existing water quality in the Driftless Area. We also analyzed well water samples for tritium to get a sense of the residence time of groundwater in the study area and compare tritium to selected ions that may be introduced by surface-borne contamination.

**Geology and Hydrogeology**  
Jo Daviess County lies within the Driftless Area of northwestern Illinois; the county, for the most part, lacks glacial drift that covers the bedrock of most of the upper Midwestern U.S. (Figure 1). However, glacial till is exposed along U.S. Route 20 between Galena and East Dubuque (Willman and Frye 1970; D. Kolata, personal communication, 2014), as well as a small area of glacial till in an isolated portion of the far eastern part of the county. Bedrock consists of Middle-Ordovician age (443 – 490 Ma) carbonate rocks of the Galena-Platteville Group, thin remnants of the Ordovician-age Maquoketa Shale, and Silurian dolomite (412 – 443 Ma) whose resistant rock caps the mounds and highlands of the county (Figure 2).

Tectonic compression and extension occurred in this area during and following the formation of the Wisconsin Arch that began in Cambrian time (490 – 543 Ma) and continued to be active in late Silurian or Devonian time (354 – 417 Ma) (Nelson 1995). As a result of compression and extension, bedrock along the Wisconsin Arch has a well-developed vertical joint system. Heyl et al. (1959) stated that “All the rock formation in the [Upper Mississippi Valley mining] district [most of Jo Daviess County] contain well-developed vertical and inclined joints (Figure 3). The vertical joints are traceable for as much as 2 miles (3.2 km) horizontally, and for as much as 300 feet (91 m) vertically. Joints are especially well developed in the Galena Dolomite.” Trowbridge and Shaw (1916) stated that crevices within the Galena Dolomite “…are frequently encountered in wells, and drilling tools are sometimes lost in them.”

In the mid-1990s, the Illinois State Geological Survey (ISGS) identified Jo Daviess County as karst (Weibel and Panno 1997; Panno et al. 1997) (Figure 1). Subsequent work by McGarry and Riggs (2000) identified most of Jo Daviess County as having “…a very high aquifer sensitivity because fractured dolomite bedrock aquifers lie beneath the glacial drift or loess. Areas where dolomite bedrock is exposed are most sensitive.” The main karst water-bearing formation in the county is the Galena-Platteville Group. Silurian dolomite is also karstified, but is rarely used as an aquifer in the county. Ekberg (2008) subdivided the secondary porosity of the Galena-Platteville Group into matrix, fracture and conduit porosity. These subdivisions are supported by spring hy-
drographs and drawdown curves from aquifer tests that support a triple porosity aquifer. The fracture porosity through which groundwater flows consists of northeast- and northwest-trending vertical fractures (consistent with Heyl et al. 1959), and bedding planes (Ekberg 2008).

Panno and Luman (2008) examined sinkholes and the abundant secondary porosity (crevices) exposed along road cuts and in quarries in eastern Jo Daviess County and concluded that the Galena Dolomite constitutes a karst aquifer. While cover-collapse sinkholes are present over the Galena Dolomite, they are typically small due...
to the thin soils of the area and are easily buried (albeit temporarily) in croplands (Figure 4). However, because of their presence, the county falls into a “medium” to “high” category of aquifer vulnerability as outlined by Lindsey et al. (2010).

METHODS

Groundwater Chemistry

Groundwater chemistry data from private wells and karst springs in eastern Jo Daviess County were available through the ISWS Groundwater Quality Database described in Panno and Luman (2008), and from Maas (2010), respectively. In addition, 11 domestic shallow wells were sampled across the county (one in September, 2013, the other ten in June 2014) and analyzed for dissolved cations, anions, dissolved Kjeldahl N (DKN), ammonia, dissolved organic carbon (DOC), D/H and 
\(^{18}\)O/\(^{16}\)O isotopic ratios, and bacterial indicators. Groundwater samples from the 10 wells sampled in June 2014 were analyzed for tritium.

Wells were purged for at least 15 minutes until field parameters (temperature, pH, dissolved oxygen (DO), specific conductance (SpC) and oxidation-reduction potential (ORP) stabilized. The field parameters were measured using a multisonde (Hydrolab®, Loveland, CO). Samples designated for chemical analyses were filtered through 0.45-µm membranes, placed in polyethylene or glass bottles and stored at 4°C prior to analysis. All samples were transported in ice-filled coolers to the laboratory, and kept refrigerated until analysis. Analyses for inorganic chemicals were conducted in accordance with standard methods (APHA, AWWA, WEF 1999) at the ISWS using standard methods.

Sterile techniques were used for samples collected for bacterial indicators (total coliform and Eschericia coli). Outside spigots were flame sterilized then water was collected in sterilized bottles and stored in ice-filled coolers before transporting to the analytical laboratory the same day. Bacterial indicators were determined using the Colilert method (IDEXX 2013) at the City of Dubuque Laboratory, Dubuque, Iowa.

Tritium was analyzed using electrolytic enrichment (Os- tlund and Dorsey 1977) and liquid scintillation counting as described in Hackley et al. (2007). Results are reported in tritium units.

Results and Discussion

Other Hydrologic Features

Springs are a common feature throughout Jo Daviess County and the locations of some have been mapped by Reed (2008) and Maas (2010). The only available data on the chemical composition of springs in the county are from Maas (2010) for six springs in northeastern Jo Daviess County within the Warren Quadrangle. The springs lay along prominent lineaments identified in this investigation and are consistent with discharge of groundwater along bedrock crevices where the overburden thins near stream valleys. Groundwater, under hydrostatic pressure, would be able to breach land surface in low-lying areas with relatively thin overburden (usually near streams). Bedrock springs typically appear to be large circular to elliptical depressions with small sections breached, providing openings through which the spring water discharges to a nearby stream.

Chemical Composition of Groundwater

Because spring water in karst regions is an amalgam of groundwater from various sources, the chemical composition of spring water is useful in characterizing background concentration ranges of constituents and trends for contaminants (Table 1). The overall chemical composition of groundwater and relationships between and among selected cations and anions were used to identify the source(s) of contaminants. Background concentration ranges of selected ions are useful for comparing natural compositions of surface water and groundwater with waters that have been affected by anthropogenic and/or natural contamination. When making such comparisons, one must be aware that, for example, concentrations of Na, Cl\(^-\) and NO\(_3\)-N that are somewhat elevated above background do not constitute water that is harmful to humans or to natural flora and fauna of an area. It does, however, indicate that surface-borne contaminants from land-use activities have entered groundwater and will ultimately discharge to surface waters. Further, it has been shown that elevated concentrations of Na and Cl\(^-\) can be deleterious to vegetation (e.g., Panno et al. 1999), aquatic organisms (e.g., Kelly et al. 2012), can impart a salty taste to drinking water at Cl\(^-\) concentrations exceeding 250 mg/L, and elevated Na concentrations in drinking water may be a problem for people with high blood pressure (USEPA 2014). Nitrate-N concentrations greater than 10 mg/L in drinking water have been shown to cause methemoglobinemia (blue-baby syndrome) and may be linked to stomach cancer (O’Riordan and Bentham 1993). For the purposes of this investigation, we considered concentrations exceeding the upper end of background as anthropogenic tracers that may be used to investigate aquifer recharge areas, recharge rates and groundwater movement through the underlying karst aquifer.

Groundwater in Jo Daviess County is a calcium-magnesium-bicarbonate (Ca-Mg-HCO\(_3\)^-) type groundwater with elevated concentrations of Cl\(^-\) and NO\(_3\)-N in some
areas (Panno and Luman 2008). The background concentration range for Cl\textsuperscript{-} in shallow groundwater in northern and central Illinois is between 1 and 15 mg/L (Panno et al. 2006a). The range for background concentration of NO\textsubscript{3}\textsuperscript{-}-N in Illinois is between 0 and 2.5 mg/L (Panno et al. 2006b; Hwang et al. 2014). Based on chemical compositions of groundwater from wells and springs, the distribution of these ions and relatively high DO concentrations (4.7 to 8.7 mg/L) in the underlying aquifer is indicative of an open, oxygenated, unconfined karst system. Potential sources of Cl\textsuperscript{-} and NO\textsubscript{3}\textsuperscript{-}-N include road salt, human and animal waste, and fertilizers. Water resources in open aquifer systems such as this are especially vulnerable to surface-borne contaminants. There is little or no attenuation of contaminants discharged into sinkholes, macropores and fissures; consequently, wells, springs, and streams down-gradient of contamination sources can show effects within a few days or even hours (Green et al. 2006). The convergent nature of flow in karst aquifers may result in contaminants becoming concentrated in conduits (Field 1993).

Maas (2010) sampled six springs in the eastern part of Jo Daviess County, and observed that water from five of the springs’ discharges from open, oxygenated systems typically containing elevated levels of Cl\textsuperscript{-} and NO\textsubscript{3}\textsuperscript{-}-N. Nitrate-N concentrations were commonly greater than 10 mg/L and as high as 30 mg/L. Spring water from all six springs was undersaturated with respect to calcite and dolomite which suggests that the springs are either dominantly shallow groundwater or are a mixture of deep and shallow groundwater. The lack of saturation with respect to calcite and dolomite indicates that karstification of the Galena Dolomite is an ongoing process in this area. The elevated concentrations of NO\textsubscript{3}\textsuperscript{-}-N found in all but one of the springs are similar to those of tile drain waters of Illinois. Tile drains have been described by Schilling and Helmers (2008b) as analogous to karst drainage basins with regard to nutrient losses in an agricultural watershed. The elevated nutrient concentrations in the springs suggest that the springs studied by Maas (2010) may be affected by recharge water containing relatively high concentrations of surface-borne contaminants. The relative depths of the groundwater flow system feeding the springs may be estimated by the changes in temperature of the spring water with time. Specifically, the shallowest flow systems would be most affected by seasonal changes that would be manifested in the fluctuation of spring water temperatures. All but one spring were affected by seasonal changes.

Because of the nature of groundwater flow in karst aquifers, groundwater pathways may be discrete conduits/crevices and/or bedding planes fed by numerous, smaller crevices within carbonate bedrock. Because springs are discharge points for groundwater, they may be fed by flow paths of various ages. Some of the input may have a shallow component containing surface-borne contaminants from a variety of land uses, whereas other inputs may originate from deeper, usually less contaminated, sources. The percentages of each component source can vary depending on the groundwater flow paths to the springs and with time/season.

Background concentrations of major ions and contaminants are presented in Table 1. These concentrations are from Panno et al. (2006a, b) and groundwater samples from the karst regions of southwestern Illinois, and also inferred from the water quality data from the spring

<table>
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<th>Ion/Parameter</th>
<th>Range</th>
<th>Background Threshold</th>
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<tbody>
<tr>
<td>pH</td>
<td>6.35 – 8.90</td>
<td>ND</td>
</tr>
<tr>
<td>Specific Conductance</td>
<td>520 – 1100</td>
<td>700**</td>
</tr>
<tr>
<td>Sodium (Na)</td>
<td>5.70 – 13.4</td>
<td>10 mg/L**</td>
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<tr>
<td>Potassium (K)</td>
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<td>0.6 mg/L**</td>
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<tr>
<td>Calcium (Ca)</td>
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<td>Magnesium (Mg)</td>
<td>35.8 – 56.6</td>
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</tr>
<tr>
<td>Silicon (Si)</td>
<td>4.5 – 8.0</td>
<td>ND</td>
</tr>
<tr>
<td>Bicarbonate (HCO\textsubscript{3})</td>
<td>175 – 400</td>
<td>ND</td>
</tr>
<tr>
<td>Chloride (Cl)</td>
<td>0.13 – 26.7</td>
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<tr>
<td>Fluoride (F)</td>
<td>0.081 – 0.38</td>
<td>0.4 mg/L**</td>
</tr>
<tr>
<td>Sulfate (SO\textsubscript{4}\textsuperscript{2–})</td>
<td>8.32 – 31.1</td>
<td>35 mg/L**</td>
</tr>
<tr>
<td>Nitrate-nitrogen (NO\textsubscript{3}\textsuperscript{-}-N)</td>
<td>&lt;0.30 – 30.1</td>
<td>2.5 mg/L*</td>
</tr>
</tbody>
</table>

Table 1. The range of concentrations and estimated background threshold values for parameters and ions determined from spring-water samples collected in eastern Jo Daviess County by Maas (2010). Threshold values are the upper bounds of background concentrations and were estimated based on previous work by Panno et al. (2006a, b)*, unpublished data from the sinkhole plain of southwestern Illinois (Panno, ISGS), and spring data from Hicks Spring** (Maas 2010) located in eastern Jo Daviess County. ND = Not determined.
which appears to be the least affected by recharge events and surface-borne contaminants (Maas 2010) (Table 1). There are distinct differences among the six springs based on concentrations of indicators of surface-borne contamination which include K, Na, Cl⁻, and NO₃⁻-N. Maas (2010) came to slightly different conclusions using various non-site specific references (e.g., Hem 1986) that relied on ion concentrations in soils and carbonate rock to determine whether ion concentrations of the springs were above or below background threshold values. He concluded that Na and sulfate (SO₄²⁻), as well as Cl⁻ and NO₃⁻-N were anomalous and were derived from surface sources. Because the geology of Jo Daviess County is not solely carbonate rock, other strata (e.g., shale) could have contributed Na and SO₄²⁻ to the groundwater. Potential contaminants in the area include fertilizers and road salt (NaCl). Fertilizers used in the area include urea, 28% solution N, anhydrous ammonia, diammonium phosphate, potash, and, on a more local basis, hog and dairy manure. The McPhilips Spring is in the vicinity of a dairy where manure is applied to the field (Mr. Jim Frances, Warren, IL, personal communications, June 2013). Because households in this area are on private septic systems, waste from septic effluent is also a possible contaminant (Panno et al. 2007).

Preliminary work by Panno and Luman (2008) on available water quality data from public water wells from Jo Daviess County and Stephenson Counties showed that Cl⁻ concentrations ranged from less than 1 to 55 mg/L and NO₃⁻-N concentrations from less than 0.1 to 31 mg/L.

Denitrification probably accounts for the very low concentrations to absence of NO₃⁻-N at depths greater than 400 feet. Background concentrations may have a lower upper threshold than the estimated 15 mg/L Cl⁻ and the 2.5 mg/L NO₃⁻-N (Table 1). Additional estimates for background concentrations of Cl⁻ and NO₃⁻-N (based on municipal well data) are slightly lower, about 1 to 10 mg/L and 0 to 2 mg/L, respectively. The deeper groundwater in the area from wells that intersect the St. Peter Sandstone have Cl⁻ concentrations between < 1 and 3 mg/L in Jo Daviess County and about 1 mg/L in the Galena area. Chloride concentrations in aquifers at that depth are anomalously low with the lowest concentration (0.6 mg/L) being only slightly greater than the concentration of Cl⁻ in rainwater (0.1 to 0.3 mg/L) and slightly less than in soil water (0.7 to 1.7 mg/L) (Panno et al. 2006a). Processes in the unsaturated zone (e.g., evapotranspiration) and rock water interactions typically increase the Cl⁻ concentration of groundwater up to 15 mg/L. Recharge of glacial melt water would be capable of leaching soils and bedrock aquifer flow paths to the point where they would impart little Cl⁻ to the groundwater via rock-water interactions. This effect is apparent in part of the Mahomet Aquifer in east central Illinois (Hackley et al. 2010).

The pH values of spring water sampled by Maas (2010) were always below 7.0 which are consistent with the spring water being undersaturated with respect to calcite and dolomite. Water in equilibrium with calcite or dolomite would have a pH value of about 7.5 as noted earlier. These lower pH values indicate that groundwater from the springs is not in equilibrium with the carbonate rock supporting the interpretation of an open system.

Specific conductance (SpC) is a measurement of the electrical conductance of the water that can be easily taken in the field. While SpC is a very general measurement, it does yield an indication of the presence of contaminants to groundwater of surface water. The addition of contaminants (e.g., road salt) can increase the SpC proportionally. The approximate background values in southwestern Illinois and northwestern Illinois groundwater are 650 and 700 μS/cm, respectively based on available data from Maas (2010) and Panno (ISGS, unpublished data).

Potassium and Cl⁻ concentrations in spring water suggest that at least two of the springs may have been contaminated with potash fertilizer, manure, and/or human waste. Similar effects were seen in tile drain-fed stream water samples from the same area during the same period (Maas 2010). Groundwater samples from the other springs all fall below the K background threshold of 0.6 mg/L (Table 1).

Nitrate-N concentrations for most of the springs were greater than those of shallow groundwater of Illinois and far exceeded the background value of 2.5 mg/L (Table 1). Nitrate-N concentrations ranged between 2.9 and 30 mg/L (median = 12.3 mg/L) for five of the six springs, with all but one sample exceeding 10 mg/L. These concentrations exceeded those found in tile drains of central Illinois measured by the authors (e.g., ranging from 0.51 to 23.1 mg/L, median = 11.3 mg/L (Panno et al. 2005)) and in tile-drain fed surface streams in Jo Daviess County (i.e., ranging from 4.19 to 14.6 mg/L, median = 9.31 mg/L (Maas 2010)). The elevated levels are most likely due to the application of N-fertilizer and manure to agricultural fields in the area. Nitrate-N concentrations in one of the springs were very low, equivalent to what Panno et al. (2006b) determined to be indicative of background conditions (< 0.4 mg/L). Whereas it is possible that the low NO₃⁻-N concentration in this spring is the result of denitrification, the background-level concentration of Cl⁻ and K suggest that it is more likely that the spring represents background concentrations for shallow
groundwater in the area. Hicks Spring appears to be the least affected by anthropogenic sources and recharge events among the six springs sampled by Maas (2010), and its flow path may be deeper than the other springs.

All springs are in close proximity to land supporting row-crop agriculture and one is in close proximity to a livestock operation. However, the actual recharge areas of the springs are not known. Consequently, it is not possible at this time to use land use as an indicator of the sources of surface-borne contaminants in any more than a general way. Chloride and NO$_3$-N concentrations in Hicks Spring are well below their background thresholds. Because background concentrations for Cl$^-$ and NO$_3$-N are exceeded in five of the six springs and because the ions covary, the data suggest that these springs are being fed by an aquifer with a steady input of these ions. The data also suggest that the ions are entering the aquifer together perhaps as manure and/or septic effluent. A plot of Na vs Cl$^-$ for all spring data shows no discernible covariance, indicating that road salt (NaCl) is not the likely source of either ion in the springs. However, there are not enough chemical data to determine the source(s) of these ions at this time.

Overall, the degree of contamination of the spring water with regard to NO$_3$-N is greater than typically found in shallow groundwater of Illinois. This, plus the fact that concentrations of K and Cl$^-$ exceed background concentrations in all but one spring indicates that N-based fertilizers and/or manure/human wastes are likely entering the groundwater systems. The high DO concentrations in the spring samples (Maas 2010) suggest rapid movement of water into the subsurface which would limit attenuation within the soil zone. In situations where water recharges more slowly through the soil zone, oxygen is consumed as organic matter is reduced, resulting in anoxic conditions that promote denitrification which would decrease NO$_3$-N concentrations. The thin soils and the presence of macropores and sinkholes in this area appear to promote rapid recharge to bedrock aquifers, a common feature of karst regions.

**Shallow Wells within the Galena Dolomite**

Our sampling is the first time wells in the Driftless Area of Illinois have been sampled on a systematic basis. The wells ranged from 70 feet to 160 feet in depth. As with the spring water samples, groundwater from all wells is a Ca-Mg-HCO$_3$ type water with pH values ranging from 6.56 to 6.73. Tritium in groundwater samples ranged from 0.60 to 5.45 tritium units (TU) (Figure 5). The concentration of tritium in today’s rainfall (ca. 2014) is about 5.5 TU (Fanta, ISGS, personal communications, September 2014). The presence of tritium in the well water samples indicates that there is some component of modern recharge.

The Ca-Mg-HCO$_3$ type water was expected because each well intersected the fractured and creviced Galena Dolomite aquifer. Groundwater in equilibrium with dolomite should have a pH value of about 7.5 (e.g., Parkhurst and Plummer 1993). The Mg/Ca ratio of the samples ranged from 0.471 to 0.623 (median = 0.545); pure dolomite dissolved by fresh water would have a Mg/Ca ratio of 0.61. Thus, most of the samples (all but one between 0.516 and 0.623) are consistent with dissolution of dolomite. The sample with the lowest ratio of 0.417 may be affected by a thin bed of limestone. It is clear that dolomite has not been dissolved to equilibrium within the groundwater probably due to the influx of and mixing with recent recharge.

Evidence of modern recharge is also present as surface-borne contaminants such as Na, Cl$^-$ and NO$_3$-N. Tritium does not decrease with depth, and it does not co-vary with Na or Cl. This lack of correlation may be a reflection of the highly fractured and creviced nature of the Galena Dolomite aquifer and the vagaries of well drilling. Available data from wells screened within the deeper Plattville Formation and St. Peter Sandstone show anomalously low concentrations of Cl$^-$ (about 1 mg/L). This suggests that the upper part of the Galena Dolomite aquifer is more affected by surface-borne contaminants than deeper units, as would be expected.

Nitrate-N concentrations range from below detection (< 0.04 mg/L) to 5.42 mg/L; only two of the wells had NO$_3$-N concentrations that exceed the background threshold concentration of 2.5 mg/L. Nitrate concentrations in the wells samples correlate well with tritium concentrations; NO$_3$-N concentrations were below detection (<0.04 mg/L) for TU concentrations less than 3 TU (Figure 5).

**Figure 5.** Nitrate concentrations decrease with time within the karst aquifer, probably due to denitrification (e.g., Panno et al. 2001).
This shows a strong correlation between NO$_3^-$-N and groundwater age may be the result of denitrification and/or mixing with NO$_3^-$-N –free groundwater. Fluoride (F$^-$) appears to be negatively correlated with tritium (Figure 6) suggesting that F$^-$ is derived from rock-water interaction and might serve as a surrogate/proxy for tritium in this area.

Calcium and sulfate concentrations co-vary within the shallow aquifer (Figure 7). Given the abundance of pyrite and other sulfides within the Galena Dolomite, the oxidation of pyrite is probably the main source of SO$_4^{2-}$.

Delta δD and δ$^{18}$O data from precipitation in Jo Daviess County should be similar to that of precipitation in the Chicago area. Precipitation data collected in Chicago from 1960 to 1979 (IAEA 2014) provide a local meteoric water line similar to that of the Global Meteoric Water Line. The stable isotope data for the groundwater samples collected in Jo Daviess County fall along this line and indicate that little has occurred to the precipitation/recharge (e.g., evaporation) prior to entering the Galena Dolomite aquifer.

Combining these data with those of the ISWS database yielded a representation of the vertical distribution of selected ions. The vertical distribution of surface borne contaminants such as Cl$^-$ and NO$_3^-$N is distinctive (Figures 8 and 9). Chloride concentrations are due, in a large part, to road salt contamination. The highest Cl concentrations...
Concentrations are generally found in wells less than 240 feet deep; below around 380 feet (116 m) deep, Cl- concentrations are always below background levels (Figure 8). The vertical distribution of surface-borne contaminants is consistent with that of a fractured and creviced karst aquifer. Elevated NO$_3$-N concentrations are probably due to the application of N-fertilizers in the area given that land use in Jo Daviess County is dominated by row-crop agriculture. The relatively large NO$_3$-N concentrations near the surface are typical of agricultural areas, but as with the Cl- concentrations, the sporadic nature of the profile is consistent with a fractured and creviced karst aquifer.

Overall, the chemical composition of relatively shallow groundwater within the Galena Dolomite aquifer is consistent with that of the springs and reflects an open fractured karst aquifer whose mineralogy is dominated by dolomite. In particular, elevated NO$_3$-N concentrations at depth are indicative of an open, oxygenated system with discrete and focused flow paths.

### Conceptual Hydrologic Model

A preliminary conceptual model for the hydrogeology of the Galena Dolomite, based on previous studies in counties east of Jo Daviess County suggests that the northeast-southwest and northwest-southeast trending fractures/crevices and bedding planes constitute the greatest porosity for the Galena Dolomite aquifer (Ekberg 2008). Panno et al. (2015c) and Luman and Panno (2015) showed that these fracture/crevice systems are oriented more north-south and east-west in Jo Daviess County (Figure 10).

Based on this investigation and Panno et al. (2015b), we have documented that the Galena Dolomite has its greatest porosity within at least the upper 15 to 25 feet (4.6 to 7.6 m) of the dolomite with fractures and solution-
enlarged crevices ranging from < 0.4 in to 3 feet (< 1 cm to 1 m) or wider. This is in spite of fine-grained sediment found in the upper parts of crevices greater than about 0.4 inches (1 cm) in width; the depth of this material and its effects on groundwater flow is currently under investigation. In general, the crevices provide a network of pathways through which infiltrating surface water and groundwater can flow rapidly. Below 25 feet (9.6 m), many crevices become narrower and range from less than 0.25 to greater than 1 inch (0.6 to 2.5 cm); east-west trending crevices tend to retain their widths of 3 feet or more with depth, thereby providing large conduits for the karst aquifer. Bedding planes may also provide pathways for groundwater movement. The effect of depth on the porosity and permeability associated with shear zones is currently under investigation. However, elevated concentrations of NO\textsubscript{3}-N and Cl\textsuperscript{-} suggest the system is susceptible to surface-borne contamination to depths greater than 300 feet.

It is reasonable to assume that relatively rapid recharge to the karst aquifer occurs throughout the county, but probably less so where Maquoketa Shale is present. In areas where Maquoketa Shale constitutes the bedrock surface, and where drain tiles are used to lower the water table, relatively greater amount of recharge may discharge to streams before entering the karst aquifer. Sinkholes and macropores are present throughout the county and are locations of focused recharge. Sinkholes are not commonly seen due either to cultivation, which tends to obscure all but the very largest ones, or perhaps their natural scarcity. During dry periods, macropores (desiccation cracks) are ubiquitous and form easily due to the thinness of the soil and the depth of the water table (below the soil-rock interface) (Panno et al. 2013).

References


SEEPS AND SPRINGS AT A PLATTEVILLE “OBSERVATORY” ON THE RIVER BLUFFS

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Abstract
Residential building construction along the Mississippi River bluffs in the 1970s created a unique enclosed outcrop of the Late Ordovician Platteville Formation at Lilydale, Minnesota. This outcrop was examined in early 2013 after a newly-formed spring flooded an elevator shaft the previous year, drawing attention to the foundation conditions.

The Lexington Riverside property is a six story condominium complex constructed within the top of the bluff. A two-level underground parking garage was built into the bluff. Bedrock was mechanically excavated to accommodate the construction of the building, creating an unweathered rock surface. The space between the structure and the excavated rock face, running for 150 meters, was roofed over, and is used as a utility space. At least three dominantly carbonate members of the Platteville Formation are visible: Mifflin, Hidden Falls, and Magnolia, in ascending order. The foundation of the structure was constructed on the lowermost Platteville limestone and Glenwood shale and is tile-drained to the nearby river gorge.

Most of the seeps and springs on the property, both inside and on the grounds, belong to the three Platteville spring-lines identified for the Twin Cities Metropolitan (TCM) area by Brick (1997). Groundwater emanates from both vertical joints and horizontal bedding plane partings within the Platteville Limestone and at the Platteville–Glenwood Shale contact. Overall, the hydrostratigraphic attributes of this site are consistent with how the Platteville has been recently characterized in the TCM area in a fractured bluff edge setting (Anderson et al., 2011).

The enclosed outcrop features many seep- and spring-related mineral deposits. Most notable were the iron-stained flowstone and microgours near the seeps and springs along fractures in the limestone, and calcite rafts on the surfaces of the pools. At some damp locations a fungal ecosystem has developed. Gypsum beards have grown in dry portions of the cavern.

This man-made cavern, and others nearby, present unique opportunities to research groundwater flow in fractured bedrock settings. Studying the spring locations relative to joints and bedding, changes in spring flow rate over time, and mineral deposition rates, are possible in this accessible location without the complication of surface water inputs or instrumental interference from the general public.

References
HYDROGEOLOGICAL DYNAMIC VARIABILITY IN THE LOMME KARST SYSTEM (BELGIUM) AS EVIDENCED BY TRACER TESTS RESULTS (KARAG PROJECT)

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Abstract
Paleozoic carbonate aquifers represent major groundwater resources in Belgium. Karstification processes affect most of them and Belgium counts many hydrologically active karst networks. Given the intrinsic vulnerability of such geology, comprehensive studies are required in order to improve their protection and management.

The KARAG project (Karst Aquifer ReseArch by Geophysic, 2013-2017) aims to identify the specific dynamic of karst aquifers by using geophysical and hydrogeological tools. This research is funded by the Belgium National Fund for Scientific Research (FNRS) and conducted by the University of Namur, University of Mons and the Royal Observatory of Belgium.

The LKS – Lomme Karst System (Rochefort, southern Belgium) was chosen as it is a major Belgian karst system (10 km long) in the Givetian carbonate aquifer (Middle Devonian). The system is formed by two parallel components: the surface system (the Lomme River) and a complex underground system (multiple sinkholes with one main resurgence). Based on this layout, it is possible to study the aquifer dynamic and its relationship with the surface river.

A high resolution monitoring network has been in place since July 2013 in order to follow the system dynamic during several hydrogeological cycles.

Multi-tracing experiments with different injections and monitoring points highlight the complexity of underground flow dynamics. Investigations enlightened the connectivity between monitoring points and how dependent of the hydrological conditions were these connections. The breakthrough curves analysis allows to characterize the hydrodynamic behavior of the underground flows within the aquifer.

Modeling will be necessary to link hydrological and tracer tests data in order to build a detailed conceptual model for this karst system. This model will also be used to interpret geophysical data (ERT, gravimetry) collected in order to study the unsaturated and epikarst zones.

Introduction
Karst areas represent an important part of the Belgian territory, especially in the south (Wallonia) where Paleozoic limestones (Carboniferous and Devonian) have extensive cave development. A report established in 2011 stated that groundwater coming from those limestones represents 53% of the total water volume extracted from Walloon aquifers (200 million cubic meters; SPW-DGO3, 2014). In addition, Belgium is a densely populated country with heavy agricultural and industrial development. Many human activities represent a threat for the karst media especially in terms of water quality.

There are several active karst systems in Wallonia and only a few of them are well understood. The recharge areas, the underground connections and the hydrological behavior regarding hydrological conditions are the main questions to be answered in order to protect those systems. It is clear that comprehensive studies about
The behavior of karstic systems will help to improve the management of their water resources.

The KARAG project, launched in 2013, aims to understand the characteristics of epikarst and karst aquifers with a multidisciplinary approach: hydrogeology and geophysics (ERT and gravimetry). This 4-year study is funded by the National Fund for Scientific Research in Belgium (FNRS) and conducted by the University of Namur, University of Mons and the Royal Observatory of Belgium. The Lomme Karst System (LKS) was chosen as it is one of the main karst systems in Belgium. One of the key work packages concerns the hydrogeological behavior of the karst system that may be synthesized in three main questions:

- What is the hydrogeological organization of the karst system (flow connections)?
- What is the dynamic of those connections during variable hydrological conditions?
- What is the relationship between the subterranean karst system and the surface Lomme River?

This short paper introduces this particular research topic presenting our methodologies and first results.

**The Lomme Karst System (LKS)**

The LKS is located in south Belgium, near the city of Rochefort, within the Lomme River valley (Figure 1). It is one of the longest active karst systems in Belgium with almost 10 km of a complex underground network with multiple sinkholes and one main resurgence (Figures 3 and 4). The different cave networks are formed within the Givetian carbonated aquifer (Middle Devonian), which is one of the most karstified aquifers in Belgium. At the surface, the Lomme River is flowing on the Givetian limestones, almost parallel to the underground network(s).

The first item of interest in this system is the double (surface and subterranean) flow system which allows a comparison of the fate of water in both.

The general organization of the underground flow paths seems to be simple because all the water entering the karst system (through sinkholes or dolines) will flow to the Eprave resurgence (the biggest in Belgium with an average discharge of 800 liters/sec), the only final point of this karst network.

However, the dynamic of underground flows is very complicated because every sinkhole feeds different subterranean rivers. The relationships between these rivers is not yet fully understood.

The large number of caves with hydrogeological features (15) provides a great opportunity to understand the functioning of this system. Moreover, recent speleological discoveries bring new information about underground rivers in this karst system.

**Methodology**

In order to collect enough data to feed our future hydrogeological model, 20 water probes were installed throughout the LKS and in the surrounding karst aquifer and surface rivers (Figures 3 and 4). A 15-minute time interval will provide high-resolution data (water level, temperature and electrical conductivity) to understand
the system dynamics during various hydrogeological cycles. Treatment and analysis of piezometric data will be done in a future step of the project.

The main part of the hydrogeological study is devoted to dye tracing. In 2014, five GGUN-FL30 fluorometers (Schnegg, 2002) were used to conduct tracer tests in varying hydrologic conditions. Uranine (C_{20}H_{10}Na_{2}O_{5}) and sulforhodamine B (C_{27}H_{30}N_{2}O_{7}S_{2}) were used for injections performed at various injection locations. Two dye tracing campaigns were conducted: in February (high water conditions) and in September (low water conditions).

Different sites were chosen for a total of 9 tracer injections during the two campaigns (Figures 3 and 4):
- The Kerwée sinkhole, on the Wamme river (point 1);
- The Mortier sinkhole, on the Lomme river (point 2);
- The Pré-au-Tonneau sinkhole, on the Lomme river (point 3);

Figure 3. Geological map of the Lomme karst System with the main Givetian limestone formations, speleological networks, water probes and main interest karst sites for the hydrogeological study and tracer test experiments (geological background : Barchy et al. submitted; Blockmans and Dumoulin, submitted).

Figure 4. Zoom of Figure 3 showing the caves and interest points at the south of the city of Rochefort; Yellow points represent water probe locations, numbers refer to the Figure 3 list.
Figure 5. Breakthrough curves for the 9 tracer tests carried out in the LKS in February (High piezometric conditions) and September 2014 (Low piezometric conditions). Each graph represents one experiment showing the dye concentration at the downstream monitoring stations (one color for each station). The yellow star points the injection time. The dashed black line represents the surface Lomme River outflow in Rochefort and is representative to the general hydrological conditions.
The Lorette cave river (point 4; Figure 2); The Fosse aux Ours cave river (point 7); The Trouvée sinkhole (point 8).

Dye sampling was done at five points with field fluorometers:
- The river in the Lorette cave (point 4);
- The north rivers of Thiers des Falizes:
  - Hôtel cave river (point 5);
  - Muret cave river (point 6);
- The south river of Thiers des Falizes:
  - Fosse aux Ours river (point 7);
- The Eprave resurgence (point 10).

Tracer Test Results
The breakthrough curves for the 9 tracer tests experiments are presented in Figure 5 that also gives reference to the injection location and time. The outflow of the Lomme River is also given as it is representative to the hydrological conditions of the LKS.

Based on the tracer test results, the karst flow functioning within the LKS has been defined and appears to be much more complicated than expected. Indeed, the flow connections between underground rivers seem to vary with the hydrological conditions. Two cases were identified from our tracing results: one for high piezometric conditions, another for low piezometric conditions.

Figures 6 and 7 are two schematic maps illustrating the main hydrogeological connections of the LKS for both high and low piezometric conditions. Hydrodynamic characteristics (transit time, maximal speeds in meters/hour) are indicated for each of the underground connection.

LKS Dynamic During High Piezometric Conditions
Dye tracer injection from the Kerwée sinkhole demonstrates the connections with the northern underground rivers of the Thiers des Falizes area: the Hôtel and Muret caves (Figure 6). From those points, the underground river joins the southern river of Thiers des Falizes within the Fosse aux Ours cave. The north river of the Fosse aux Ours confluence is the underground Wamme River, flowing in the Fromelennes Formation with a high speed of more than 140 meters/hour.

Tracer injection inside the Mortier sinkhole highlights a connection with the Lorette cave river, as expected from previous studies (Delbrouck, 1974). This underground

**Figure 6.** Schematic map of LKS dynamic during high piezometric conditions based on the tracer test campaign of February 2014.
Lomme River continues toward the west and emerges after 40 hours at the Eprave resurgence. For these hydrological conditions, no evidence was found for a connection with the Thiers des Falizes area. During high piezometric condition, the underground Lomme seems to be drained by an adjacent system.

A last injection from the Trouvée sinkhole shows a connection with the Eprave spring in almost 60 hours. This little tributary system is also independent from the Thiers des Falizes area. The origin of the Fosse aux Ours river for high piezometric conditions is still unknown, further tracer test campaigns will be made in order to complete the high water condition scheme.

### LKS Dynamic During Low Piezometric Conditions

The tracing during low piezometric conditions shows a very different organization of the flow paths inside the LKS, both for the Wamme and the Lomme systems (Figure 7).

Regarding the underground Wamme River, a very clear tracer signal was detected in the Thiers des Falizes area at the three monitoring stations (Hôtel, Muret and Fosse aux Ours rivers). Low piezometric conditions seem to connect the Wamme river with the southern river of this area, which was not the case during high piezometric conditions. The location of the diffluence of the Wamme river is still unknown (?? on Figure 7).

As for the underground Lomme, a connection was proved between the Pré-au-Tonneau sinkhole and the Lorette cave river. The flow speeds are very slow (30 m/h) indicating a larger flow section, very different with the Mortier sinkhole – Lorette river connection.

From the latter, the water flows to the Fosse aux Ours cave in 8 to 13 hours, which means a maximal speed of 115 to 72 m/h. Flow conditions seems to be relatively identical to the previous section as shown by the BTC. At the Fosse aux Ours Cave, the underground Lomme meets the underground Wamme and the resulting river flows toward the Eprave resurgence.

### Conclusions and Perspectives

The KARAG project aims to understand the dynamic of karst aquifers by using both geophysical and hydrogeological tools. For this purpose, the Lomme Karst System (Rochefort, Belgium) has been chosen and studied since 2013. One of the first objective of the
project is to draw the hydrogeological behavior of this karst system: flow connections inside the karst system, reactions to various hydrological conditions, relations to the surface Lomme River. So far, 20 water probes have been installed within the LKS, providing high resolution data for further analysis. In 2014, 9 tracer tests using 5 GGUN-FL30 fluorometers were conducted, leading to a better understanding of the LKS dynamics. Two main underground rivers were identified but the results also highlight the variability of the karst hydraulic connections as aquifer water level conditions change.

Further work will be focused on the probes data: piezometric levels to define the aquifer dynamic, temperature, electrical conductivity recording and chemical analysis. All this information together with the tracer test results will help us to build a complete conceptual model for the LKS. Finally, the hydrological data will be used in addition to geophysical tools (ERT, gravimetry) in order to evaluate the capability of such techniques to measure karst and epikarst aquifer dynamics.

References
RECHARGE AREA OF SELECTED LARGE SPRINGS IN THE OZARKS

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Abstract

Ongoing work by the Missouri Geological Survey (MGS) is refining the known recharge areas of a number of major springs in the Ozarks. Among the springs being investigated are: Mammoth Spring (Fulton County, Arkansas), and the following Missouri springs: Greer Spring (Oregon County), Blue Spring (Ozark County), Blue/Morgan Spring Complex (Oregon County), Boze Mill Spring (Oregon County), two different Big Springs (Carter and Douglas County) and Rainbow/North Fork/Hodgson Mill Spring Complex (Ozark County). Previously unpublished findings of the MGS and USGS are also being used to better define recharge areas of Greer Spring, Big Spring (Carter County), Blue/Morgan Spring Complex, and the Rainbow/North Fork/Hodgson Mill Complex (Ozark County).

MGS is applying a graphical method of data analysis using spectrofluorometric scan results. Comparing the dye peak intensity to the intensity of the valley preceding the peak yields a ratio that can be used to standardize and quantify water traces. This method can be applied to current and some legacy traces with comparable results.

In some cases, past tracer injection sites were utilized in attempts to replicate older traces from those locations. The data clearly show that there is value in applying spectrofluorometric dye detection techniques in attempting to replicate older traces. Some repeat injections and subsequent monitoring confirmed earlier traces. Other replication efforts revealed multiple recovery points that were undetected by the legacy traces, thus expanding known recharge areas. Still other replication efforts indicate that some older traces are not repeatable. The effect of the replication efforts significantly changes the logical interpretation of a number of recharge area boundaries.

Among the findings of the overall study to date: Mammoth Spring and Greer Spring share a portion of their recharge, with the majority of Greer Spring’s flow apparently passing under a gaining segment of the 11 Point River, ultimately emerging more than four kilometers to the southeast. This and other findings raise questions about how hydrology in the study area may be controlled by deep-seated mechanisms such as basal faulting and jointing. Research and understanding would be improved by 1:24,000 scale geologic mapping and increased geophysical study of the entire area.

Introduction

In the latter half of the 20th century, the US Forest Service and the National Park Service began serious efforts to define the recharge areas of some of the largest springs in the Ozarks of south central Missouri using water tracing techniques. The scientific community is indebted to committed professional investigators such as Tom Aley, Everett Chaney, Mickey Fletcher, and Chuck Tryon who contributed to the work in this region. In addition, Jim Vandike as a graduate student conducted numerous traces in the North Fork watershed.

These water tracing efforts were groundbreaking since there had been no previous organized effort to determine recharge areas of these springs that are among the largest in the nation. Techniques used were state-of-the-art for US investigators at the time. Most of the work employed activated carbon packets to collect fluorescein dye. Packets were anchored in springs and other monitoring points, and replaced on a routine basis. The removed packets were then washed in water and dye eluted via a basic solution. Dye detection was routinely conducted by visual analysis of the eluate from each packet.

While most of the older traces still seem to fit into existing paradigms with respect to the recharge areas of the major springs, questions have arisen about the accuracy of some. Beginning in 2013, the Missouri Geological Survey (MGS) conducted a number of dye traces with the goal of better defining the recharge areas of Big Spring in Carter County, Missouri, Greer Spring,
Blue Spring and Morgan Spring in Oregon County, Missouri, Rainbow Spring, North Fork Spring, Hodgson Mill Spring, and Blue Spring in Ozark County, Missouri and Mammoth Spring in Fulton County, Arkansas. Most of this work was completed using techniques that were similar to those used decades ago: carbon packets as dye collectors, eluted in a basic alcohol solution. The biggest changes follow:

- Multiple tracers were used with eluate analyzed with a fluorescence spectrometer at the MGS Water Tracing Laboratory.
- Additional monitoring points were used based on past work and new findings.
- Monitoring was more extensive with multiple packets collected to characterize background conditions and to obtain more complete dye recovery curves.
- Recovery curves were plotted using Peak Valley Ratios (PVR) to compare relative dye intensities.

Fluorescence spectrometry methods remove some of the inherent subjectivity of visual detection and quantification methods and allow the use of multiple fluorescent tracers. They also improve detection limits. Some time-of-travel information was obtained with a fluorometer and associated logger placed in Mammoth Spring.

It should be noted that the intent of the current study is not to undermine the pioneering work that was completed decades ago by a number of forward-thinking individuals and organizations. Our goal is improvement of the data-base by adding to, enhancing, and refining the work of those pioneers.

Geohydrology of the Study Area

Detailed geologic mapping is not available for most of the study area. Reconnaissance quality mapping shows that the bedrock near the surface is nearly horizontal, fractured and solutioned dolostone of the Ordovician age Cotter, Jefferson City, Roubidoux and Gasconade formations with some fractured sandstone of the Roubidoux Formation. In many areas the Jefferson City and Cotter formations display relatively low permeability, but locally they are prone to extremely deep weathering with attendant sinkholes, and losing streams. The Roubidoux Formation is typically fractured and permeable and host to karst features such as sinkholes and losing streams. The upper part of the Gasconade Formation is one of the prime strata for cave development in the central Ozark region. While the Cambrian aged Eminence Formation (also largely comprised of dolostone) is not well exposed in the study area, deeper groundwater circulation does move through voids in this unit.

Past tracing efforts (Kleeschulte and Duley, 1985) have shown that geologic structures can be crucial in directing flow in karst aquifers in Missouri. The limited geologic mapping that has been done in the study area has shown that faulting is common with some notable structures discernible even on reconnaissance mapping (Figure 1). Work by Lowell et al. (2010) and Weary et al. (2014) implies a number of sizable structures apparently interpolated from remote imagery. But without detailed geologic mapping, the impact of these structures can only be surmised.

The Ozarks of southern Missouri and northern Arkansas contain some of the best examples of karst features observed anywhere in the US. The big spring region in south-central Missouri contains the recharge areas of a number of first order magnitude springs including the three largest in the Ozarks: Big Spring (with a mean discharge of 12.7 m³/s in Carter County, Missouri) Mammoth Spring (with a mean discharge of 9.9 m³/s and located just south of the Missouri state line in Fulton County, Arkansas) and Greer Spring (with a mean discharge of 9.8 m³/s in Oregon County, Missouri). In addition, water tracing has shown that Rainbow Spring, North Fork Spring and Hodgson Mill Spring are all part of a single complex. While discharge information is limited for the springs in the Rainbow/North Fork/Hodgson Mill complex, the combined mean discharge is about 6 m³/s. Thus the complex would the considered to be the fourth largest system in the Ozarks.

Prioritizing Traces for Replication

While many historical traces completed in the karst region of south central Missouri and north central Arkansas agree with current interpretations of recharge areas, it should be noted that recharge area boundaries shown in recent publications (Mugel et al, 2009; Imes et al, 2007) are largely based on traces completed in the latter part of the 20th century with some seepage run and potentiometric data for support. Prior to the current study, few of the traces had been replicated with fluorescence spectrometry methods. As a result some of the older traces have come under scrutiny for a variety of reasons.

In order to prioritize traces for replication, MGS staff considered a number of issues. One overriding concern was selection of traces where a reasonable monitoring effort could produce the most valuable data in a relatively short period of time due to budget constraints. The following factors were considered as additional reasons to attempt to replicate traces in the study area:

- A small number of traces have been discounted in subsequent studies.
A limited number of traces appear to conflict with available potentiometric data.

Essentially all of the older traces in the study area were completed with subjective (visual) detection methods.

Several traces were found to have questionable timing of injections causing a possible overlap with other injections.

Several traces had a single reported recovery, or no detailed recovery information at all.

Traces near the boundary of the recharge area in question were replicated in some instances because of the relative importance placed upon them with respect to planning and impact on future studies.

Traces with inadequate, or no other monitoring points recorded were considered as needing replication because of the uncertainty with respect to other potential recoveries related to the original injection.

**Current MGS Techniques for Dye Collection and Analysis using Carbon Packets**

The standard protocol used by MGS in recent years involves use of activated coconut charcoal (6 to 12 mesh) to collect fluorescent tracers in the field. Charcoal is placed in packets constructed of nylon window screen. The packets are then secured in the water to be monitored. They are replaced on schedules varying from weekly to monthly and analyzed in the MGS Water Tracing Laboratory using fluorescence spectrometric techniques. Background packets are normally collected and analyzed for periods ranging from weeks to months prior to dye injection.

Charcoal packets are washed and placed in 100 milliliter cups with screw top lids. The dyes are eluted from the charcoal using a solution of 5% ammonium hydroxide in ethyl alcohol for one hour. Eluate is pipetted from the charcoal and placed in disposable polystyrene cuvettes.

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**Figure 1. Geologic Map of the Study Area.**
produced specifically for fluorescence studies. Currently analysis is completed using a Hitachi F7000 synchronously scanning fluorescence spectrophotometer. Analysis involves synchronously scanning the eluate with a seventeen (17) nanometer (nm) separation between excitation and emission using procedures similar to those developed by Duley (1986). Using current protocol the Hitachi measures and records fluorescence intensity from 450 nm to 620 nm excitation (467 nm to 637 nm emission). Results are plotted on the excitation wavelength axis.

Fluorescent dyes are typically indicated by the presence of background-modified normal (Gaussian) peaks. Dye peaks are superimposed on background fluorescence, with the greatest background impact occurring at lower wavelengths. Fluorescein is indicated by a well-defined peak between 495 nm and 505 nm. Eosine is indicated by the presence of a well-defined peak occurring between 521 nm and 525 nm. Rhodamine WT is indicated by a well-defined peak occurring between 544 nm and 553 nm.

When more than one tracer is present in the same eluate, their fluorescent spectra overlap somewhat. If one tracer is found at a particularly high concentration, detection of other tracers may be problematic. For this reason MGS staff often staggers dye injections to allow the typically large pulse of one tracer to dissipate prior to recovery of a second or third tracer. MGS staff have also developed a simple chemical method to lower the pH of samples to a level that lowers fluorescein fluorescence so that it does not overshadow eosine or Rhodamine WT. It should be noted that the peak wavelengths of both eosine and Rhodamine WT increase (up to several nanometers above normal) when this method is used.

In order to compare data from a broad array of analytical instruments used over the last thirty years, MGS developed a simple method to normalize spectral data. When possible, instrumental dye peak height value (as measured from zero on the X-axis in any instrumental unit) is divided by the value of the lowest intensity of the spectral valley that occurs between the fluorescent background maxima and the dye in question (Figure 2).

This peak to valley ratio (PVR) gives the investigator a quick and simple way to compare data from a variety of instruments ranging from those with analog printouts to digital data. By definition the PVR must be greater than 1.0 to indicate the presence of dye. If there is no dye peak, there is no valley and the PVR is assumed to be 1.0. MGS interpretations of different PVRs are shown below:

- A PVR of 1 is indicative of background conditions since, by definition, there is no dye peak.
- A PVR of 1 to 2 is a small dye recovery. This level in background packets is indicative of sewage or another limited source of tracer.
- When the PVR is greater than 2 and up to 5 a moderate dye recovery, normally from a current injection, is indicated.
- A PVR greater than 5 and up to 10 is a large dye recovery that indicates a strong connection to a current injection.
- A PVR greater than 10 is a very large recovery that indicates a strong and direct connection to a current injection.

Results and Recommendations
A combination of new injection points and repeat injections at locations used for legacy traces has helped to create a clearer picture of the recharge areas of some of the springs in question (Figure 3 and Table 1).

Replication efforts of older traces show the value of newer techniques. A limited number of repetitions show that some past investigators may have reported positive recoveries of dye erroneously. They also show that some actual resurgence points were not monitored or tracers were not detected due to limitations of the older detection methodologies. In addition, injection overlap may have been a serious problem in some traces.

While geologic conditions such as degree of bedrock fracturing, degree of weathering, and location of faults,
Figure 3. Recharge Areas of Selected Springs in the Southern Ozarks. Older dye traces referenced in this figure are discussed in Aley (1972; 1975), Vandike (1979), and Aley and Aley (1987). Replication attempts are discussed in Imes and Frederick (2002) and Imes et al. (2007). Results of more recent traces are discussed in Gilman et al. (2008).

Folds and discontinuous low permeability beds are the prime factors that determine how groundwater moves in this region, insufficient data are available to predict groundwater recharge areas using existing geologic mapping alone. Thus recharge area boundaries on Figure 3 are approximate and were drawn based on the following factors (listed in order of importance):

1. Sound and repeatable dye traces
2. Potentiometric data from groundwater and surface water
3. Losing/gaining stream determinations
4. Water quality

Data gathered during this study shows that the recharge area of the Blue Spring/Morgan Spring Complex extends to the west and south of the city of Alton, likely encompassing most of the drainage of Frederick Creek (excluding the upper reaches of Piney Creek tributary). Piney Creek, which flows through the city of Alton, is largely in the recharge area of Boze Mill Spring. It is likely that during wetter periods, surface flow carries some water into the lower reaches of Piney Creek and/or Frederick Creek basins where it is subsequently lost to the subsurface and resurfaces at the Blue Spring/Morgan Spring Complex. This connection is minor as it is transitory in nature with little actual water movement between the upper reaches of Piney Creek and the Blue Spring/Morgan Spring Complex.

New traces in the recharge area of Mammoth Spring show that some early traces are accurate or at least partially so. However, new traces from replicated injection points show that a significant portion of the northern end of the Mammoth Spring recharge area is shared with Greer Spring and Blue Spring on the North Fork. This extends the known recharge area of both Mammoth Spring (to the northeast) and Greer Spring (to the southwest). Significant questions still remain about the boundary between Greer Spring on the north and Mammoth Spring on the south.
Table 1. Basic Information about Recent MGS Traces in the Study Area.

<table>
<thead>
<tr>
<th>Trace #</th>
<th>From</th>
<th>To</th>
<th>Dye* (Amount)</th>
<th>Distance in Km</th>
<th>Injection Date</th>
<th>First Recovery Velocity (m/hr)</th>
<th>Peak Recovery Velocity (m/hr)</th>
<th>Peak PVR</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Frederick Creek</td>
<td>Blue/Morgan Complex</td>
<td>Fl (4.5 kg)</td>
<td>25.9</td>
<td>8/27/2013</td>
<td>36-49</td>
<td>30-36</td>
<td>33.9</td>
<td>Principle dye recovery point.</td>
</tr>
<tr>
<td>2</td>
<td>Frederick Creek</td>
<td>Boze Mill Spring</td>
<td>Fl (4.5 kg)</td>
<td>22.4</td>
<td>8/27/2013</td>
<td>27-31</td>
<td>22-27</td>
<td>2.1</td>
<td>Secondary dye recovery point.</td>
</tr>
<tr>
<td>3</td>
<td>Grand Gulf</td>
<td>Mammoth Spring</td>
<td>RWT (11.4 l)</td>
<td>11.0</td>
<td>10/9/2013</td>
<td>88</td>
<td>49</td>
<td>15.9</td>
<td>Only dye recovery point. Conffirms earlier trace.</td>
</tr>
<tr>
<td>4</td>
<td>Pinney Creek in Alton</td>
<td>Boze Mill Spring</td>
<td>RWT (11.4 l)</td>
<td>18.2</td>
<td>12/4/2013</td>
<td>54-96</td>
<td>40-54</td>
<td>133.2</td>
<td>Principle dye recovery point.</td>
</tr>
<tr>
<td>5</td>
<td>Pinney Creek in Alton</td>
<td>Blue/Morgan Complex</td>
<td>RWT (11.4 l)</td>
<td>24.1</td>
<td>12/4/2013</td>
<td>29-39</td>
<td>25-29</td>
<td>1.3</td>
<td>Secondary dye recovery point.</td>
</tr>
<tr>
<td>6</td>
<td>Mastion Creek</td>
<td>Mammoth Spring</td>
<td>Fl (4.5 kg)</td>
<td>40.7</td>
<td>2/14/2014</td>
<td>85-153</td>
<td>68-85</td>
<td>217.3</td>
<td>Only dye recovery point.</td>
</tr>
<tr>
<td>7</td>
<td>Rattlesnake Spring</td>
<td>Mammoth Spring</td>
<td>Eos (4 kg)</td>
<td>56.6</td>
<td>2/26/2014</td>
<td>81-117</td>
<td>69-81</td>
<td>9.2</td>
<td>Did not agree with legacy traces.</td>
</tr>
<tr>
<td>8</td>
<td>Rattlesnake Spring</td>
<td>Greer Spring</td>
<td>Eos (4 kg)</td>
<td>60.5</td>
<td>2/26/2014</td>
<td>87-90</td>
<td>74-87</td>
<td>13.9</td>
<td>Did not agree with legacy traces.</td>
</tr>
<tr>
<td>9</td>
<td>Rattlesnake Spring</td>
<td>Blue Spring N. Fork</td>
<td>Eos (4 kg)</td>
<td>13.5</td>
<td>2/26/2014</td>
<td>12-14</td>
<td>7</td>
<td>4.2</td>
<td>Did not agree with legacy traces.</td>
</tr>
<tr>
<td>10</td>
<td>Dry Creek near Pomona</td>
<td>Mammoth Spring</td>
<td>RWT (11.4 l)</td>
<td>52.7</td>
<td>3/26/2014</td>
<td>81-110</td>
<td>81-110</td>
<td>8.4</td>
<td>Recovery point that replicates legacy trace.</td>
</tr>
<tr>
<td>11</td>
<td>Dry Creek near Pomona</td>
<td>Greer Spring</td>
<td>RWT (11.4 l)</td>
<td>54.6</td>
<td>3/26/2014</td>
<td>84-114</td>
<td>81-114</td>
<td>10.2</td>
<td>Additional recovery found in replication attempt.</td>
</tr>
<tr>
<td>12</td>
<td>Upper Tabor Creek</td>
<td>Blue Spring N. Fork</td>
<td>Fl (2.2 kg)</td>
<td>13.9</td>
<td>6/17/2014</td>
<td>28-85</td>
<td>21-28</td>
<td>246.9</td>
<td>Only dye recovery point.</td>
</tr>
<tr>
<td>13</td>
<td>Granny Meyers Spring</td>
<td>Mammoth Spring</td>
<td>Eos (2.2 kg)</td>
<td>23.4</td>
<td>7/16/2014</td>
<td>36-48</td>
<td>29-36</td>
<td>29.9</td>
<td>Only dye recovery point. Conffirms legacy trace.</td>
</tr>
<tr>
<td>14</td>
<td>Upper Fox Creek</td>
<td>Rainbow/North-Fork Comp.</td>
<td>Eos (2.2 kg)</td>
<td>38.2</td>
<td>8/19/2014</td>
<td>116-206</td>
<td>78-116</td>
<td>98.7</td>
<td>Recovery point closely tied to Hodgson Mill.</td>
</tr>
<tr>
<td>15</td>
<td>Upper Fox Creek</td>
<td>Hodgson Mill Spring</td>
<td>Eos (2.2 kg)</td>
<td>37.5</td>
<td>8/19/2014</td>
<td>122-198</td>
<td>75-112</td>
<td>88.3</td>
<td>Recovery point tied to Rainbow/N. Fork Comp.</td>
</tr>
<tr>
<td>16</td>
<td>11 Point at Lost Hill</td>
<td>Greer Spring</td>
<td>Eos (2.2 kg)</td>
<td>26.1</td>
<td>10/28/2014</td>
<td>69-186</td>
<td>69-186</td>
<td>18.2</td>
<td>Only dye recovery point.</td>
</tr>
<tr>
<td>17</td>
<td>Middle Fork 11 Point</td>
<td>Bill Mac Spring</td>
<td>RWT (11.4 l)</td>
<td>8.4</td>
<td>12/11/2014</td>
<td>14-19</td>
<td>9-11</td>
<td>5.0</td>
<td>Did not agree with legacy traces.</td>
</tr>
<tr>
<td>18</td>
<td>Warm Fork Losing Stream</td>
<td>Warm Fork Spring</td>
<td>Fl (0.45kg)</td>
<td>4.8</td>
<td>1/21/2015</td>
<td>&gt;32</td>
<td>&gt;32</td>
<td>48.5</td>
<td>Only dye recovery point.</td>
</tr>
<tr>
<td>19</td>
<td>Cave Spring on Spring Cr.</td>
<td>Greer Spring</td>
<td>Fl (2.2 kg)</td>
<td>17.7</td>
<td>1/21/2015</td>
<td>26-37</td>
<td>18-26</td>
<td>7.5</td>
<td>Only dye recovery point.</td>
</tr>
</tbody>
</table>

*Fl=Fluorescein, RWT=Rhodamine WT 20%, Eos=Eosine

Table 1. Basic Information about Recent MGS Traces in the Study Area.

Data also show that, while Upper and Lower Greer Spring are clearly part of the same system, on one occasion, dye recovery at Upper Greer Spring was shown to lag behind the recovery at Lower Greer Spring, presumably due to the head differential and circuitous route for discharging water flowing to the surface at the higher cave outlet of Upper Greer. Yet another trace demonstrates that the Eleven Point River contributes recharge to both Greer Springs from the region about 8 km upstream of Thomasville. Figure 4 is a graphical depiction of this trace. The new information shows that much of Greer Spring’s recharge travels beneath a gaining section of the 11 Point River, surfacing 4 km to the southeast. One new trace was completed from an upper reach of Tabor Creek to Blue Spring on the North Fork which helps to define the boundary between the recharge areas of Blue, Greer and Mammoth springs. A new trace was completed during this study to the Rainbow/North Fork/Hodgson Mill Complex which extends the known recharge area to include upper Fox Creek. At least one legacy connection to the Complex completed using visual methods was discounted by an attempt to replicate the old trace. It is recommended that additional tracing be conducted to determine what amount (if any) of the recharge area of the Complex is located east of the North Fork River. It is also recommended that additional work be done to better define the relationship between Hodgson Mill Spring and the main resurgences of the Complex at Rainbow and North Fork Spring.

As a final recommendation, the entire region would benefit from detailed geologic mapping. The limited amount of available detailed mapping gives clues helpful in predicting groundwater flow direction and resur-
gence points but much more work is required to get a more complete picture of bedrock influences on the karst system. Geologic mapping at a scale of 1:24,000 is in dire need to show the location of structures involved in directing groundwater movement. This mapping would benefit those dealing with water supply, waste disposal, and watershed protection issues alike.

References

HYDROLOGICAL AND HYDROGEOLOGICAL CHARACTERISTICS OF THE PLATFORM KARST (ZEMO IMERETI PLATEAU, GEORGIA)

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Abstract  
The article discusses the hydrological and hydrogeological characteristics of the platform karst of Zemo Imereti plateau. The structural plateau of Zemo Imereti is the part of the intermountain plain karst zone of Georgia and one of the interesting parts of the karst relief development. The above mentioned karst region includes the easternmost part of western Georgia, which is characterized by peculiar natural conditions (relief, tectonics, climate, surface and underground streams) and represents one of the significant platform karst regions in the Caucasus. On the basis of the cartographic materials analysis and borehole data, the general scheme of hydrogeological situation of the Zemo Imereti structural plateau (two hydrogeological basins were defined) was created as confirmed by experiments. In addition, it was identified that underground karst water flowing from the periphery to the center determines sedimentation together with the broken dislocations within the frame of the structural plateau. The study found that within the Chiaatura structural plateau the joint karst hydrogeological system (with enough dynamic water resources) has been established, which mainly is unloaded in sources of Ghrudo vauculise and the surrounding area (local erosion basis). Ghrudo hydrogeological system and Chiautra structural plateau are characterized by the systems of isolated karst-fissure waters with different hypsometric location and orientation. Therefore, based on these studies, it could be said that in karst areas the structural features can define the characteristic of groundwater circulation, but karst age can also make a significant adjustment.

Introduction  
Conflicting views on underground karst water circulation has been formed in the very beginning of the systematic study of the karst. Certain Western European researchers argued that there should be a common underground water level in the karst districts (Grund, 1903). In contrast, others gave advantage to the theories of the isolated systems (Martel, 1894; Knebel, 1906; Katzer, 1909). In this discussion, the results of which still did not lose their importance, A. Kruber (1915), V. Varsanofieva (1915) and other researchers actively participated. Later the mentioned issues were generalized based on the rich actual materials in the works of the former Soviet researchers as well (D. Sokolov, 1962; Z. Tintilozov, 1976 and G. Gigineishvili, 1973).
There were attempts to explain these opposing theories by (1) structural and geological conditions; the overall level of underground waters are related to the platform conditions and the isolated ones – to the geosynclinal conditions (G. Gigineishvili, D. Tabidze, 1975; G. Tintilozov, 1976); (2) different capacities of karsting rocks; the overall level of underground waters are related to the “thin” karst and the isolated ones – to the “deep” karsts (Gèze, 1965); and (3) different stages of karst development; the overall level of underground waters are related to the “mature” and “old” stages and the isolated ones – to the “young” stage (Kruber, 1915).

In the limestone belt of the western Georgia, which is a part of the Caucasus orogenic section, the isolated underground basins (Bzipi and Arabika limestone massifs) are identified by the Georgian researchers on the basis of fundamental researches (Kiknadze, et al., 1973; Maruashvili, et al., 1971; Tintilozov, 1976). In addition, in the easternmost part of the mentioned karst zone, which is known as the Zemo Imereti karst plateau, there is an assumption of the existence of the common level of the karst waters (Gigineishvili, et al., 1975). This point of view gained precision by the methods and experiments used during the study.

Description of the study area and materials

Zemo Imereti plateau is a compound part of the limestone belt and is critical to the development of the karst relief in Georgia’s intermountain plain. The aforementioned region includes the easternmost part of the limestone belt of the western Georgia, which is characterized by the peculiar natural conditions (relief, tectonics, climate, surface and underground flows) and is one of the most important regions of the plate karst in the Great Caucasus (Figure 1).

The boundary of the Zemo Imereti karst region lays through the Cretaceous limestones surface contact line with the formations of older age (Bajocian porphyritic series to the north and east, and to the south and west – the Middle Paleozoic granitoids). Cretaceous limestones base is created by the Paleozoic formations, which suffered denudation during the Upper Jurassic and partially during the Lower Cretaceous periods, resulting in formation of the Likhi peneplain (Maruashvili, et al., 1971).

The presence of the solid platform with a peneplained surface caused a laying character (still, almost horizontal or slightly inclined layers laying) of the Mesozoic-Cenozoic cover on it, which is represented by the Valangin-
ian - Hauterivian, Barremian and Turonian-Danish limestones, Tertiary clays and sandstones. Sedimentation of the mentioned deposits occurred in the platform conditions and in connection with it their total capacity does not exceed 500-550 meters.

Morphologically the Zemo Imereti karst region is a structural plateau, fragmented down to the 150-250 m depth. The Kvirila River valley divides the Zemo Imereti Plateau into the two parts – of northern (the right side of the Kvirila River with the area of 116 sq. km) and southern (the left side of the Kvirila River with the area of 208 sq. km). The latter is represented in the form of a single plateau, the surface of which is built by a powerful Tertiary clay layers. The clay layers cover limestones and hinder the development of karst. Morphologically the northern part looks somewhat complex. Its once-fragmented plateau surface is separated by the narrow gorges of the Kvirila River’s right tributaries: Jruchula, Nekrisa, Bogiristskali, Tabagrebistskali, Rganisghele and Katshkura, among which are erected morphologically separated plateaus: Bajiti, Darkveti, Mghvimevi, Rgan and Katskhi (their absolute heights are 550–650 m above mean sea level). Turonian-Danish limestones, which build the surfaces of the mentioned plateaus, are covered with the clays and sandstones containing the Tertiary manganese ore of little capacity over the small area. The slopes of the mentioned plateaus are steep; their relative heights are gradually decreased from the 230 m to 100 m from the west to the east along the Kvirila River gorge. Dense network of water absorption of the relief on the structural plateau of Zemo Imereti, surface and underground karst forms, etc. indicate the intensity of karst development.

Kvirila River and its tributaries do not cross the limestone cover above the city of Chiatura up to its base – the Likhi peneplain. Therefore, all morphologically isolated plateaus to the both sides of the Kvirila River have a common limestone base, which creates the favorable conditions for the formation of the common level of the underground waters in case of proper structural conditions.

Research goals, methods and results
We have carried out a karst-hydrological and hydrogeological works in order: 1) to identify the peculiarities of the underground karst waters laying conditions, movement and their determining factors; and 2) to solve certain practical issues within the Zemo Imereti structural plateau (the underground karst waters are widely used in the water supply of the city of Chiatura and the villages

Figure 2. Fault Dislocation Scheme of the Chiatura Structural Plateau (Compiled by decoding of the aerial images)
located near the plateau. Periodic pollution of these waters causes a number of delays in water supply system. In this regard, it was necessary to identify the water pollution centers, their movement lines, etc.). We used driller’s well logs and geological sections data, aerial images structural decoding and electrometric search methods, as well as the groundwater tracing experiments. On the basis of the conducted studies we obtained the results of interesting scientific and practical value.

For advanced study of the tectonic setting (faults, shearing, tectonic fractures) in Zemo Imereti (Chiatura) structural plateau we carried out structural decoding of the aerial images of the territory. It enabled us to draft a detailed schematic of a fault dislocation and to identify the regularities of karst forms distribution. Decoding has revealed a dense network of previously unknown faults and fractures of different direction (Figure 2).

We determined that the tectonic shearnings and fractures control the underground waters absorption, movement and discharge. Fractures of sublatitudinal and submeridional direction are particularly distinguished, to the sectors of intersection of which are related the majority of surface and underground karst forms and the karst processes (karst forming) intensity in general.

Figure 3. Geological Section of the Drilling Wells in the Chiatura Structural Plateau

Borehole data (Figure 3) and cartographic material (geological profiles, etc.) analysis were basis for the topographic scheme of the limestone (Mesozoic-Cenozoic) bed, where the buried morpostructures were identified (Figure 4).

Analysis of the limestone bed topography and buried morphostructure clearly shows that to the east of the structural and denudation plateau dividing shearing submergence of karsting rock beds starts to the north-east or to the Sachkhere syncline mold. It seems that the conditions of their occurrence and inclination stipulate the groundwater movement and direction in the study area (the role of the shearing and fractures is considerable as well). This situation completely changes our opinions and views about the underground water discharge centers within the plateau. According to these views, the water discharge was assumed to be in the western part of the Kvirila River canyon gorge composed of limestones and they thought it was possible for that part of it was flowing into the Kolkheti artesian hydrogeological basin.

Based on the Mesozoic-Cenozoic bed and fault dislocations’ (decoding) schemes we compiled the general scheme of the structural plateau’s hydrological setting. In the scheme, we singled out the two hydrogeological
Figure 4. Topography of the Tectonic Upper Floor (Mesozoic-Cenozoic) Underlayer of the Chiatura Structural Plateau
basins with the water storage area or the vast water reservoirs, which are the prospective areas for receiving the drinking water (Figure 5).

During distinguishing the separate blocks by the fault lines the following fact was taken into account: the crystalline basement is upraised with the different heights within the Zemo Imereti structural plateau and the cross section and capacity of the sedimentary cover is also different.

In addition, the hotbed structures on forming the mechanical strain often play a role of the tectonic stamp and they control the faults. It seems that the directions of the underground karst water movement is different inside the individual blocks singled out by the faults, though it is possible that there is a water cycle among the blocks as well. Directions of the underground karst water movement within the individual blocks and within the hydrological basins in general are confirmed by the dye tracing experiments carried out by us (Figure 6).

In addition, in many cases the dyed water crossed the single plateaus and the river beds dividing them from the bottom and was fixed in the springs flowing out in the Kvirila River gorge. It should be noted that the dye tracing experiments were conducted in conditions where the underground waters were at different levels, which showed interesting results. Namely, in one case, the water was spread over the area and it came out simultaneously through the several karst springs. In another case, a number of isolated systems were observed, which are connected to each other by tunnels or narrow fracture systems; they are located at different heights and it seems the water cycle takes place during periods of high water levels.

**Figure 5.** Paleomorphostructural Zoning of the Submerged Block of the Chiatura Structural Plateau (By the General Scheme of the Hydrogeological Situation)
Thus, for the first time it has been shown that within the Zemo Imereti (c. Chiatura) structural plateau the movement directions of the underground flows are mainly determined by the structural features, namely: total submergence of the karsting rocks from the periphery to the center, as well as shearing dislocations, which greatly control the groundwater flow absorption and their movement ways. In addition it was identified that the Zemo Imereti structural plateau’s underground karst water regime is characterized both by the streams with common level (tracings painted widely open water) and by separate streams as well, which makes doubtful the recognition of absolute reality of the platform type. As a result of the dye tracing experiments the boundaries of the feeding basins of Ghurdo and other springs have been specified. It seems that within the mentioned boundaries the common karst-hydrogeological system (with the dynamic water resources) is formed, the waters of which are discharged mostly the Ghurdo and the springs running in its surroundings. The fact is of great practical importance because it is possible the contaminated waters to be flowed into the Chiatura water supply system from any karsted place within the mentioned boundaries, and which should be considered during the anthropogenic impact on the environment. Based on the dye tracing experiment results, the recommendations have been prepared, which included the removal of some of the karst waters involved in the water supply system.

**Discussion and Conclusion**

Thus, the morphostructures of the Chiatura structural plateau (western Georgia) are presented as a mosaic of different sizes elevated or lowered blocks, which were emerged during the evolution process of the regional and local submeridional and sublatitudinal shearings. Most of the old faults undergo the transformation and rejuvenation along with the tectonic movements, which is promoted by the intense karst processes.

The study area is located within the vertical (epirogenetic) uplifting zone. Contemporary geomorphological cycle (erosion, karst and other forms origination and development) has begun in the Post-Miocene in the Zemo Imereti plateau and its neighboring regions. The Post-Miocene tectogenesis was strong in the Caucasus folded zone while the activity in the Zemo Imereti platform was expressed mainly in vertical (epirogenetic) uplifting, which was accompanied by a small local faults and wave folds. The last manifestation of magmatic activity would have occurred in the same period which emerged the Goradziri, Perevisa and other laccolite-extrusions; also, the central type structures (volcanic apparatus or intrusives of isomeric forms) in the Pre-Cretaceous substrate which control the faults and play the tectonic stamp role (Dvalashvili, et al., 2014).

The Post-Miocene period’s uplifting of the study area is continuing even today. Neotectonic movements obviously played a considerable role in the evolution of the erosion fragmentation and karst genesis processes and fissure-karst water vertical hydrodynamic zones. Along with the uplifting tectonic movements, the rivers have generated deep canyon gorges (tectonic fractures played the important role in the canyon gorges genesis and the identification of the river direction), which created favorable conditions for deep-water circulation.

The wide spreading of the disjunctive dislocations in the Zemo Imereti plateau determined the formation of the independent streams, the common basis of which is the Kvirila River. In addition, at present stage the single karst caves and vaucluse springs canals which were formed still at early stages, were combined into the Ghurdo common hydrogeological system (Figure 7). It is still inaccessible, though, the development of the hollows in its basins is intense.

It seems that the ascending tectonic movements were not continuous, but had periods of slowdown, as evidenced by the level distribution of the caves in the Kvirila River.

**Figure 6. The Groundwater Stream Movement Scheme Obtained by the dye tracing experiments carried out in the Chiatura Structural Plateau.**

![Diagram](image-url)
and Jruchula River gorges. The caves of different evolution stages are placed in 4 to 5 levels (Lezhava, et al. 1992).

In addition, it appears that the slowdown epochs were marked by a short duration, to which indicates the widespread of the tunnel (or hole) type of undeveloped caves and poorly (in addition, fragmented) expressed terrace steps in the river gorges. Comparison of the relative heights of the cave levels and terrace steps with the absolute heights of the cave entrances and karst springs exits allows for the assumption that the ascending tectonic movements’ slowdown (delay) epoch would be at least 4 t- 5.

**References**


TRACER STUDIES CONDUCTED NEARLY TWO DECADES APART ELUCIDATE GROUNDWATER MOVEMENT THROUGH A KARST AQUIFER IN THE FREDERICK VALLEY OF MARYLAND

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Abstract
A pair of groundwater tracer studies at a single karst test site were completed 18 years apart. The results of these studies have provided evidence of both relatively rapid advective transport via conduits and an extreme capacity for dye storage and retardation. The tracer results, coupled with other subsurface investigation data, are used to develop a conceptual model for groundwater movement through this karst aquifer in the Frederick Valley of Maryland, as well as identify implications for remediation.

Three fluorescent tracer dyes used in the initial study were detected in several background monitoring locations established for the second study conducted 18 years later; demonstrating the persistence of these dyes in the aquifer. One of these dyes was not detected during the original study, providing useful information regarding flow and transport in the aquifer. At some of these sampling locations, at least one of the dyes was degraded, and would have gone undetected without the use of activated carbon samplers. Lastly, even though relatively rapid first detections occurred during both studies (as compared to non-karst groundwater systems) the majority of injected dye mass remained in the aquifer after the studies were completed. This suggests that the aquifer has a large capacity to store contaminants and that low levels of contaminants can be expected to persist in groundwater discharged from springs for a long period of time.

Introduction
The Frederick Valley lies near the western edge of the Piedmont Physiographic Province of Maryland. The Frederick Valley stretches northeastward from the Potomac River for 36 km (22 mi) and is composed of Cambrian and Ordovician carbonate rocks that are generally overlain by Triassic rocks that were deposited in two basins: the Culpeper basin to the south and the Gettysburg basin to the north. These two basins are separated by a narrow region where the Triassic rocks are absent. The Frederick Valley is bounded to the east by the regional Martic fault that juxtaposes low-grade metamorphic phyllites against the Paleozoic carbonates (Southworth, 1996). To the west, the Valley abuts the clastic rocks of the mountains of Maryland’s Blue Ridge Physiographic Province. The southward-flowing Monocacy River drains the Valley.

The Frederick Valley represents Maryland’s second largest karst terrane (Brezinski and Reger, 2002). The most common karst features are dolines (closed depressions), a subset of which are active sinkholes, and springs. Caves are rare in the Frederick Valley.

The tracer studies described herein were conducted to investigate the nature of groundwater movement near several former waste-disposal areas and identify associated groundwater discharge points. The 1995 study focused on shallow groundwater, with dyes mostly injected at or near the bedrock surface. The 2013 study focused on...
deeper intervals in the aquifer, with dyes injected into wells that intercepted solution porosity at two different depth intervals, many meters below the water table.

**Study Area Description**

The study area for this paper is located along the southern edge of the Frederick Valley (Figure 1). The topography of the area is gently rolling and is cut by a series of small streams that drain into Carroll Creek. Carroll Creek is the principal tributary in the area and flows generally southeastward, eventually joining the Monocacy River southeast of the study area. Several springs are located along Carroll Creek.

The southeastern portion of study area is underlain by fractured limestones and dolomites of the Cambrian Frederick Formation. These rocks are underlain by sandstones and siltstones of the Araby (Car, middle-lower Cambrian) and Antietam (Ca, lower Cambrian) Formations. The northeastern portion of the study area is underlain by the Triassic New Oxford Formation of the Gettysburg basin.

**Lithology**

The Frederick Formation is divided into several members. The member that directly underlies the study area is the Rocky Springs Station Member (Cfr). It is characterized as dark gray, very thinly bedded, lime mudstone interbedded and interlaminated with black, calcareous shale. This lithology is interbedded with intervals of very thick to massively bedded limestone breccia that includes various limestone lithologies and sandy limestone. The thickness of the member beneath the study area is estimated to exceed 300 m (1,000 ft) (Brezinski 2004a; 2004b).

The New Oxford Formation is characterized by brownish red to reddish gray, and locally greenish gray sandstone, interbedded with red claystone (Tn) with a conglomerate (Tnc) at its base. Beneath the study area, the conglomerate is composed of pebbles and some cobbles of white and gray limestone in a fine-grained red matrix with calcite cement. The thickness of the unit is unknown (Brezinski 2007).

**Structure**

Rocks in the region have been tightly folded and faulted; however, bedrock exposures in the Valley are rare. The study area is located on the western limb of the Frederick Valley Synclinorium, whose axis is approximately 5.5 km (3.5 mi) to the east. Bedding in the Rocky Springs Station Member near the study area generally dips toward the southeast, with dip angles ranging from 25 to 75 degrees. Geophysical logging of boreholes in the study area shows considerable local structural variability, particularly near unit contacts. Bedding in the New Oxford Formation has an approximate strike of N 10° W and dips to the west-southwest at approximately 30° near the study area (Brezinski, 2004b).

Little information is available regarding jointing in the Frederick or New Oxford Formations near the study area. In his study of the Frederick and Hagerstown Valleys, Nutter (1973) noted that a joint set parallel to the strike of the bedding is almost always present, as is a near vertical set with a strike that is nearly normal to the bedding strike. Brezinski (2004a) reports that jointing in the Grove Formation (late Cambrian-early Ordovician), which overlies the Frederick Formation, has a pervasive joint system with a primary orientation of N 72° W, and a secondary set oriented N 20° E. Jointing mapped in the Triassic Leesburg Formation during the same study exhibits a completely different trend with the main trend oriented N 87° E and a conjugate set oriented N 45° W. The Frederick and New Oxford Formations are truncated along the northwestern edge of the study area by the steeply-dipping Bull Run Mountain Fault. The study area is on the downthrown side of the fault. Across the fault are younger clastic rocks of the Antietam Formation. Faults are expected to be more extensive and numerous in the region than is shown on published maps because very detailed mapping is required to locate all but the largest and most obvious (Nutter, 1973).

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**Figure 1.** Location map of the karst regions of Maryland, bedrock geology of the Frederick Valley, and location of the study area. Modified from Brezinski and Reger (2002).
Karst Development

Studies related to karst development in the Frederick Valley focus on the occurrence and availability of groundwater in bedrock (Nutter, 1973) and susceptibility to sinkhole formation (Brezinski and Reger, 2002; Brezinski 2004a; Brezinski 2007).

Nutter (1973) believed joints to be the most important factor in controlling the formation of solution channels, and that solution activity decreased with depth. He also notes that solution cavities are apparently rare below a depth of 90 m (300 ft). It should be noted, however, that he provides little region-specific information to substantiate these statements.

Brezinski and Reger (2002) and Brezinski (2004a) examined the nature and distribution of karst features (closed depressions, active collapse sinkholes\(^1\), and springs) in the eight main carbonate rock units of the Frederick Valley. For the Rocky Springs Station Member, which underlies much of the study area, they found its susceptibility to karst formation to be low-to-moderate. They attributed the relatively low susceptibility to the thin-bedded, shaly nature of the member. The other carbonate formation that underlies the study area, the New Oxford Conglomerate, was not included in the above studies.

Approach

Fluorescent dyes and activated carbon samplers (supplemented with groundwater grab samples) were employed in both studies. During each sampling event, activated carbon samplers were collected for analysis and replaced with a fresh sampler. A grab water sample was also collected at each location during each event; but analyzed only selectively to confirm where dye had been detected in the activated carbon sampler. Groundwater tracing using fluorescent dyes is an appropriate method for investigating groundwater flow connections in karst settings. The use of activated carbon samplers during both tracer studies was critical in identifying sampling stations that received small concentrations of the introduced tracer dyes.

Reconnaissance

Prior to both the 1995 and 2013 traces, a thorough field reconnaissance of the study area was performed (Figure 2). Reconnaissance methods consisted of traversing all accessible stream reaches in the study area where stream

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\(^1\) Brezinski (2004a) identified active sinkholes as typically having an open throat and/or steep, unvegetated sides that suggest recent activity.

Figure 2. Map of study area and dye-monitoring locations. Note that several stream-monitoring locations from the 1995 study fall outside the figure to the southeast.
The groundwater tracing study was conducted from May to early October, 1995. Three different tracer dyes were introduced in different areas on June 7, 1995:

- 3.18 kg (7 lbs) of eosine dye were introduced into each of two Epikarstic Dye Introduction Points (EDIPs). EDIPs are borings that bottom in the epikarstic zone (approximately 7-15 m [23-50 ft] below grade) and are designed for introducing dye and water. EDIPs 1 and 2 were used for introducing the eosine dye and bracketed a former waste disposal area in the southwestern part of the study area. Using two EDIPs to bracket a relatively small waste area is an effective approach for identifying all the points to which contaminants from the waste area may flow.
- 2.27 kg (5 lbs) of fluorescein were introduced into each of EDIPs 3 and 4. These EDIPs were located in the northwestern portion of the study area and bracketed a second former waste disposal area.
- 4.54 kg (10 pounds) of rhodamine WT were introduced into a sinkhole where an intermittent surface stream terminated (called EDIP 5). It is in the west central portion of the study area, generally north of the eosine introduction locations and southwest of the fluorescein introduction locations.

Approximately 9,500 liters (2,500 gallons) of water were introduced into each of the five locations following the dye introduction. The water was slowly discharged from poly tanks temporarily located adjacent to each EDIP. A total of 139 sampling stations were monitored. Background sampling was conducted at most stations for three weeks prior to dye introductions. The background sampling identified some small emission fluorescence peaks in or near the wavelength range of fluorescein and the wavelength range of eosine (but not rhodamine WT) at a few sampling stations. In the case of fluorescein, these sampling stations were typically surface streams at points a short distance downstream of highways and parking areas. For eosine, the detections also occurred at a few surface water locations. It has been the experience of the OUL that such shoulders and small peaks are commonly attributable to algae or other aquatic plant materials. Sampling continued at most stations for 13 weeks after dye introduction; four additional weeks of sampling were conducted at selected stations.

2 Note: this stream was not flowing at the time that the tracer was introduced.
Results

Results of the 1995 study are depicted on Figure 3. Eosine was detected at 14 groundwater sampling stations: six monitoring wells and eight springs. Fluorescein was detected at two monitoring wells and six springs. Rhodamine WT was not detected at any of the sampling stations during the study period. At that time, this was attributed to adsorption of this dye onto the aquifer matrix. Relative to the other dyes introduced, a smaller mass of rhodamine WT dye was used (based on the dye equivalent) due to an expectation that water and dye introduced into a sinkhole would be more readily connected to area springs than dye introduced into EDIPs.

Based on dye concentrations measured in water samples and flow rates of the springs, approximately 90% of the eosine and fluorescein that discharged during the study is estimated to have done so from Spring A, which has several discharge points. The mean flow rate of this spring during the 1995 study period was about 13 L/s (221 gal/min). The straight-line travel distance from the nearest eosine introduction point to Spring A was 1,425 m (4,675 ft); for fluorescein this distance was 1,340 m (3,350 ft). Both eosine and fluorescein first arrived at Spring A 19 days after dye introduction. Fluorescein concentrations in water samples from Spring A continuously increased during the 99 days after the first arrival of this dye at the spring. In contrast, the concentration of eosine increased throughout the first two months after first dye arrival at Spring A and then was relatively constant for the last month of sampling.

The 1995 dye tracing study demonstrated that the area to which shallow groundwater near the former disposal areas discharges is limited to springs that discharge along an approximately 670 m (2,200 ft) reach of Carroll Creek. During the study period 0.2% of the fluorescein and 0.9% of the eosine introduced into the groundwater system discharged from the springs.

2013 Tracer Study

In the intervening years between the tracer studies a number of wells were drilled to further investigate groundwater beneath and downgradient of the former disposal areas. A number of these wells were completed...
at greater depths than wells that existed during the 1995 study. The additional drilling work revealed that, while the intensity of solution porosity (in terms of frequency and height) decreased with depth, sizeable cavities exist as deep as 96.5 to 96.8 m (316.5 to 317.5 ft) below the ground surface.

The objectives of the 2013 tracer study were to determine where the groundwater moving through this deeper solution porosity discharged and assess whether or not separate shallow and deep flow systems existed.

**Background Sampling**
Comprehensive background sampling was conducted for three weeks prior to introducing tracer dyes for the 2013 study. This sampling detected small concentrations of residual fluorescein, eosine, and rhodamine WT dyes. As discussed below, these detections are interpreted to represent dyes introduced during the 1995 tracing study.

**Eosine**
Background fluorescence in the OUL’s acceptable emission wavelength range of eosine was detected in samples from two wells and was attributed to eosine dye introduced for the 1995 tracer study (eosine dye was detected in these wells, post injection, during the 1995 study). Eosine persisted in these wells throughout the 2013 tracer study.

Eosine was also detected during background sampling at Spring B and at a location named “Pond Outfall”. Water from this latter location is derived from several springs including Spring A. These background eosine detections represent residual dye from the 1995 tracer study.

**Rhodamine WT**
Residual concentrations of weathered rhodamine WT from the 1995 study were detected in activated carbon samplers from eight springs and one well during background and routine sampling for the 2013 tracer study (Figure 3). In total, there were 98 activated carbon samplers that were positive for weathered rhodamine WT dye. The mean dye concentration in the elutants from those samplers was 2.00 µg/L and concentrations ranged from 0.482 to 5.70 µg/L. The OUL’s detection limit for rhodamine WT in carbon sampler elutants is 0.170 µg/L. No weathered rhodamine WT was detected in water samples because concentrations were below the 0.015 µg/L detection limit for this dye in water.

The OUL’s normal acceptable emission wavelength range for rhodamine WT in elutant samples is 565.2 to 571.8 nanometers (nm). Twenty-one of the elutant samples (21 percent of the samples where rhodamine WT was detected) had fluorescence peaks in this acceptable range. The mean measured emission fluorescence peak for the 98 samples was 563.9 nm and the range was from 560.0 to 567.0 nm. In the OUL’s experience, emission fluorescence peaks for rhodamine WT are often slightly shorter than the normally acceptable wavelength range at low dye concentrations and when the dye has been in the ground for periods of several months or longer. In this case the rhodamine WT had been in the ground for 18 years. The cause of this behavior is uncertain, but may be related to the physical nature of rhodamine WT. Specifically, previous studies have shown tracer-grade rhodamine WT to contain two fluorescent isomers (Hofstraat et al. 1991; Shiau et al. 1993; Sutton et al. 2001). These isomers exhibit different emission spectra and sorption kinetics. In sand column experiments, Sutton et al. (2001) found one isomer to sorb more strongly than the other.

Rhodamine WT is also subject to alteration through deamination (Kass 1998). This alteration shortens the wavelength of emission fluorescence peaks. This is an occasional problem at waste sites where emission fluorescence peaks for weathered rhodamine WT may be even shorter than those encountered in this study. The decreases in emission peak wavelengths occur both in water and in the elutant from carbon samplers, and typically increase with time.

**Tracer Introduction**
Following the background study, eosine and fluorescein dyes were introduced into deeper portions of the aquifer:

- On May 21, 2013, 6.8 kg (15 lbs) of fluorescein dye was introduced into Well X, constructed in the Frederick Formation with a screened interval of 95.4 m to 100.0 m (313 to 328 ft) bgs. The target interval was a void at 96.5 to 96.8 m (316.5 to 317.5 ft) bgs. This is viewed as a deep well.
- On May 22, 2013, 10.4 kg (23 lbs) of eosine dye was introduced into Well Y, constructed in the Frederick Formation with a screened interval from 42.7 to 47.2 m (140 to 155 ft) bgs. During well construction a fracture was noted in this well at 45.7 m (150 ft) bgs. This is viewed as an intermediate-depth well.

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4 i.e., ≥ 0.3 m (1 ft).
5 After the 1995 study, it was learned that fluorescein had been disposed along with other wastes in one of the disposal areas. As a result, we do not discuss the background detections of fluorescein because its source is uncertain.
Approximately 5,700 L (1,500 gal) of water were introduced into each of the wells following the dye introduction to flush the dye out of the well bore and into the groundwater system. The water was introduced through a tremie pipe at a rate that maintained less than 3.3 m (10 ft) of head rise in the well.

A total of 127 stations were sampled during the 2013 study. Sampling stations consisted of monitoring wells, springs, streams, ponds, and some private wells. Samples were collected at most sampling stations for over six months following dye introductions. An additional three months of sampling were conducted at selected stations in the southeast portion of the study area.

Results
Results of the 2013 study are depicted on Figure 4. Eosine, introduced into the intermediate-depth well, was detected at 17 groundwater monitoring sites; 11 wells and six springs. The first arrival times for dyes at the wells ranged from less than eight days to about 155 days. The first arrival times for eosine dye at these springs ranged from about 19 days at Spring B to about 61 days at all other springs.

The mean straight-line travel rate for the first arrival of eosine at the monitoring wells was 15.6 m/d (51.1 ft/d). The mean straight-line travel rate for the first arrival of eosine at Spring B was 51.2 m/d (168 ft/d), while it was 27.9 m/d (91.5 ft/d) to the remaining, more distant springs.

Fluorescein, introduced into the deep well was detected at five wells and Spring A. The estimated mean straight-line travel rate for the first arrival of the dye at the wells was 3.2 m/d (10.5 ft/d). The estimated mean straight-line travel rate for first arrival of fluorescein at Spring A was 9.3 m/day (30.4 ft/d).

During the 2013 study period 0.01% of the fluorescein and 0.1% of the eosine introduced into the groundwater system discharged from the associated springs.

Discussion
The karst literature is replete with information on tracer studies conducted in karst aquifers that fall near the “conduit” end member of the conduit flow-diffuse flow continuum. These aquifers typically have near-ideal locations to inject tracers directly into the conduit porosity, and a significant proportion of the tracer mass moves relatively rapidly through the aquifer, generating what can be considered “classic” breakthrough curves at springs where dye discharges. Such curves exhibit a clear leading edge, peak concentration, and trailing edge of the tracer, and the travel times to each are reasonable in the context of field tracer studies – that is, the time between

Figure 4. Results of 2013 tracer study.
the leading edge and trailing edge of tracer is on the order of days to perhaps several months.

Less-well documented are tracer studies conducted in aquifers that exhibit an appreciable component of diffuse flow. In these aquifers, flow pathways may be poorly integrated or developed, more convoluted, and/or more occluded with clastic sediment. In our experience, such conditions are not uncommon in the folded and faulted carbonates of the Piedmont and Valley and Ridge Provinces. In such aquifers, tracer may remain in the aquifer for many years (Quinlan et al. 1990; Quinlan and Ray 1991; Quinlan and Ray 1992). As a result, tracer travel times can be prolonged, resulting in extended times between the leading edge and trailing edge of tracer and a significant flattening of breakthrough curves.

Figure 5 shows the breakthrough curves for fluorescein and eosine at Spring A during four months of the 1995 tracing study. The curves are based on three composite water samples per day with each composite containing three individual samples. Based on fourteen samples and their duplicates, the mean relative percent difference (RPD) values were 2.8% for fluorescein and 4.0% for eosine. Variations in dye concentrations from one sample to another are substantially greater than the RPD values; the cause of these variations is unknown. As can be seen in Figure 5, the time to reach the trailing edge of tracer, particularly for fluorescein which does not appear to have reached its peak concentration, could be on the order of several months to years.

The data collected during the two tracer studies described in this paper demonstrate that this aquifer contains an appreciable amount of stored water and exhibits a significant component of diffuse flow. Furthermore, these data lead to several useful findings.

Utility of Activated Charcoal Samplers
Activated charcoal samplers have proven their utility in tracing studies for many years. Their ability to accumulate tracer dyes allows detection of dyes where their concentration in water may be below detection limits in grab samples of the water. Our data show that, in aquifers with a significant component of diffuse flow, their use can significantly enhance the outcome of a tracer study. For the 1995 trace, 30 wells were sampled. Eosine was detected in carbon samplers from five of them and also in water samples from two of the five. Fluorescein was detected in carbon samplers from two of them and also in water samples from one of the two. A total of 10 springs were sampled. Eosine was detected in carbon samplers from eight of them and also in water samples from six of the eight. Fluorescein was detected in carbon samplers from six of them and also in water samples from four of the six.

Figure 5. Breakthrough curves for fluorescein and eosine in water samples from Spring A – June 7 (introduced) to September 8, 1995. These dyes were first detected in carbon samplers deployed from June 23 to 29 – estimated first arrival = June 26 (Day 19).
For the 2013 tracing 32 wells were sampled. Eosine was detected in carbon samplers from 11 of them and also in water samples from 5 of the 11. Fluorescein was detected in carbon samplers from five wells and also in water samples from four of the five. A total of nine springs were sampled. Eosine was detected in carbon samplers from six of these and also in water samples from two of the six. Fluorescein was detected in carbon samplers from two of the nine springs. Water samples from the same two springs also contained fluorescein.

Based on the above data it is clear that many dye-detection locations would have been missed if activated carbon samplers had not been used.

**Dye-Residency Time and Aquifer Storage Capacity**

Results of this study demonstrate that in this aquifer, tracer dyes were largely sequestered and remained at detectable concentrations for at least 18 years. Remnants of eosine and rhodamine WT dyes from the 1995 tracing were found during background sampling prior to dye introductions for the 2013 tracing.

Small concentrations of rhodamine WT were detected during 2013 in carbon samplers from eight of the nine sampled springs and in carbon samplers from one of the monitoring wells. None of the water samples from these stations contained detectable concentrations of rhodamine WT. This finding further demonstrates the utility of the activated carbon samplers used in this study. The percentages of introduced dyes masses discharged from springs draining the aquifer during each of the two study periods\(^6\) (Table 1).

The very small percentages of introduced dyes detected in associated springs, and the presence of remnant dyes in the aquifer 18 years after their introduction, is consistent with the conclusion that much of the introduced masses of tracer dyes are retained within the aquifer and have not discharged from the associated springs; that is, the storage capacity of this aquifer for organic tracer dyes, and by extension organic contaminants, is large. It is reasonable to conclude that an appreciable portion of contaminants disposed within the study area and not removed during remediation are also detained within the aquifer.

**Aquifer Structure and Groundwater Movement**

The findings of the tracer studies, when combined with boring data from wells drilled in the study area, provide insight into the nature of solution porosity in the aquifer and its role in groundwater movement. There is a well-developed epikarst beneath the soil and, as expected, solution porosity decreases with depth (Figure 6). Despite this decrease, solutional cavities have been discovered as deep as 105.3 to 105.45 m (345.5 to 346 ft) below ground surface in the Frederick Formation. In the New Oxford Conglomerate, the deepest cavity penetrated spanned from 82.6 to 82.9 m (271 to 272 ft). Such deep porosity can be caused by meteoric waters. The steeply dipping beds of limestone of the Rocky Springs Station Member, interspersed with occasional beds of insoluble shale, can serve to focus aggressive water downward, particularly along bedding-plane fractures among beds of purer, more soluble limestone. Near-vertical joints or faults serve to link such penetrated bedding planes creating phreatic loops (Ford 1971; Ford and Ewers 1978).

It is clear from the tracer studies that the aquifer consists of one flow system with variable travel rates. In a similar setting in the Shenandoah Valley, located about 65 km (40 mi) to the west of the study area, Doctor et al. (2011) found that the carbonate aquifer beneath the valley was karstified to depths exceeding 300 m (984 ft) and that springs in the region integrated flow through both shallow and deep flow paths.

An interesting observation from this study is that flow in the aquifer is not only integrated vertically within the Frederick Limestone; but also is integrated between the Frederick Limestone and the New Oxford Conglomerate. Dyes introduced in the Frederick Limestone discharged mainly from springs issuing from the New Oxford Conglomerate.

The estimated first-arrival groundwater velocities throughout the aquifer are appreciably slower than those typical of many karst aquifers, and the dyes dispersed more widely in the aquifer than is common in conduit-dominated karst aquifers. These facts indicate that conduit flow is subdued in this aquifer. Anecdotal observations from long-time residents of the area indicate that the flow of larger springs in the area is remarkably constant. Flows are not observed to increase dramatically in response to storms, and spring

<table>
<thead>
<tr>
<th>Dye Location</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eosine in epikarst zone (1995)</td>
<td>0.9%</td>
</tr>
<tr>
<td>Fluorescein in epikarst zone (1995)</td>
<td>0.2%</td>
</tr>
<tr>
<td>Eosine in intermediate depth zone (2013)</td>
<td>0.1%</td>
</tr>
<tr>
<td>Fluorescein in deep zone (2013)</td>
<td>0.01%</td>
</tr>
</tbody>
</table>

**Table 1.** Percentages of introduced dyes masses discharged from springs draining the aquifer during each of the two study periods.
water remains visibly turbid-free after such events. This suggests that aquifer recharge is primarily diffuse, and that flow velocities are too low to transport appreciable volumes of clastic sediments to springs.

The model for aquifer structure and groundwater flow described above also offers an explanation for the behavior of the rhodamine WT dye that was injected into the sinkhole in 1995. Since the rhodamine WT did not discharge during the 1995 study period, it is reasonable to conclude that the dye was detained in the aquifer and then slowly discharged over an extended period. It is clear that the flow paths followed by the rhodamine WT were longer, more complex, and less efficient than originally anticipated, and groundwater velocities were lower than anticipated.

The primary area of groundwater discharge from the karst aquifer is at one or more of the springs that discharge along an approximately 670 m (2,200 ft) reach of Carroll Creek. This is the case regardless of the depth at which 7 A small amount of eosine from the 1995 tracing was detected at Hospital Spring, located about 230 m (750 ft) east-southeast of Spring A.

Figure 6. Depth distribution of solution-enlarged features observed in video logs. Percentages have been normalized to the footage of borehole logged in each depth interval. Data set consists of 30 solution features identified in 462 meters of video logging conducted at 17 boreholes.

Tracer dyes were introduced. The 1995 tracer study included 139 sampling stations and sampling lasted for 13 weeks after dye introduction, and for 17 weeks at some selected locations. The 2013 tracer study included 127 sampling stations and lasted over six months and for an additional three months at selected locations. Both studies included many sampling stations at streams, springs, and wells located hundreds to several thousands of meters from the disposal areas. There are no data suggesting any significant discharge of dyed groundwater other than from the springs identified in this paper.

Conclusions
Tracing studies conducted 18 years apart elucidated aquifer structure and groundwater movement in the karst aquifer beneath the study area. Both the Cambrian Frederick Formation and the Triassic New Oxford formation have been karstified and act as one flow system to a depth of at least 100 m (328 ft). Flow velocities are sluggish by classic karst standards, attributed to flow pathways that are long, convoluted, and potentially occluded in places by clastic sediment. Diffuse flow, rather than conduit flow, plays a large role in the aquifer. As such, dye- and contaminant-residency times are long, on the order of decades. Eosine, rhodamine WT, and possibly fluorescein remained detectable in this aquifer 18 years after they were introduced.

Tracer studies conducted in this, and similar, aquifers may benefit from a longer duration, and greater introduced dye masses, than in aquifers with a greater component of conduit flow. The concentration of dyes discharged at springs can be expected to be relatively low; increasing the utility of activated carbon samplers. Had only grab samples of water been used in these two studies, many dye detections would have been missed.

References
Brezinski DK. 2004b. Geology of the Frederick Quadrangle, Frederick County, Maryland. Maryland Geological Survey Geologic Map, scale 1:24,000.
Abstract
During the 1990s, in an attempt to better understand threats posed by surface developments overlying the cave, National Park Service staff at Wind Cave National Park in Custer County, South Dakota carried out a series of dye traces through portions of the vadose zone overlying the cave. Wind Cave is located within the 100 m-thick Madison formation (limestone and dolomite), which in most locations is capped by varying thicknesses of the basal units of the Minnelusa formation (intermingled beds of sandstone, limestone, and shale). A variety of cave locations with dripping or pooled water were monitored for up to five years following dye injection. Transit times to the cave varied from less than six hours to as much as 4.8 years. Despite a variety of positive results, there appears to be little correlation between transit time and lateral or vertical distance from the injection site. Data analysis produced traditional-shaped dye recovery curves in some locations, albeit stretched out over hundreds and possibly even thousands of days beyond dye injection. The results strongly suggest that chemical or sewage spills in the vicinity of the dye injection sites would quickly enter multiple sites in the cave system, and could persist for years.

Introduction
Wind Cave is located within Wind Cave National Park in the southern Black Hills in Custer County, South Dakota. The cave is an extensive three-dimensional labyrinth that as of this writing consists of more than 230 kilometers of surveyed passages contained within an area of roughly 2.5 square kilometers (Ohms, 2015, personal communication). It is contained within the Madison formation, which consists of beds of limestone and dolomite of Mississippian age roughly 100 meters in thickness in the Wind Cave area (Palmer and Palmer, 2008).

Overlying the Madison formation in the vicinity of the cave is the Minnelusa formation, consisting of sandstone, limestone, dolomite, and shale (Strobel and others, 1999). The vast majority of Wind Cave is capped by the lower 30 to 100 meters of the Minnelusa formation, but a small number of cave passages are located beneath areas where the overlying Minnelusa formation has been largely or entirely eroded away. The only known completely-natural entrance to the cave is located in such an area, in a drainage known as Wind Cave Canyon (Figure 1).

The contact between the Minnelusa and Madison formations is complex. The Madison was exposed soon after deposition, and a karst topography formed, with many caves, sinkholes, and other typical karst features. As a result, the base of the Minnelusa was deposited atop an irregular surface, and in the process it filled many former solutional features up to tens of meters below the highest remaining reaches of the Madison (Palmer and Palmer, 2008).

The area above Wind Cave is semi-arid with a mean-annual precipitation of 45.8 cm/year (Ohms, 2012). Streambeds above the cave area have highly intermittent flow, and there are few locations within them that would appear to capture what flow they occasionally have in any obvious way.

Wind Cave National Park was established in 1903, making it one of the oldest national parks in the United States. For the convenience of the visiting public, the park headquarters and its associated buildings, utilities, and parking lot were established near the cave’s en-
trance. During the 1980’s and 1990’s, park staff became increasingly concerned about the aging infrastructure lying above the cave. Water and sewer lines were known to be compromised, and untreated runoff from the 1-hectare asphalt (at that time) parking lot was entirely disappearing within 15 meters of the drains, which are located in Wind Cave Canyon (Nepstad, 1997). How much of the cave was affected by these potential contaminating sources? How long did it take to reach the cave? Once there, how long was the contamination present?

To answer these questions, it was necessary to trace water from the surface to the cave. Although dye tracing with fluorescent dyes has been commonly employed for this purpose for decades, the circumstances at Wind Cave presented challenges. The surface overlying the cave lacks traditional karst features such as sinking streams or sinkholes, requiring ingenuity with respect to the selection of dye injection sites and dye injection methods. Rather than injecting dye into a sinking stream and testing for it at resurgences or within cave streams, it would be necessary to trace diffuse flow through the vadose zone directly above the cave.

**Previous Work**

In 1986, Marsha Davis, a graduate student at the University of Minnesota working under E. Calvin Alexander, conducted two dye traces at Wind Cave (Alexander and others, 1989). The objective of the traces was to determine whether links existed between the park’s visitor center parking lot and wet areas in the underlying cave. On July 19, 1986, 1122 grams of fluorescein (Acid Yellow 73) was injected into a small swallow in the middle of Wind Cave Canyon in the park’s picnic ground, along with roughly 1,000 liters of water. On August 16, 1986, 1076 grams of rhodamine WT (Acid Red 388, often referred to generically as rhodamine throughout this paper) was injected with several thousand liters of water beneath the south parking lot drain - located near the elevator building.

At the time of these traces, there was no database documenting the location of wet areas within Wind Cave. Park staff and volunteer cavers, acting on personal recollections, were able to inform Davis of a limited number of sites in the vicinity of the injection points with known dripping or pooled water. Frustrated at not being able to provide researchers with more information for such a critical study, in 1989 park staff instituted an aggressive inventory of cave features at Wind Cave. By the early 1990’s, inventory data was already showing that Davis’ work had missed several key wet areas.

Despite the fact that sampling was unintentionally incomplete, Davis reported some positive results for both traces. Fluorescein was reported at survey stations NFP5 (Fairy Palace) and NM4 (Before Fairy Palace). Possible fluorescein was reported at RS11 (Minnehaha Falls). Following some background sampling in these areas in 1993-1995, park staff learned that Fairy Palace and Before Fairy Palace consistently produce false positives for fluorescein. Minnehaha Falls still had small traces of fluorescein present at that time, but due to its location beneath the south parking lot drain, park staff suspected that the fluorescein present there was from antifreeze spills on the parking lot. All of this, combined with the scarcity of sampling sites and the short duration of sampling (only about three weeks), created some skepticism regarding the reported fluorescein results.

Davis’ rhodamine WT trace also produced some positive results. One site, RS11 (Minnehaha Falls), produced dye within two days of injection. A few other sites also eventually tested positive for dye including RS13 (Mule’s Ears), UB10 (Caving Tour Drips), and C58 (Garden of Eden Drips). Dye may have shown up at additional sites, but once again monitoring was limited to a small number of sites in the cave and was cut off relatively early.

A year following the injection of the rhodamine, dye began showing up in the park’s well, located within Wind Cave Canyon about 2.7 km east of the rhodamine injection site off the parking lot. Davis attributed the dye in the well to a rhodamine WT injection two months earlier into the Beaver Creek sink, located about 3.8 km northwest of the well, but since the same dye had been used for both traces, an additional trace would be necessary to produce conclusive results.

By 1992, the number of cave management staff at Wind Cave National Park had grown sufficiently that it was possible to contemplate additional dye traces. Park staff, advised by Joe Meiman, a hydrologist who worked for Mammoth Cave National Park at the time, began to experiment with alternative methods of tracing water through the overlying vadose zone. The persistence of Davis’ dyes in the vicinity of her traces led them to experiment with Lycopodium spores. Given the limited number of fluorescent dyes available, it was hoped that if such a trace could be successfully accomplished, a nearly unlimited number of tracing elements would be available, since the spores could, in theory, be dyed with a large number of colors.
An experimental trace was attempted in 1992, when roughly 200 grams of Lycopodium spores, together with approximately 200 gallons of water, were injected in the same small swallow that Davis had injected her fluorescein into back in 1986. Sampling stations containing 10 micron filter paper were placed beneath several cave drips that were known to respond quickly to rain events on the surface, including some that Davis had believed tested positively for dye during the 1986 trace. The filters were removed from the cave on an occasional basis and examined carefully with a microscope. No spores were ever detected, so the Lycopodium experiment was largely a failure, probably due to the filtering effect of overlying soil and sediments in the epikarst.

By coincidence, however, park staff had been background testing the water at many of these same sites for the presence of optical brightener, in the hopes of using that dye as a tracer element in the future. Untreated cotton had been left in several of these locations, and these samples were observed under a handheld UV lamp after being removed from the cave. During the Lycopodium experiment, a cotton sample from NM4 (Before Fairy Palace) was found to be strongly positive for optical brightener. Park staff were puzzled by this occurrence, since all known potential sources of optical brightener (leaking sewage lines) were located down-dip from this location. Eventually it was recalled that the Lycopodium spores had been prepared for injection by mixing them with four liters of water mixed with laundry detergent (to prevent clumping). Although the Lycopodium experiment proved to be a failure, it is interesting that it indirectly led to a successful dye trace of sorts.

By 1993, it appeared as though future traces would have to resort to utilizing the limited number of fluorescent tracer elements available at that time. A series of three traces were planned for that year utilizing fluorescein, rhodamine WT, and optical brightener. All of the traces were conducted considerably to the west of the 1986 traces. Over 60 sampling locations were set up, blanketing nearly all of the known portions of the cave. Small samples of untreated cotton and activated charcoal “bugs,” consisted of a few grams of activated charcoal contained within a small screen pouch, were tested on at least a quarterly basis for a full year prior to dye injection in order to gather background data.

Once they were removed from the cave, the charcoal bugs were eluted with roughly 10cc of an elutant consisting of 1-propanol, de-ionized water, and ammonium hydroxide (mixed at 5:3:2), then tested with a Turner Model 111 fluorometer with the filters appropriate for the spectra of the dyes to be used. Cotton samples were visually examined for optical brightener exposure with a handheld UV lamp.

With the exception of the single cotton sample that tested positive for optical brightener described in the Lycopodium experiment and a few weak positives for fluorescein or rhodamine in the vicinity of the 1986 work, the entire cave proved to be free of the presence of all three dyes.

On June 14, 1993, 1000 grams of a 20% rhodamine WT solution was injected into the bottom of a small sink in a drainage roughly overlying the Club Room portion of the cave, along with roughly 350 liters of water from a pumper truck. The water in the truck had been left exposed to the atmosphere for four days to allow any residual chlorine from the park’s water treatment system to dissipate. Roughly 125 liters went in prior to the dye, with the remaining 225 liters going in after the dye had been poured into the sink.

On July 6, 1993, 1008 grams of fluorescein powder mixed with 10 liters of water was injected into a small depression above the southeast portion of the cave, along with roughly 1200 liters of chlorine-free water from a pumper truck. About a third of that water was injected prior to the dye, while the remainder was used to flush it into the system. Later that same day, 4000 grams of Tinopal CBS-X (Fluorescent Brightener 351), an optical brightener, was injected along with roughly 25000 liters of chlorine-free water into a small depression in a drainage overlying the southwestern portion of the cave.

During the 1986 traces in Wind Cave Canyon, dyes were injected either directly into the cave-bearing Madison formation, or into the lowest five to ten meters of the overlying Minnelusa formation. The dyes used in the 1993 traces were injected considerably further up into the Minnelusa “cap” above the cave. The rhodamine trace above the Club Room was roughly 25 meters further up in the Minnelusa, while the fluorescein and tinopal traces were injected 50-65 meters up into it.

No longer dependent on visiting researchers for sampling, these traces were sampled by park staff according to a far more rigorous schedule, and for a much longer period of time. Some sites were monitored on a daily or weekly basis for months, and the remainder of the sites were monitored at least quarterly. After two full years of monitoring and over a thousand samples being collected, no dyes were detected in the cave from any of the 1993 injections.
While the 1993 traces appeared to be failures at first glance, they nonetheless demonstrated a critically important principle. Unlike dyes injected at the very base of the overlying Minnelusa formation, which could arrive in the cave in a matter of days, dyes injected further up into the Minnelusa were either carried away laterally from the cave area, or diffused or delayed by the Minnelusa cap to such an extent as to render them undetectable for at least two years.

The 1993 traces taught park staff that the majority of its developments - its parking lots, buildings, and sewer and water lines - were perched above the most hydrologically vulnerable portion of the cave, where the protective Minnelusa formation was thinnest. It was time to supplement the results of the 1986 traces to get a better idea of the extent of this problem.

The 1996 Traces
In early 1996, additional background sampling demonstrated that dye from the 1986 traces was either entirely absent or barely detectable at cave sites.

Utilizing the park’s extensive cave inventory database, park staff were able to identify over 60 locations suitable for sampling in areas including and surrounding Davis’ earlier sampling locations. Following a final round of background sampling in the spring of 1996 (including both grab samples and bugs), plans were made to once again inject the same dyes into the same locations utilized in 1986 (see Figure 2) in an attempt to better quantify the extent of the cave influenced by these surface locations, and to better measure travel times and dye persistence.

On July 29, 1996, 2142 grams (1800 ml) of a 20% rhodamine WT solution, representing 428 grams of pure dye, was injected beneath the south parking lot drain, along with enough water to simulate a one-inch rain event - roughly 140000 liters for this particular drain. About a third of that water was injected prior to the dye, while the remainder was used to flush it into the system. Unfortunately, due to the large quantity of water needed to accomplish this, there was no way to ensure the water was completely chlorine-free. Chlorine levels in the water were very low, however, as chlorination had been shut down for over 24 hours prior to opening the hydrant used to supply the water. As chlorine levels in the water are ordinarily very low to begin with, it likely destroyed very little of the dye.

Later that same day, 2339 grams of fluorescein powder mixed with about 20 liters of water was injected into the same small swallow near the park picnic ground that was used in 1986. As in the earlier experiment, roughly 1,000 liters of chlorine-free water from a pumper truck was used to inject the dye - about 350 liters before injection, and about 650 liters following it.

As in the 1993 traces, samples were collected in 40ml glass vials that were specifically labeled for each site. Due to a limited budget and the very large number of samples taken over the course of these traces, vials were reused. To avoid contamination once a site tested positive for dye, vials were rinsed with de-ionized water at least three times between uses, and droppers used to fill vials in the cave (emptied and rinsed prior to each use) were kept at each sampling site to avoid cross-contamination. Activated charcoal bugs were placed at each site as a means of confirming very weak positives, or detecting dye arriving at quantities too low to measure in grab samples.

Water samples from the first 444 days were analyzed with a Turner Model 111 fluorometer, and subsequent samples were analyzed with a newly-acquired Turner TD-700. For three weeks, both fluorometers were used to analyze samples to ensure consistency of results between instruments.

The Rhodamine WT Trace
It had been known for years that a site in the cave near survey station BI25 (Upper Minnehaha Falls), located almost directly below the parking lot drain, consistently

![Figure 2. The locations of the 1986 and 1996 dye injection points on the surface above Wind Cave. The same dyes were injected in the same locations utilizing the same methods for both traces.](image)
responded to rain events within a matter hours. It was believed that the dye would make its initial appearance in the cave here, so an auto-sampling device was installed to automatically collect samples on an hourly basis following the rhodamine injection. Unfortunately it missed the first five hours following injection.

The first sample collected at BI25 was more than double the barely detectable pre-injection background level of 0.07 parts per billion (ppb), and one hour later it was nearing 1 ppb. By midnight that evening, it was approaching 20 ppb. Already a site had been discovered that received dye several times faster than the site that first went positive in 1986 (RS11, Minnehaha Falls).

BI25 would go on to become the most carefully documented site during the life of this dye trace. The site is easy to access, and was sampled on an hourly basis for weeks, on a nearly-daily basis for over a year, and several times per week following the first year, until the author left the park in October 1998. The rhodamine results from this site are very interesting. The first two years are illustrated in Figure 3.

A plastic tarp was arranged beneath the site to capture most of the flow to allow for the measurement of flow rates, which proved to be highly variable. Following heavy rains during the spring and summer, steadily streaming or heavily dripping water will pour in from a dome above the site at up to 1700 ml per minute. Flow rates quickly reach a peak within a day of the rain event, and just as quickly fall off in the days immediately following. The steep peaks in the drip rate in Figure 3 all relate to heavy rain or melting events on the surface. During cold, dry stretches in the winter, the drip rate slows dramatically, and on some occasions even ceases completely.

Rhodamine levels rose steadily at the site for a week, rising to 87 ppb before declining to 43 ppb 11 days after injection. It appeared as though the peak dye concentration had come and gone, and the site would continue to rapidly decline over the coming weeks. On the twelfth day after the rhodamine was injected, the first significant rain event on the surface occurred. As expected, drip rates quickly rose, but dye levels surged even faster, peaking at 250 ppb two days later before starting yet another decline. Another rain event occurred on the seventeenth day, creating another rapid surge in dye levels that peaked at 380 ppb within 48 hours. This reading on the nineteenth day proved to be the true peak at this site. Although spikes in dye concentration continued to occur

Figure 3. The results of the first two years of sampling at BI25 (Upper Minnehaha Falls). This was the most carefully studied site in the 1996 traces.
immediately following subsequent precipitation or melting events on the surface, the magnitude of the spikes gradually decreased in the months following injection. They become barely noticeable after about 300 days. Despite the spikes which occurred early in the trace, the overall shape of the concentration curve strongly resembles a traditional “positively skewed, bell-shaped curve that is steeper on the rising limb than on the falling limb,” (Mull, et al, 1988), except that instead of a time scale measured in hours or days, the time scale here stretches over multiple years.

The persistence of the dye in the system was rather remarkable. Nearly 1200 days following the injection, dye concentrations were still double the background levels measured prior to the 1996 dye injection. This persistence perfectly illustrates the threat posed by having the parking lot perched above the cave in this area. If contaminants in parking lot runoff behave like the rhodamine used in this trace, then relatively minor spills on the parking lot could impact cave resources for years following an incident.

BI25 was an interesting site, and it received more attention than any other sampling location during the 1996 rhodamine trace. But it was by no means the only site in the cave where positive results were found. Table 1 documents all of the sites that received measurable quantities of rhodamine, together with the number of days it took for the leading edge of the dye to reach those sites and the number of days it took dye concentrations to reach their peak.

There appears to be no clear correlation between the emergence of dye at a site, its peak concentration, or the time it took to peak and any of the other variables associated with a site. Some sites at a greater horizontal distance away from the injection point went positive well before other sites that were nearly beneath it. Likewise, vertical depth is a poor indicator of when dye will arrive or at what concentration it will peak, as is the drip rate associated with a site. Rather, travel times and peak concentrations appear to depend on highly local but unmeasurable variables, such as the site’s proximity to hidden flow networks in the epikarst, the presence or absence of clay-bearing paleofill deposits in the overlying strata (which could prevent or slow more direct flow to the site), or the mixing characteristics of other waters encountered in transit.

The concentration curves for many of these other sites also resemble the classic shape of curves associated with traditional dye traces in karst aquifers. Other sites exhibit concentration curves that are anything but traditional. See Figure 4 for examples of both. Concentrations at C47 (Assembly Room), for instance, rise abruptly once dye arrives, but remain relatively flat over the course of an entire year before increasing once again. One cannot even be certain the peak concentration had been measured after nearly 1000 days, when sampling finally ceased. Likewise, we cannot be certain that the peak concentration was measured at either C53 (Escape Route) or C53A (Silent Lake) after more than two years of sampling. Years of additional sampling may have been necessary in order to capture an entire curve at these sites.

Rhodamine Recovery
Accounting for the dye introduced in a vadose zone dye trace is challenging. In a traditional dye trace in a karst...

<table>
<thead>
<tr>
<th>Site</th>
<th>Depth (m)</th>
<th>Horiz (m)</th>
<th>Days to Cave</th>
<th>Peak (ppb)</th>
<th>Days to Peak</th>
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<td>31</td>
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Table 1. Cave sites that tested positively for rhodamine WT following the July 1996 injection. Also listed are the vertical and horizontal distances from the injection point, days following injection that dye was first detected, the peak concentration measured, and the number of days following injection that the peak concentration occurred. Missing data for RS10, BI12, and UYA4 are due to very infrequent sampling at those sites.
landscape, where dye is poured into a stream at an insur- 
gence, the researcher can at least be relatively certain that 
a large proportion of the dye injected got into the system. 
Although park staff could be certain about the amount of 
dye poured onto the ground in these Wind Cave dye trac- 
es, it is impossible to know how much of that dye might 
have been adsorbed by soils and sediments at or near 
the injection point. Water and dye pooled over an area 
of roughly 150-200 square meters before gradually sink- 
ing, so the injection was not perfectly discrete. Although 
this was deemed acceptable, since the experiment was 
designed to model the fate of parking lot runoff and that 
was how parking lot runoff in the area behaved, it non- 
etheless introduces some complexity that is difficult to 
completely resolve. We can calculate roughly how much 
dye reached the various sites in the cave, but the fate of 
the dye beyond the sum total of all sites will remain a 
subject of speculation.

The vast majority of the dye that we can account for 
traveled through BI25. This site experienced not only 
the highest concentrations, but the highest flow rates as 
well. Over the course of the first two years following in-
jection, flow rates averaged 227 ml/min at the site, and at 
times exceeded 1700 ml/min. Other sites had far lower 
concentrations, or far lower flow rates, or both.

To arrive at an estimate for the quantity of dye captured 
at BI25, certain assumptions had to be made. It was 
assumed that each day’s observed flow rate could be
viewed as the mean flow rate for that day, and that each sample’s dye concentration could be viewed as the mean concentration for that day. Likewise, if more than a day passed between samples, then the values for the most recent sample were used for the day(s) not sampled. While this inevitably resulted in over-calculating the daily mass flux on days when concentrations and/or flow rates were declining, the opposite surely happened on numerous occasions as well. It is assumed that the errors thus introduced into the estimate will cancel each other out, since the number of samples collected was quite large. Data was collected on 538 of the first 752 days, including 344 days during the critical first year following injection.

If this assumption is correct, then roughly 12.06 grams of rhodamine WT (dye, not dye solution) traveled through BI25 during the course of the first 752 days. Figure 5 illustrates the temporal distribution of the dye recovery, with the solid red line representing the running total of rhodamine received at the site by a given day. Roughly 9.1 grams arrived during the first summer and fall following injection. The graph flattens significantly during the course of the winter of 1996-1997, as flow rates at the site markedly slowed. The total stood at about 9.9 grams at the beginning of the spring of 1997, and rose to 11.9 grams by the beginning of the winter of 1997-1998. During the spring, summer, and early fall of 1998, the total slowly rose to 12.06 grams. Although dye was still arriving at low concentrations at that time, extrapolating on this curve, it is difficult to imagine the total ever rising above 12.5 grams.

The sampling site at BI9 is beneath and just down-dip from BI25. Due to its location, and due to the fact that it tracked so closely with data from BI25 (they peaked at precisely the same time), it is strongly suspected that the water emerging at BI9 is water that has already flowed through BI25. Although it peaked at a lower concentration, this could be due to mixing that occurs after the water leaves BI25. Unless future evidence indicates otherwise, it is probably not appropriate to consider this site when accounting for dye recovery.

Figure 5. Cumulative totals for dyes recovered at BI25 by a certain period in time. Note the rhodamine WT curve flattening significantly during the first year, compared to the fluorescein curve, which does not. Graph represents total dye recovered, not total dye mixture recovered.
Another site where very high concentrations of dye were encountered was RS11, Minnehaha Falls. Although this site peaked at 235 ppb about a month after injection, the flow rates here are much slower than those encountered at BI25 or BI9. Average measured flow was 4.9 ml/minute at RS11, compared to an average 226.9 ml/min at BI25. Site conditions at RS11 prevented all flow entering the site to be measured, however. It is estimated that the total amount of water entering the site at any time is roughly double the measured flow.

A total of 309 samples from RS11 were collected and analyzed during the first 760 days following injection. As with BI25, if we assume that daily observed flow rates and dye concentrations are the mean values for that particular day, then a total of 0.18 grams of rhodamine passed through RS11 during the first 760 days of this trace. If our assumption is correct that twice as much water enters the site as a whole, and if that additional water had the same concentration as the water that was analyzed (which is likely, due to its proximity), then it is possible that a total of up to 0.36 grams of rhodamine WT entered the cave at RS11 during this time period.

The remaining sites were sampled far less frequently, and had significantly lower dye concentrations. Due to large gaps between sample collections, the author believes that recovery estimates are not precise enough to cite, but even considering simple order of magnitude basis estimates, most of these sites can account for a few hundredths of a gram total at best, with the possible exception of SA1, which may be able to account for a few tenths of a gram.

In all, it is fairly safe to say that less than 13 grams of dye can be accounted for from the rhodamine WT trace, a little over 3% of the amount injected. One possible explanation for the remaining dye was that it adsorbed to soil, sediments, or bedrock above the cave. It is also possible that the dye passed into a series of fractures unconnected to the cave, or too small for humans to enter, and thus passed to the water table unobserved. Additionally, it is possible that the dye was intercepted by lateral flow prior to entering the Madison formation and was carried away from the sampling area, and possibly even the cave area. Finally, it is possible that the rest of the dye traveled through unexplored cave passages. This last possibility is the least likely. The sample area is located in close proximity to the elevator entrance, and as a result is one of the most thoroughly explored portions of the cave.

More than likely, the fate of the missing rhodamine was a combination of the other three possibilities. Rhodamine WT is known to adsorb to soils and sediments (Smart, 1977), and as discussed previously, the injection did not occur at a discrete point. Also, the Madison formation is highly fractured in the cave area, as evidenced by the incredibly complex nature of Wind Cave itself. It is not difficult to imagine a myriad of fractures that could have diverted a significant portion of the flow. And finally, intriguing evidence of lateral flow beneath Wind Cave Canyon would come as results from the fluorescein dye trace began to emerge.

The Fluorescein Trace

Following the injection of the fluorescein on July 29, 1996, a total of 25 sites were repeatedly sampled in the northern portion of the cave. These sites included NFP5, which Davis believed tested positive in her 1986 fluorescein trace, and NM4, which Davis also believed tested positive in her trace and which definitely tested positive for optical brightener following the Lycopodium experiment. Given this past history of the area, it was assumed that a number of sites would receive measurable quantities of dye in this trace.

After 15 months of exhaustive sampling, no sign of the fluorescein was ever discovered in the northern portion of the cave. This came as quite a surprise, since the injection point is very near the Minnelusa/Madison contact. Due to that proximity, it was believed that the dye would travel in a mostly vertical fashion, with little if any lateral movement. While some of it may have traveled vertically via fractures too small for humans to enter, in this case we discovered direct evidence of lateral flow beneath Wind Cave Canyon.

Despite our early assumptions, background sampling was rigorously performed prior to dye injection. Background grab samples and bugs were tested for all dyes that were going to be used, regardless of how unlikely we believed a particular dye might be encountered at a particular site in the future. Between the background sampling performed in June 1996 and routine sampling that occurred during the course of the 1993 dye traces, we knew that fluorescein levels at BI25 and RS11 tended to test a little higher for fluorescein than other cave sites, but always hovered at levels equivalent to 10 - 14 ppb (Figure 6).

In the days following the dye injection, fluorescein levels at BI25 and RS11 remained at familiar levels. Lulled into complacency by the initial readings and desperate to save time due to the crush of data coming in (we collected nearly 1000 samples during the first month), we briefly stopped measuring for fluorescein at these sites and others that were distant from the injection point.
This proved to be a major mistake, as we missed the fluorescein’s arrival at these sites.

Thus, no fluorescein data for BI25 and RS11 exist between day 19, when it was still within the background range, and day 78, when we resumed fluorescein testing for these sites. By that time, fluorescein levels had nearly doubled at both sites. Levels slowly rose during the course of the fall and winter, and eventually peaked at 40 ppb at BI25 and 38 ppb at RS11 the following spring.

The fluorescein dye curves for these sites were unlike anything we had seen. For more than two years, concentrations fluctuated on what appears to be an annual cycle. Levels were highest during the winter months, as flow rates in the cave slowed, then dropped in the spring as the flow rates accelerated. One possible explanation for this would be that relatively local flow, such as the flow carrying the rhodamine into the cave from the parking lot drain area, dilutes the lateral flow that is presumably carrying the fluorescein down from the north. In other words, the flow we observe at BI25 is a mixture of two separate components, one very local, another with a more distant origin. Such unexpected results cried out for verification. Can we be certain it was indeed fluorescein that was being measured? If so, could it possibly be from an unreported antifreeze leak on the parking lot that exited the drain at the rhodamine injection point?

The answer to the first question arrived in fairly short order. After elution, charcoal bugs retrieved from the site displayed fluorescent green that, especially when illuminated with a bright beam of light, was visible even through the pink cloud produced by the rhodamine at the sites. Later, we discovered that the same could be done with raw samples of water from the site. Joe Meiman at Mammoth Cave provided a more definitive verification via a spectrofluorophotometer. Fluorescein was clearly entering the cave.

The antifreeze spill theory was dismissed after a discussion with an employee from a major antifreeze company, who, after a bit of arm twisting, revealed the rough concentration of fluorescein present in antifreeze. Suffice it to say that it would take at least 1000 liters of raw antifreeze to produce the amount of dye recovered at BI25 over the first two years of sampling. Most of the fluorescein we were encountering simply could not

Figure 6. Rising from background levels equivalent to 10-14 ppb, fluorescein settled into a cyclic pattern for the next two years. This fluorescein appears to have travelled 449 meters laterally beneath Wind Cave Canyon to arrive at this location in the cave.
have come off the parking lot. The only other location that fluorescein had ever been injected above the cave area was a considerable distance down dip from BI25 and RS11, near the southeast corner of the cave.

The fluorescein at these sites simply had to be coming from the picnic ground injection area, and it had not been present at these levels in any of the samples we had tested over the course of the previous three years. In light of all this, and considering the fact that Davis also reported fluorescein at RS11 soon after her 1986 trace, the fluorescein we were detecting is almost certain to have originated with our own 1996 injection in that same area. The lateral distance from the fluorescein injection point to BI25 is 449 meters.

As surprising as it was to verify lateral flow extending up to 449 meters from the injection point, the results closer to the injection point were equally surprising. Davis had reported fluorescein at NFP5 and NM4, and our own serendipitous discovery of optical brightener at NM4 during our Lycopodium experiment seemed to provide independent verification of a hydrologic connection between the picnic ground and NM4. If that connection exists, it did not exhibit itself during the course of this trace, suggesting it is sporadic in nature. Results from other studies suggest this is not unusual for vadose zone traces (Aley, 2012).

An examination of local precipitation data showed that 1986 and 1993 were both much wetter than normal, with an annual total of 58.2 cm in 1986 and 69.9 cm in 1993 (Ohms, 2012). Annual precipitation for 1996 was a more average 47.8 cm. Further, precipitation in the two months preceding the 1996 injection was only half of the 30 year average for the area (7.3 cm, compared to the average 14.5 cm). It is possible the system wasn’t active enough to move dye to NM4 in 1996.

Not surprisingly, fluorescein also arrived at BI9 at the same time it arrived at BI25, further leading us to believe that most of the water arriving at that site is water that has previously traveled through BI25. The concentrations detected at BI9 consistently lagged behind the concentrations measured at BI25 by 2-3 ppb. The consistency of the data between these two sites, together with a single fluorescein reading from a sample gathered at BI9 on day 29 (ten days into the fluorescein data gap for BI25 and RS11), hints that the fluorescein may have just started to arrive by that date.

On day 29, fluorescein readings were at 14 ppb at BI9, compared to 9 ppb on the day of injection. This was higher than any background sample from the site, and given the consistency of the 2-3 ppb lag behind BI25 readings, suggests that BI25 would have been reading 16-17 ppb. The increase is small and inconclusive, but it does suggest that it is at least possible that the leading edge of the fluorescein was arriving in the area about a month after injection. If true, the lateral flow beneath Wind Cave Canyon was traveling about 15-16 meters per day.

No other sites that were sampled could conclusively be deemed positive for fluorescein based on grab sample readings, although a small number of sites, including SA1 and UB10, eventually rose slightly above background levels. Charcoal bugs left for months at these sites were visibly fluorescent green after elution, suggesting these sites were positive for fluorescein, albeit very weakly.

**Fluorescein Recovery**

Background levels of fluorescein – either due to interference induced by a substance naturally in the water, or due to the presence of actual fluorescein – was present at both BI25 and RS11 in small but measurable levels prior to the July 29, 1996 injection. These background readings were subtracted from readings measured after dye began to arrive in order to arrive at an estimate for the quantity of fluorescein captured at these sites.

When calculated in the same manner as the rhodamine WT totals discussed above, it appears that 3.704 grams of fluorescein dye moved through BI25 during the first 752 days following injection. The total for RS11 during the same time period was 0.071 grams. As with the rhodamine calculations for this site, if we assume that twice as much water is passing through the RS11 area as we were able to capture and measure, then this total could be as high as 0.142 grams. This means that about 3.846 grams of the 2339 grams of fluorescein injected on July 29, 1996 (0.16% of the total) can be accounted for during the 752 days following the injection.

A major difference between the rhodamine and fluorescein recovery totals is that while the rhodamine total would not have increased significantly with additional monitoring (since the sites that were still positive when sampling ceased were only weakly positive, and mostly decreasing), the fluorescein levels being measured after 752 days were still quite positive, so fluorescein recovery totals likely would have continued to significantly increase had sampling continued. Figure 7, which documents the daily mass flux for each dye at BI25, illustrates this beautifully.
Epilogue

None of the dyes injected in the 1993 and 1996 dye traces ever emerged in measurable quantities at the park well. Grab samples were regularly collected from a spigot in the well house that draws water from the system prior to chlorination, and charcoal bugs were placed in a small piece of PVC pipe that was exposed to water continuously flowing from this spigot. None of the grab samples or bugs ever tested positively for any dye.

This seems to lend validity to Davis’ assertion that the rhodamine she found in the park well in the 1980s was the same dye she injected into Beaver Creek two months earlier. During the 1990’s park staff also discovered a powerful correlation between mean monthly flow rates in Beaver Creek – as measured by a USGS gaging station – and the rise and fall of the water table in the Lakes portion of the cave. A lag time of roughly two months was observed between a rise in Beaver Creek flow and a corresponding rise in the Lakes – roughly the same amount of time that passed between Davis’ rhodamine injection and the appearance of the dye in the well (which is located about the same distance from Beaver Creek as the Lakes area of the cave).

By 1998, sampling in the cave near the 1993 dye injections had grown so infrequent that we could no longer justify keeping the tarps in place that had been used to collect water and measure flow rates. Trips were scheduled to retrieve the tarps, collect final grab samples, and retrieve any bugs that were still present in the cave in those areas.

Four years, 10 months, and one day following the injection of the rhodamine WT in 1993, the collection site at CI17 (Pop Secret) was finally cleaned up. The site is nearly beneath the injection point of the 1993 rhodamine trace, and surprisingly, this sample turned out to be weakly positive for rhodamine – roughly 0.14 ppb. Additional grab samples collected 51 and 89 days later both produced readings of 0.36 ppb. Although these are the only three samples collected during this time, it appears we inadvertently caught the near-arrival of the dye during the April 1998 sampling.

Figure 7. Quantity of both dyes passing through BI25 on any particular day over time during the first two years of sampling. Graph represents dye recovered, not dye mixture recovered.
Due to CI17’s proximity to the 1993 injection site, and since no dye was recovered in any of the sites between CI17 and the 1996 rhodamine WT injection site, the likeliest explanation is that this represents a weak positive from the 1993 dye traces. Unlike the park’s developed area, water apparently takes almost five years to reach the cave here. If a longer lag time between dye injection and initial recovery during a dye trace has ever been measured, this author is not aware of it.

The discovery of dye at CI17 provided us with an additional insight. We had often wondered how long an activated charcoal bug could be left at a drip site in the cave and remain active. The charcoal bug that we retrieved in April 1998 had been in the cave since July 1995. If we are correct in our belief that we discovered the dye relatively soon after its initial appearance in the cave, then a charcoal bug exposed to dripping water at this site could still adsorb dye after being in the cave for nearly three years.

Complete turnover of the park’s cave management staff occurred during 1998, and as a result, regular sampling came to an end by late October of that year. Sampling was sporadic during 1999, and continued at some sites on an occasional basis into 2000. Following the collection and analysis of more than 4,400 grab samples and hundreds of bugs throughout the 1990’s, this exercise in patience finally came to an end.

Acknowledgments

It requires discipline to stick to a sampling schedule for years even when positive results fail to materialize. But as we learned, negative results in such experiments can produce valuable information as well, and surprising results can sometimes be attained years after all hope is gone. The author is grateful for the exceedingly dedicated cave management employees – both paid staff and volunteers – who assisted with data collection and analysis, especially Stan Allison, Bob Kobza, Brice Leech, Bonnie-Ann Burnett, and Tonya McDermott.

Marc Ohms continued collecting samples after the author left Wind Cave National Park in October 1998, and has provided very useful information and advice in the intervening years. His patience and diligence in assisting with his predecessor’s project helped squeeze the last bits of valuable data out of this long-term experiment.

Finally, NPS hydrologist Joe Meiman provided critical advice and training in study design, and also provided sampling equipment and our first fluorometer. He helped out on numerous occasions when we had trouble interpreting results, and served as our lead cheerleader when positive data began to pour in. I am deeply grateful for all of his assistance.

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A COMPARATIVE STUDY BETWEEN THE KARST OF HOA QUANG, CAO BANG PROVINCE, VIETNAM AND TUSCUMBIA ALABAMA, U.S.A.

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Abstract
Some of the most beautiful karst features created by the dissolution of limestone are residual hills with steep or vertical sides rising from a flat plain, known as tower karst. Tower karst to be developed requires a “mean annual temperature of minimum 17°C to 18°C and 1,000 to 1,200 mm/m² of annual rainfall (Jakucs, 1977). Two sites matching this criteria were selected: the karst of Hoa Quang District, Cao Bang Province, Vietnam, and Tuscumbia, Colbert County, Alabama, U.S.A. Preliminary observations regarding similarities and differences between these two sites are presented in this paper.

Introduction
The Hoa Quang karst area is located in the northern Vietnamese Province of Cao Bang. In 2014, a large number of karst springs, caves, sinking streams, and karst landforms were identified. Eighteen water samples were collected and analyzed for anions, cations, oxygen, and hydrogen stable isotope ratios.

The pH values are typical for karst waters and ranged from 7.23 to 7.97. Specific conductance values ranged from 153.2 to 421.6 µS/cm, the total alkalinity as CaCO₃ varies from 125 to 207 mg/L, and the carbon dioxide varies between 40.8 and 123.4 mg/L; whereas the values for the total hardness (as CaCO₃) are between 143 and 220 mg/L.

The local meteoric water line, based on our measurements is $\delta^2H = 7.93(±0.10)\delta^{18}O + 10.45 (±0.86)$ with $r^2=0.998$, which is close to the global meteoric water line (GMWL) $\delta^2H = 8.17\delta^{18}O + 10.35$ defined by Craig (1961) and revised by Rozanski, et al. (1993). The intercept value differs very slightly from both local and global water lines. Due to the short sampling period, the
information provided by the water stable isotopic composition is limited.

Carbonate rocks underlie many areas of north Alabama. Karst features can be found around Tuscumbia, in northwestern Alabama, which is part of the Tennessee-Alabama-Georgia karst area that is called TAG. TAG has the highest concentration of caves in United States, and home for a few large springs. Tuscumbia Spring is a municipal water supply with a base flow of 1,500 L/s. The field parameters, measured in January 2014, were: pH 6.81, specific conductance 292 uS/cm, and temperature 5.31°C Celsius.

In 1989-1990, the Geological Survey of Alabama conducted an extensive investigation in the area, performing dye studies in storm water drainage wells (SDW-1 through SDW-20) to define the recharge area of Tuscumbia Spring. The storm water drainage wells can be a potential source of contamination for the springs. Two rock samples from Vietnam and one from Tuscumbia, Alabama (U.S.A.) were collected and examined using the X-ray diffraction (XRD) analysis, microscopic analysis in polarized light and Differential Scanning Calorimetry-Thermogravimetry (DSC–TG) analysis. The quality of limestones in Vietnam and Tuscumbia (38.7 percent and 39.6 percent versus 31.10 percent calcium concentration) and the amount of precipitation (1,500 to 2,000 mm/m² in Vietnam versus 947 mm/m² to 1,960 mm/m² per year in Tuscumbia) are comparable.

Thick limestone beds, massively jointed, combined with frequent tectonic uplifts and a complex geologic pattern result in the tower karst landscape in Vietnam versus a leveled landscape in Tuscumbia, Alabama. Tectonics is the primary driver for the formation of tower karst landscape in Cao Bang Province, Vietnam.

**Karst of Hoa Quang District, Cao Bang Province, Vietnam**

**Geographic and Climatic Setting**

Vietnam is in the eastern part of the Indochinese Peninsula (Figure 1, Insert). It covers a total area of 331,210 km² of which mountains, mostly covered by forest, represent 40 percent. Carbonate rocks are exposed over 60,000 km², which represents 18.12 percent of Vietnam (Clements et al., 2006).

Cao Bằng Province is located in northeastern Vietnam, 270 kilometers north of Hanoi, on the border with China. The province comprises 6,724.6 km² and has a population of 632,450. The topography of the region is characterized by mountain ranges with elevations over 900 m mean sea level (MSL), and karstic plateaus developed between 500 and 700 m MSL. The elevation of the town of Cao Bang is 300 meters MSL and it has a temperate climate throughout the year. Annual rainfall ranges between 1,500 and 2,000 mm/m² and temperature varies from 5°C Celsius in December and January to 37°C Celsius in July and August. The average temperature in the province is 22°C Celsius. Winter temperatures in some areas occasionally fall below freezing with snow.

The study area is located in the Hoa Quang Districts. The topography consists of small alluvial plains along the Bac Vong River and its tributaries, with large poljes surrounded by limestone pinnacles and tower karst (Figure 1), sinkholes (dolinas), closed depressions, sinking streams, springs, and large underground streams or submerged passages (Ponta et al., 2013).

**Regional Geology**

Northeastern Vietnam is in a tectonic active region located at the boundary between the Indian Plate and Eurasian Plate, which created the Himalayas Mountain System and the Tibetan Plateau (Strong Wen et al., 2015). The area is underlain by numerous rock types and ages, which have undergone numerous phases of tectonic deformation from middle Paleozoic to present (Tran Thanh Hai, 2009). The site is located a few kilometers north of Ailao Shan-Red River Shear Zone and Song Ma Fault Zone, which divides North Vietnam into two main tectonic units: The Bac Bo Fold Belt (part of Eurasian Plate/South China Terrane) and the Indochina Fold Belt (part of Indian Plate/Indochina Terrane).

The Bac Bo Fold Belt is composed by three fold systems: Tay Bac, Viet Bac and Dong Bac. Cao Bang Province is located in the Dong Bac Fold System. Cao Bang Province is traversed by two deep fault systems, Lo-Gam and Cao Bang - Lang Son, which divide the region into three parts: the western uplift block (the “Viet BAC” with the Bong Son anticline), the central part (Song Hien), and the eastern Ha Lang zone (anticline). The province is underlain by a variety of rocks ranging in age from Cambrian to Quaternary (Long, 2001).

The Viet Bac uplift has Paleozoic rocks outcropping along the NW-SE axis. Between the two deep faults is a Mesozoic low with sedimentary and volcanic rocks of Triassic age (not shown on Figure 1), including iron and manganese deposits. Northeast of the Cao Bang - Lang Son fault, outcrops comprise Paleozoic sediments with complex folds and fault systems (Ha Lang zone, Figure 1).
Figure 1. Geological map of Hoa Quang District, Cao Bang Province, Vietnam. Geology after Pham Dinh Long, 2000 (carbonate rocks are shown in white).
Geology of the Hoa Quang District

The Hoa Quang karst area is 15 km long and 14 km wide (210 km²). The central part of the study area is mainly a NW-SE trending faulted anticlinorium plunging northwestward. The core of the anticline is formed by Cambrian and Devonian sandstones and siltstones belonging to the Than Sa (C1ts), and Luoc Khieu (D1lk) Formations, intercalated with conglomerates, shale, and clay layers of Lower Devonian age (Son Cau Group [D1sc] and Mia Le Formation [D1ml]). The western flank is crossed by two parallel faults one km apart, oriented north-south, and the eastern flank by several faults and thrust faults parallel with the anticline’s axis, which separates the Na Quan (D1–2nq) and Toc Tat (D3tt) Formations of Devonian age. In the northeastern part of the study area the Bac Son Formation (C-Pbs) outcrops with siliceous shale, shale-limestone, light gray limestone, and gray limestone (Long, 2001).

The limestone unit is up to 800 m thick, finely crystalline, and light gray to dark gray. Bedding is generally 30 to 50 cm thick, oriented NW-SE, dipping 19° to 25° SW, E, or NE.

Karst Landforms

The northeastern part of Vietnam provides a unique cave-forming environment with large rivers draining into the limestone valleys from the mountains. The Hoa Quang karst has developed into a tower karst landscape that overlooks valleys, poljes, and closed basins. The area underlain by limestones is extensive. Rainfall is abundant and the karst’s vertical potential exceeds 200 m. Exokarst landforms are well represented by a variety of karrens, tsingi, small- to medium-sized sinkholes, and poljes. The limestone massifs (tower karst) are traversed by caves that carry rivers from one polje to the next that overlooks valleys, poljes, and closed basins.

East of the anticline, the tributaries of the Bac Vong River flow through a sequence of caves, springs, and poljes (Figure 2, Cross Section A-A’). The Ban Vong’s tributaries collect waters a few hundred meters south of the border with China, and disappear underground through temporary or permanent swallets six times along a 13-km path, before resurfacing in the Hang Ban Nua stream cave (HQ-32) with a flow rate of 800 L/s.

The Hoa Quang area is an old karst where isolated hills (tower karst) remain upon residual plains developed on an 800-m thick limestone unit. As illustrated on cross section A-A’ (Figure 2) tower karst emerges at the intersection of the land surface with faults/joints or bedding planes. These structural features appear on the 1:200,000-scale geologic map. More than likely, on a geologic map at a larger scale (1:10,000 to 1:100,000), more structural features will be documented.

The presence of caves, phreatic or vadose, at the contact between towers and alluvial plains (foot-hill caves) are common. In some areas, two levels of cave passages are developed, one as a stream passage at the present flood plain level, and a fossil cave at 50 m higher in elevation, marking a former flood plain level. The springs are “free draining,” recharged by allogenic underground streams with a significant autogenic contribution in the rainy season. The flow pattern is dendritic (a branch-work cave pattern where underground passages are present). The type of tower karst land-

Figure 2. Cross Section A-A’ along Bac Vong River’s tributaries, traversing tower karst landscape in Hoa Quang District, Cao Bang Province, Vietnam; B-B’ Cross section in Tuscumbia, Alabama, U.S.A.
scape developed along the Bac Vong’s tributaries and Tra Linh River is “residual hills on a planed limestone surface” (Ford and Williams, 2007).

### Sampling Methods

Eighteen water samples from cave waters (rimstone pools), rainwater, sinking streams, and springs were collected and analyzed for anions, cations, and oxygen and hydrogen isotopes. Selected water-quality parameters (pH, temperature, specific conductance, and salinity) were monitored at each sampling point with an YSI 63 instrument. Additionally, a Digital Tittrator (Hach Model 16900) was used in the field to determine total hardness as alkalinity, total hardness as calcium, and carbon dioxide (CO₂); whereas iron was measured with a Hach Ferrous Iron test kit Model IR-18C. UTM coordinates and elevation were recorded with a GPS (Garmin 62S) at each sampling point (Table 1).

Groundwater samples from each source were collected and pre-treated as shown below. Samples for cations were collected in 40 mL vials pre-treated with nitric acid. Samples for anions and stable isotopes were collected in 40 mL vials with no preservatives.

Anion analyses were performed at the University of Alabama using a Dionex DX 600 Ion Chromatograph. The trace metals were analyzed at the Department of Geology, University of South Florida, using a Perkin-Elmer Elan DRC II Quadrupole inductively coupled plasma mass spectrometer (ICP-MS) analytical instrument. Standards used were formulated stock standards with metals in concentrations from 1,000 mg/L to 10,000 mg/L.

The stable isotopes analyses were performed at Stable Isotope Laboratory of the Babeş-Bolyai University, Cluj-Napoca, Romania, using a Picarro L2130-i Cavity Ring Down Spectroscopy (CRDS) instrument.

To evaluate the limestones from the study area, three rock samples (two from Vietnam, and one from Tuscaloosa, Alabama, U.S.A) were collected and examined at R&D National Institute for Nonferrous and Rare Metals Romania using the X-ray diffraction (XRD) analysis, microscopic analysis in polarized light, and Differential Scanning Calorimetry- Thermogravimetry (DSC–TG) analysis. The DSC-TG thermic analyses were conducted with Setsys Evolution instrument (Setaram). A Zeiss Axiosmager A1m polarizing microscope was used for the thin section analyses, and the X-ray diffraction (XRD) analysis was performed with Bruker D8 Advance diffractometer.

### Assessment of Results

#### Water-Quality data

The field parameter data are presented in Table 1. Sampling locations are shown on Figure 1. The elevations of sampling points ranged between 443 m and 641 m MSL. The estimated flow rates of the sampled springs ranged from 0.5 L/s to 800 L/s. The observations were recorded

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<th>Sample Source</th>
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<th>Temperature (°C)</th>
<th>Salinity (mg/L)</th>
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<td>150.0</td>
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<td>150.0</td>
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<td>2.0</td>
<td>50.0</td>
<td>150.0</td>
</tr>
</tbody>
</table>

Table 1. Field parameters Hoa Quang District, Cao Bang Province, Vietnam

14TH SINKHOLE CONFERENCE NCKRI SYMPOSIUM 5 131
in February 2014, at the end of the dry season, so these values are characteristic for base flow.

Temperatures ranged from 13.5°C to 23.2°C Celsius, which correspond to mean annual air temperature in the area. The pH values ranged from 7.23 to 7.97, typical for karst waters. The values measured for the specific conductance are typical for cave waters (between 153.2 and 421.6 uS/cm). Total alkalinity as CaCO₃ varies between 125 and 207 mg/L, the total hardness as CaCO₃ varies between 143 and 220 mg/L, and carbon dioxide (CO₂) ranged from 40.8 mg/L to 123.40 mg/L.

Laboratory results for anions and cations are provided in Table 2. Calcium concentration ranged from 51.75 to 81.23 mg/L. The water is classified as calcium-magnesium bicarbonate.

**Stable Isotopes**

Isotope fractionation accompanying evaporation from the ocean and condensation during atmospheric transport of water vapor causes spatial and temporal variations in the deuterium and δ¹⁸O composition of precipitation (Dansgaard, 1964). Regional-scale processes such as water vapor transport patterns across landmasses and the average rainout history of the air masses precipitating at a given place control the isotopic composition of local precipitation (Ponta et al., 2013).

Globally, the relation between the δ¹⁸O and δD is mainly explained by the general Rayleigh distillation principle. The relation between the δ¹⁸O and δD values was recognized by Craig (1961), and defined as “global meteoric water line” (GMWL). The complex processes that are involved in the isotope fractionation alter the general (oversimplified) principle and produce local relations (LMWL) between the water δ¹⁸O and δD values.

The relationship between δ¹⁸O and δD in world’s fresh surface/cave waters is described by the global meteoric water line (GMWL) defined by Craig (1961) as:

\[ δD = 8 \ δ^{18}O + 10\% \]

The δ¹⁸O values of the water samples collected ranged from -8.88 to -2.34‰ and the δD values from -58.89 to -7.87‰. The water sample with δ¹⁸O value of -2.34‰ was collected from a rimstone pool in the cave TR33 (Table 1). This sample plots very close to the GMWL, which indicate a close link to a Rayleigh process. Higher values probably are linked to slower infiltration rate of precipitation water from warmer months.

The local meteoric water line (LMWL) in Vietnam is:

\[ δD = 7.91(±0.19) δ^{18}O + 12.44 (±1.25) \]

with r²=0.990 (Figure 3) on the basis of 38 measurements (between 2004 and 2007) from the IAEA rainfall monitoring station in Hanoi (N21°2'43", E105°47'55") (IAEA/WMO, 2015).

The local meteoric water line, based on our measurements (Table 3) is:

\[ δD = 7.93(±0.10) δ^{18}O + 10.45 (±0.86) \]

with r²=0.998 which is close to the global meteoric water line (GMWL) δD = 8.17 δ¹⁸O + 10.35 defined by Craig (1961) and revised by Rozanski et al. (1993). The

### Table 2. Summary of water-quality data Hoa Quang District, Cao Bang Province, Vietnam

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Location Name</th>
<th>Calcium</th>
<th>Magnesium</th>
<th>Sodium</th>
<th>Potassium</th>
<th>Sulfate</th>
<th>Chloride</th>
<th>Nitrate</th>
<th>Hardness</th>
<th>Conductivity</th>
</tr>
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<tbody>
<tr>
<td>1</td>
<td>TR-33 Cave</td>
<td>Na⁺</td>
<td>K⁺</td>
<td>Ca²⁺</td>
<td>Mg²⁺</td>
<td>Cl⁻</td>
<td>SO₄²⁻</td>
<td>NO₃⁻</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Long Cac TR-14</td>
<td>58.90</td>
<td>14.39</td>
<td>9.87</td>
<td>0.19</td>
<td>0.89</td>
<td>9.02</td>
<td>0.04</td>
<td>0.04</td>
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<td>91.84</td>
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<td>0.30</td>
<td>0.19</td>
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<td>0.04</td>
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<td>4.00</td>
</tr>
<tr>
<td>4</td>
<td>TR-84</td>
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<td>0.86</td>
<td>2.86</td>
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<td>0.04</td>
<td>0.04</td>
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</tr>
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<td>TR-IC pool</td>
<td>9.86</td>
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<tr>
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<td>TR-1 Caver</td>
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<td>56.72</td>
<td>56.72</td>
<td>0.04</td>
<td>0.04</td>
</tr>
</tbody>
</table>
Deuterium excess values, defined as $d = \delta D - 8\delta ^{18}O$ (‰) (Dansgaard, 1964), calculated for the collected waters from karst areas (Table 3) are typical for continental meteoric waters (deuterium excess values of ~10 ‰) and have a median value of 11.51 ‰. There are some samples with lesser or higher deuterium excess values due to local variations in humidity, temperature, and wind speed.

**Regional Geology**

The Tuscumbia karst area is located on the southern flank of the Nashville dome, in the Interior Low Plateaus physiographic province. The area is underlain by sedimentary formations of Mississippian age limestones, sandstones, and shales). Local geology is typical of the terrain associated with limestone bedrock, overlain by a mantle of clay-rich, unconsolidated material, 5 to 8 meters thick (Figure 4). The rocks dip generally toward the southwest at 5 m per kilometer (0.3° SW).

The Fort Payne Chert-Tuscumbia Limestone aquifer system with a thickness of 160 meters is the most important water-bearing unit in the Tuscumbia area. It is part of the Mississippian aquifer system that underlies three counties in northwestern Alabama (Colbert, Franklin, and Lawrence). The Chattanooga Shale (not shown on Figure 4), at the base of the aquifer system, restricts the downward movement of water to lower units, which are not known to be water-bearing (Chandler and Moore, 1991).
Geology of the Tuscumbia Area

The study area is 12 km long and 12 km wide (144 km²). The Fort Payne Chert (not shown on Figure 4) of Lower Mississippian age consists predominantly of medium gray to light gray and white crystalline limestone containing abundant light gray to black bedded and nodular chert (Szabo, 1975). Tuscumbia limestone (Mt) of Valmeyeran age is a medium-bedded light bluish-gray hard, dense, fine-to medium-grained bioclastic limestone containing abundant light-colored bedded and nodular chert (Szabo, 1975). The Tuscumbia limestone is overlain by the Pride Mountain Formation (Mpm) of Chesterian age and consists of shale, a basal limestone up to 3 m thick, and sandstone and siltstone (Szabo, 1975). The Hartselle sandstone (Mh) is a light gray massive to thin-bedded quartzose sandstone, which unconformably overlies the Pride Mountain Formation (Szabo, 1975). The system of fractures (joints) at the site consists of a northwest-trending set, as evidenced by the alignment of closed depressions and sinkholes in the area (Figure 4), and a southwest-trending set identified through lineament interpretations by Chandler and Moore in 1991.

Karst Landforms

The karst plain developed at 152 m MSL in the study area looks like a river bedrock terrace. It is part of a mature karst with low topographic relief except for the southern edge of Pickwick Lake, where limestone bluffs occur (Chandler and Moore, 1991). Exokarst landforms are well represented by a variety of small to medium-sized sinkholes and closed depressions (Figure 2, Cross Section B-B’).

Sinkholes in this area are very effective in collecting surface runoff waters. The drainage is primarily subterranean (Johnston, 1933). Spring Creek, with its tributaries, is the only surface stream that crosses the sinkhole plain (Johnston, 1933).

Due to urban development, some of the karst features represented on Figure 4 (based on a 1975 7.5-minute

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Sample Name</th>
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<th>δ¹⁸O</th>
<th>δD</th>
<th>dD-GMWL</th>
<th>d-excess</th>
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</tr>
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<td>Chu Doi TR-35</td>
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</tr>
<tr>
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<td>-53.68</td>
<td>10.34</td>
</tr>
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<td>-54.22</td>
<td>-55.36</td>
<td>11.14</td>
</tr>
<tr>
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<td>Tuc Po HQ-34</td>
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<td>-55.12</td>
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<td>-53.97</td>
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<td>-52.72</td>
<td>9.86</td>
</tr>
<tr>
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<td>Dua Nua Sink HQ-33</td>
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<td>-53.85</td>
<td>-55.20</td>
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<td>14</td>
<td>TR-24</td>
<td>2/5/2014</td>
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<td>-54.80</td>
<td>-55.52</td>
<td>10.72</td>
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<tr>
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<td>TR-26</td>
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<td>-58.89</td>
<td>-60.56</td>
<td>11.67</td>
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<tr>
<td>18</td>
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<td>-56.66</td>
<td>-58.00</td>
<td>11.94</td>
</tr>
</tbody>
</table>

*In Red: Water samples collected from sinking streams.*

Table 3. Summary of stable isotopes data Hoa Quang District, Cao Bang Province, Vietnam
Figure 4. Geological map of Tusculumia, Alabama U.S.A. Geology after Szabo 1975. Groundwater Flow Directions after Chandler and Moore 1993 (carbonate rocks shown in white).
Tuscumbia Spring is a municipal water supply source with a base flow of 1,500 L/s. According to the Geological Survey of Alabama’s Groundwater Assessment Program (Real Time Monitoring), between May 22, 2013, and February 9, 2015, the average flow was 2,133 L/s with a minimum of 1,423 L/s and a maximum of 6,211 L/s (Geological Survey of Alabama webpage). The field parameters, measured in January 2014, were: pH 6.81, specific conductance 292 uS/cm, and temperature 15.31°C Celsius. The recharge area of Tuscumbia Spring is 218 km².

Dye Studies
In 1989-1990, the Geological Survey of Alabama conducted an extensive investigation in the area, performing dye trace testing to define the recharge area of Tuscumbia Spring. Seventeen storm water drainage wells were selected for the dye study. The results of fluorescent dye-trace testing indicate that hydrologic interconnections exist between 17 wells and springs of the Tuscumbia area (Table 4).

Strong recoveries of tracers at Tuscumbia Spring were recorded from SDW-11 and SDW-16 (Figure 4, Table 4). The travel-time data (3 hours) and high groundwater velocities (463 m to 705 m/h) indicate channelized flow conditions between Tuscumbia Spring and storm water drainage well SDW-11. The water is of the calcium-sodium-bicarbonate-sulfate-chloride type, slightly alkaline and hard, but low in mineral and chemical constituents (Chandler and Moore, 1993).

Rock samples
Laboratory results are provided in Table 5. The two rock samples collected in Vietnam (latitude 22°N, Figure 1) had calcium concentrations ranging from 38.7 percent and 39.6 percent, while the magnesium concentration fluctuated between 0.14 percent and 0.16 percent. The presence of silica was insignificant. The microscopic analysis in polarized light revealed that both samples collected in Vietnam are micritic limestones with veins filled with secondary calcite spar. The sample collected in Tuscumbia Alabama, U.S.A., (latitude 34°N, Figure 4) had a calcium concentration of 31.1 percent, silica 8.9 percent, and 0.28 percent aluminum. In both cases, the carbonate rocks are middle to upper Paleozoic age.

Summary
1. Hoa Quang, Vietnam is at latitude 22°N., in the peritropical zone of excessive planation. Tuscumbia, Alabama, is located at latitude 34°N., in a subtropical zone of mixed relief development (Budel, 1977).
2. In the area studied in Vietnam, bedding is generally 30 to 50 cm thick, oriented northwest-southeast, and dipping 19° to 25° southwest or northeast; the Tuscumbia limestone in Alabama (50 to 100 cm thick) dips gently southwestward at approximately 4.8 to 5.7 m per kilometer (0.3° SW) (Szabo, 1975).
3. The two rock samples collected in Vietnam (Figure 1) have calcium concentrations ranging from 38.7 percent siliceous limestone to 39.6 percent, whereas the sample collected in Tuscumbia Alabama, U.S.A., (Figure 4) has a calcium concentration of 31.1 percent (with 8.9 percent silica, and 0.28 percent aluminum). In both cases, the carbonate rocks are middle to upper Paleozoic age.
4. In Vietnam, the central part of the study area is mainly a northwest-southeast trending faulted anticlinal plunging toward the northwest. The western flank is traversed by two parallel faults one km apart, oriented north-south, and the eastern flank by several faults and thrust faults parallel with the anticline’s axis.
5. In Tuscumbia, Alabama, on the 7.5-minute topographic/geological map (Szabo, 1975) no faults/joints are mapped. In Chandler and Moore’s 1991 publication, relatively dense sets of lineaments oriented mainly northwest-southeast and northeast-southwest are interpreted from LANDSAT imagery, high altitude black and white, color photography, and 7.5-minute topographic maps.
6. The amount of precipitation in Tuscumbia, Alabama, ranged between 947 mm/m² to 1,960 mm/m² per year versus 1,500 to 2,000 mm/m² per year in Vietnam.
7. In the water samples collected in Vietnam, the amount of dissolved carbon dioxide (CO₂) ranged from 40.8 mg/L to 123.40 mg/L (Table 1) and calcium concentration ranged between 51.75 to 81.23 mg/L (Table 2).
8. In Vietnam, the flow network is dendritic with streams recharged by epigenetic springs and meteoric water. In Tuscumbia, Alabama, the surface flow pattern is poorly organized.
9. The maximum discharge of a spring in Vietnam is 800 L/s versus a base flow of 1500 L/s in Tuscumbia. The waters in Tuscumbia, Alabama, are collected from 280 km² versus 20 to 30 km² in Vietnam.
<table>
<thead>
<tr>
<th>Tested Well/Injection Point</th>
<th>Tuscumbia Spring (SP-1)</th>
<th>SP-2</th>
<th>SP-3</th>
<th>SP-4</th>
<th>SP-5</th>
<th>SP-6</th>
<th>SP-7</th>
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<tbody>
<tr>
<td>Flow (L/min)</td>
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<td>0.06 - 100</td>
<td>3.7 - 35</td>
<td>0.3 - 4</td>
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</tr>
<tr>
<td>SDW-1</td>
<td>Distance (km)</td>
<td>Travel Time (Days)</td>
<td>4.5</td>
<td>20</td>
<td>5.31</td>
<td>5.63</td>
<td>5.15</td>
</tr>
<tr>
<td>SDW-2</td>
<td>Distance (km)</td>
<td>Travel Time (Days)</td>
<td>5.54</td>
<td>20</td>
<td>4.51</td>
<td>5.15</td>
<td>4.67</td>
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<tr>
<td>SDW-3</td>
<td>Distance (km)</td>
<td>Travel Time (Days)</td>
<td>4.02</td>
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<td>4.67</td>
<td>3.71</td>
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<td>SDW-4</td>
<td>Distance (km)</td>
<td>Travel Time (Days)</td>
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<td>&lt;1</td>
<td>4.51</td>
<td>5.63</td>
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<td>Distance (km)</td>
<td>Travel Time (Days)</td>
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<td>&lt;1</td>
<td>5.86</td>
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<td>6.80</td>
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<td>SDW-8</td>
<td>Distance (km)</td>
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<td>6.92</td>
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<td>SDW-9</td>
<td>Distance (km)</td>
<td>Travel Time (Days)</td>
<td>7.24</td>
<td>5</td>
<td>8.05</td>
<td>8.05</td>
<td>7.08</td>
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<td>SDW-10</td>
<td>Distance (km)</td>
<td>Travel Time (Days)</td>
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<td>7.24</td>
<td>5</td>
<td>7.56</td>
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<td>SDW-11</td>
<td>Distance (km)</td>
<td>Velocity (m/h)</td>
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<td>&lt;1</td>
<td>4.34</td>
<td>4.51</td>
<td>4.02</td>
</tr>
<tr>
<td>SDW-15</td>
<td>Distance (km)</td>
<td>Travel Time (Days)</td>
<td>3.54</td>
<td>19</td>
<td>4.51</td>
<td>5.15</td>
<td>4.67</td>
</tr>
<tr>
<td>SDW-16</td>
<td>Distance (km)</td>
<td>Velocity (m/h)</td>
<td>3.45</td>
<td>&lt;1</td>
<td>4.51</td>
<td>5.63</td>
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<td>SDW-17</td>
<td>Distance (km)</td>
<td>Travel Time (Days)</td>
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<td>&lt;1</td>
<td>4.34</td>
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<td>4.67</td>
</tr>
<tr>
<td>SDW-20</td>
<td>Distance (km)</td>
<td>Travel Time (Days)</td>
<td>3.54</td>
<td>&lt;1</td>
<td>5.00</td>
<td>4.99</td>
<td>4.67</td>
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Table 4. Summary of dye study data Tuscumbia Alabama from Chandler and Moore 1991

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<tr>
<th>Sample Name</th>
<th>Sample Type</th>
<th>Calcium</th>
<th>Magnesium</th>
<th>Sodium</th>
<th>Potassium</th>
<th>Iron</th>
<th>Trace Element</th>
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<tbody>
<tr>
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<td>34.14</td>
<td>0.40</td>
<td>1.80</td>
<td>0.61</td>
<td>0.41</td>
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<tr>
<td>2 Rock Sample</td>
<td>Calcite</td>
<td>30.60</td>
<td>28.4</td>
<td>0.26</td>
<td>0.02</td>
<td>0.02</td>
<td>0.26</td>
</tr>
<tr>
<td>3 Rock Sample</td>
<td>Dolomite</td>
<td>38.40</td>
<td>34.14</td>
<td>3.00</td>
<td>1.80</td>
<td>0.61</td>
<td>0.41</td>
</tr>
</tbody>
</table>

Table 5. Summary of rock sample data
Conclusions
The Hoa Quang karst has developed into a tower karst landscape that overlooks valleys, poljes, and closed basins. The area underlain by limestones is extensive. Rainfall is abundant and the karst’s vertical potential exceeds 200 m. Exokarst landforms are well represented by a variety of karrens, tsingi, small- to medium-sized sinkholes, and poljes.

As shown on cross section A-A’, the tower karst in Vietnam emerges at the intersection of the land surface with a fault/joint or a bedding plane parting. These structural features appear on the 1:200,000-scale (medium scale) geologic map. More than likely, when a geologic map at a larger scale (1:10,000 to 1:100,000) is prepared, more structural features will be documented.

The base map for the Tuscumbia area is a 1:24,000-scale topographic/geologic map. No fractures or faults are shown and the bedding planes are nearly horizontal. The quality of limestones in Vietnam and Tuscumbia (38.7 percent and 39.6 percent versus 31.10 percent calcium concentration) and the amount of precipitation (1,500 to 2,000 mm/m² in Vietnam versus 947 mm/m² to 1,960 mm/m² per year in Tuscumbia) are comparable.

Thick limestone beds, massively jointed, combined with frequent tectonic uplifts and a complex geologic pattern result in the tower karst landscape in Vietnam versus a leveled landscape in Tuscumbia, Alabama (Figure 2). Tectonics is the primary driver for the formation of tower karst landscape in Cao Bang Province, Vietnam.

Acknowledgments
This material is based upon work supported by the National Speleological Society Research Grant to Gheorghe Ponta. Sponsors for the hydrogeological investigation are the Institute of Geosciences and Mineral Resources Vietnam, Geokarst Adventure, the University of South Florida, U.S.A, and University Babes-Bolyai, Cluj-Napoca, Romania.

References
HYDROCHEMICAL CHARACTERISTICS AND FORMATION MECHANISM OF GROUNDWATER IN THE LIULIN KARST SYSTEM

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Abstract
The Liulin karst system is typical of hydrogeological systems in northwestern China, with a group of springs as the dominant mechanism for regional groundwater discharge. To reveal the hydrochemical formation mechanism of the Liulin karst groundwater system, we studied the hydrogeochemical processes of karst groundwater in aquifers at the base of the hydrogeological investigation. Then starting from the chemical composition of karst groundwater together with the recharge-runoff-discharge process of groundwater systems, we analyzed the solutes origin and the dissolved mineral facies of the groundwater chemical composition. The results showed that the anionic and cationic compositions of karst water were different in recharge, runoff and discharge areas, with the main anions of $\text{HCO}_3^-$ and $\text{SO}_4^{2-}$ in recharge areas, and $\text{HCO}_3^-$ and $\text{Cl}^-$ in runoff and discharge areas, as well as the main cationic for $\text{Ca}^{2+}$ and $\text{Na}^+$, of which the molar concentrations of $\text{Ca}^{2+}$ was greater than $\text{Na}^+$ in recharge areas and contrary to the runoff and discharge areas. Karst water was influenced by carbonate and evaporite dissolution while flowing through the aquifers, of which carbonate rock dissolution dominated in recharge areas, and evaporite rock dissolution increased to be the dominate lithology in runoff and discharge areas. Based on analysis of water-rock interaction, the main dissolved mineral facies included dolomite, calcite, gypsum and halite. Dolomite is the most important dissolved mineral, followed by calcite and gypsum in recharge area, as well as calcite, gypsum and halite in runoff and discharge areas.

Introduction
The forming conditions of groundwater chemical composition are obviously different from those in surface water, which are often controlled by the geological and hydrogeological factors (Shen, 1993). Groundwater interacts with surrounding media and converts the chemical composition through geologic time. By the study of groundwater chemical characteristics, we can better understand the interaction mechanism of groundwater and the environment (Zhang et al., 2000), as well as the mineral rock facies in the water-rock interaction (Pu et al., 2013; Wang et al., 2006). Previous studies in this area include analysis of the model of karst water systems from the macro perspectives of time and space (Gao et al., 2008), resources development and protection (Li, 2010), as well as the evolution of water chemical characteristics and recharge source (Pei and Liang, 2005; Wei et al., 2012). Few studies have been done on the solute source and dissolved mineral facies of karst water in the Liulin karst system (Zang et al., 2013). This paper seeks to apply the relationship of cations and anions...
of karst water in accordance with the recharge-runoff-discharge process to reveal the chemical characteristics and discuss the solute source and dissolved mineral facies of Liulin karst groundwater, aiming at providing a certain theoretical basis for reasonable development and protection of the karst water.

**Study Area**

The Liulin karst system is hydrogeologically typical of karst systems in northern China. A group of springs act as the predominant mechanism for regional groundwater discharge where carbonate rocks are widely distributed. They represent relatively independent karst systems. A distance of about 3 km along the Sanchuan River valley outcrop exists more than 100 springs with annual average discharge of 2.63 m³/s. These springs outcrop in five groups and are collectively called the Liulin springs. Liulin springs discharge from the Cambro-Ordovician carbonate karst fracture, and are mainly supplied by precipitation via a carbonate fissure within a bare area of about 1,400 km². Secondly, the springs are recharged by river leakage from Sanhe valley via six 33 km long fissures. Related subsurface flow runs from northeast to southwest controlled by geological structures. Water discharges from scattered springs on both sides of the Sanchuan from Ordovician Fengfeng Group.

**Sampling and Analysis**

In September 2011 and May 2013, fourteen groundwater samples were collected from central region. Electrical conductivity (EC), dissolved oxygen (DO), pH and temperature measurements were taken immediately after sampling. Then Atomic absorption spectroscopy (AAS) (PE - 601) was used for the cation content (Na⁺, K⁺, Ca²⁺, Mg²⁺) measurement, and high-performance liquid chromatography (HP1100) was used for the anion content (Cl⁻, SO₄²⁻, HCO₃⁻, NO₃⁻) measurement. Trace elements were measured by inductively coupled plasma mass spectrometry (ICP-MS).

**Results**

**Major Ions of Karst Groundwater**

As shown in Table 1, the pH values varied from 7.30 to 7.74. HCO₃⁻ is the dominant anion with the concentration ranging from 4.34 mmol/L to 5.19 mmol/L, and accounts for more than 70% of the total anion content. The concentration of secondary anion SO₄²⁻ and Cl⁻ range from 0.21 mmol/L to 2.35 mmol/L and 0.25 mmol/L to 2.31 mmol/L, separately. In the recharge, runoff and discharge areas, HCO₃⁻ is always the dominant anion in chemical composition of karst water. While there are differences in the secondary anion: the average concentration of karst water in accordance with the recharge-runoff-discharge process to reveal the chemical characteristics and discuss the solute source and dissolved mineral facies of Liulin karst groundwater, aiming at providing a certain theoretical basis for reasonable development and protection of the karst water.

**Study Area**

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**Table 1. The chemical composition of major ions in Liulin springs (Unit: mmol/L).**

<table>
<thead>
<tr>
<th>Time</th>
<th>Location</th>
<th>pH</th>
<th>K⁺</th>
<th>Na⁺</th>
<th>Ca²⁺</th>
<th>Mg²⁺</th>
<th>HCO₃⁻</th>
<th>Cl⁻</th>
<th>SO₄²⁻</th>
<th>NO₃⁻</th>
<th>TDS mg/L</th>
</tr>
</thead>
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<td>Wet period</td>
<td>Wucheng</td>
<td>7.56</td>
<td>0.03</td>
<td>0.41</td>
<td>1.77</td>
<td>0.66</td>
<td>4.42</td>
<td>0.27</td>
<td>0.23</td>
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<td>7.34</td>
<td>0.14</td>
<td>1.33</td>
<td>3.36</td>
<td>1.31</td>
<td>5.19</td>
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<td>2.35</td>
<td>0.27</td>
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<td>7.57</td>
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<td>1.94</td>
<td>0.67</td>
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<td>0.48</td>
<td>0.09</td>
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<td>0.80</td>
<td>4.44</td>
<td>1.03</td>
<td>0.47</td>
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<td>1.27</td>
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<td>0.16</td>
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<td>1.81</td>
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<td>1.41</td>
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<td>0.81</td>
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<td>2.02</td>
<td>1.01</td>
<td>4.44</td>
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<td>0.97</td>
<td>0.20</td>
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<td>0.93</td>
<td>4.53</td>
<td>1.52</td>
<td>0.84</td>
<td>0.17</td>
<td>444.1</td>
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SO\textsubscript{4}^{2-} is greater than that of Cl\textsuperscript{-} in recharge areas, and it is opposite in runoff and discharge areas. Therefore, the concentration of SO\textsubscript{4}^{2-} decreases and Cl\textsuperscript{-} increases along the recharge-runoff-discharge pathway. Karst groundwater contains relatively high levels of Ca\textsuperscript{2+} and Na\textsuperscript{+} as the major cations, accounting for more than 75% of the total. The concentration of Ca\textsuperscript{2+} and Na\textsuperscript{+} vary from 1.57 mmol/L to 3.36 mmol/L and 0.35 mmol/L to 1.82 mmol/L, separately, which also shows a certain difference among the zonings. The average concentration of Ca\textsuperscript{2+} is greater than Na\textsuperscript{+} in the recharge area, and it is opposite in runoff area and discharge areas.

As was shown in Figure 1, whether a wet or dry period, the cations of karst groundwater mainly plot on Ca\textsuperscript{2+}-Na\textsuperscript{+} line with close to the Ca\textsuperscript{2+} end in recharge areas, and close to the Na\textsuperscript{+} end in runoff and discharge areas. The anions mostly plot at the end of HCO\textsubscript{3}-, while scatter to the end of SO\textsubscript{4}^{2-} and Cl\textsuperscript{-} in runoff and discharge areas.

Stoichiometry of Karst Groundwater

This paper referred to the Gibbs diagram to establish the relationship between total dissolved solids (TDS) and Na\textsuperscript{+}/(Na\textsuperscript{+}+Ca\textsuperscript{2+}) as well as Cl\textsuperscript{-}/(Cl\textsuperscript{-}+HCO\textsubscript{3}-) (Figure 2) to reveal the relationship of the chemical composition and the lithology of aquifer correlation (Gibbs, 1970).

The results show that karst groundwater in Liulin springs had lower values of TDS and Na\textsuperscript{+}/(Na\textsuperscript{+}+Ca\textsuperscript{2+}), as well as Cl\textsuperscript{-}/(Cl\textsuperscript{-}+HCO\textsubscript{3}-), which reflects that river leakage is the main groundwater source of Liulin springs, including the Yukou upstream section of Beichuan springs and Yancun upstream section of Dongchuan river. There is an increasing Na\textsuperscript{+}/Ca\textsuperscript{2+} and Cl\textsuperscript{-}/HCO\textsubscript{3}- concentration trend along the recharge-runoff-discharge pathway. Groundwater in karst aquifer will react differently depending on the mineral content (calcite, dolomite and gypsum, etc.). To further explore the dissolution effect of these mineral facies and karst groundwater, calcium in groundwater could be divided into two types: calcium from non-gypsum and calcium from non-carbonate rocks. Assuming that all the SO\textsubscript{4}^{2-} in the groundwater of the study area is derived from gypsum dissolution, calcium from non-gypsum sources are equal to [Ca\textsuperscript{2+}]-[SO\textsubscript{4}^{2-}] mmol/L, which is the total concentration of Ca\textsuperscript{2+} minus the part of Ca\textsuperscript{2+} balanced out by SO\textsubscript{4}^{2-}. Similarly, calcium from non-carbonate rocks is equal to the total Ca\textsuperscript{2+} minus the part of Ca\textsuperscript{2+} balanced out by HCO\textsubscript{3}-, namely [Ca\textsuperscript{2+}]-
0.33[HCO$_3^-$] mmol/L. The chemical reaction equation is as follows: CaCO$_3$ + CaMg(CO$_3$)$_2$ + 3CO$_2$ + 3H$_2$O = 2Ca$^{2+}$ + Mg$^{2+}$ + 6HCO$_3^-$.

Figure 3a shows that samples from the recharge area generally plot between the relationship line of 1:2 and 1:4, samples from the runoff area plot near the relationship line of 1:4, and samples from the discharge area plot below the relationship line of 1:4. In Figure 3b, we can see samples from the recharge area plot near the relationship line of 1:6, samples from the runoff area plot between the relationship line of 1:4 and 1:6, and samples from the discharge area plot near the relationship line of 1:4. It can be seen from Figure 4 that samples from the recharge area plot on the relationship line of 1:1, and samples from the runoff and discharge areas plot below the relationship line of 1:1 but do not deviate far.

**Discussion**

**The Solute Source of Karst Groundwater**

The solute source and the processes that generated the observed composition of water can be revealed by the dissolved species and their relationships with each other (Hussein, 2004; Su et al., 2009). The main ion component of karst water in Liulin springs shows that HCO$_3^-$ and Ca$^{2+}$ are the dominant ions in all the areas, which indicates the control effect of carbonate minerals (such as limestone and dolomite) dissolution. Along the flow path, SO$_4^{2-}$ and Na$^+$ increase in the runoff and discharge areas. The increase of SO$_4^{2-}$ may come from sulfate evaporite mineral dissolution (such as gypsum), sulfuric acid from sulfide oxidation, or atmospheric acid deposition. The latter two may be involved in the process of carbonate mineral dissolution by carbonic acid (Lang et al., 2005). The chemical reaction is as follows:
FeS₂ + 15/4O₂ + 7/2H₂O = Fe(OH)₃ + 2SO₄²⁻ + 4H⁺
2CaₓMg₁₋ₓCO₃ + H₂SO₄ = 2xCa²⁺ + 2(1-x) Mg²⁺ + SO₄²⁻ + 2HCO₃⁻

Ratio of [Ca²⁺+Mg²⁺] and HCO₃⁻ of the study area is between 1.05 and 1.80, which can’t be explained by pure carbonate dissolution. At the same time, the concentrations of SO₄²⁻ and [Mg²⁺+Ca²⁺ - HCO₃⁻] are highly correlated with correlation coefficient of 0.901 (Figure 5b). Therefore, part of the Mg²⁺ and Ca²⁺ dissolved in the groundwater in the form of sulfate indicates that karst groundwater in Liulin springs was influenced by sulfate evaporite mineral dissolution. The positive correlation of SO₄²⁻ and Ca²⁺ as well as Mg²⁺ demonstrates that the dissolution of sulfate evaporite minerals, such as gypsum or anhydrite contributes greatly to the origin of solutes of karst groundwater. A large amount of Ca²⁺ and SO₄²⁻ ions produced by gypsum dissolution is likely due to the ion effect in the gypsum or anhydrite dissolution (CaSO₄ ⋅ 2H₂O ↔ Ca²⁺ + SO₄²⁻ + 2H₂O), which brought about the increase of Mg²⁺ concentration with that of SO₄²⁻ increase. The Na⁺ concentration increased rapidly and even exceeded the Ca²⁺ concentration indicating that the chemical composition of karst groundwater is not only influenced by carbonate dissolution, but also affected by evaporite (halite) dissolution.

The water chemistry triangle is a common means to explore the relationship of conventional ions and lithology of groundwater. Generally, when using the cation triangle, products of evaporite mineral dissolution plot at the end of Na⁺ + K⁺ line. Limestone leachates plot at the Ca²⁺ end, and dolomite leachates plot in the middle of the Mg²⁺ and Ca²⁺ line. When using the anion triangle, dissolved pure carbonate substances plot near HCO₃⁻. Products of evaporite mineral dissolution should plot between Cl⁻ and SO₄²⁻ (Pu et al., 2011). Chemical composition of groundwater is controlled by lithology (Li et al., 2006). Plotting chemical analysis resulting on the water chemistry triangle shows that groundwater flowing through aquifers is affected by carbonate and evaporite dissolution. Cations representing recharge areas distribute along the Ca²⁺ and Mg²⁺ line, closest to Ca²⁺, while anions mainly plot near HCO₃⁻. Therefore, pure carbonate dissolution is the dominant influence on the chemical composition of karst groundwater in Liulin springs, including limestone and dolomite minerals. Runoff and discharge samples scatter resulting in cations plotting along the Ca²⁺ -Na⁺ line and anions distributing along HCO₃⁻ -SO₄²⁻ line. Consequently, the influence of evaporite minerals including sulfate and halite can’t be ignored except for carbonate minerals.

The relationship of Mg²⁺/Ca²⁺ and Na⁺/Ca²⁺ molar ratios for the groundwater are illustrated in Figure 5a. The Mg²⁺/Ca²⁺ ratios vary from 0.35 to 0.62, with all the values less than 1, while Na⁺/Ca²⁺ ratios range from 0.19 to 2.12, with most of the values less than 1. This indicates an abundant Ca²⁺ in groundwater. The equilibration of groundwater simultaneously with calcite and dolomite under room temperature gives an ideal molar Mg²⁺/Ca²⁺ ratio of about 0.8 (Appelo and Postma, 1993). Thus, maybe the solubility disequilibrium resulted in a high degree of variability with respect to Mg²⁺/Ca²⁺ molar ratios. It is then reasonable to consider that limestone with low Mg²⁺/Ca²⁺ and Na⁺/Ca²⁺ ratios, and dolomite with a high Mg²⁺/Ca²⁺ ratio and a low Na⁺/Ca²⁺ ratio are the main end-members controlling the variations.

**Figure 5.** The relationships among ions in Liulin springs.
in chemical composition of these spring waters. Molar Mg$^{2+}$/Ca$^{2+}$ ratio of karst water is below the 0.8 ratio line. Thus, it likely was related with the exchange reaction among calcite, dolomite and sulfate minerals. Molar Mg$^{2+}$/Ca$^{2+}$ ratio is low in groundwater influenced by limestone dissolution, and high Mg$^{2+}$/Ca$^{2+}$ molar ratio of groundwater may dissolve from dolomite aquifers. The sulfate mineral has much higher solubility than calcium, and gypsum is far from saturated. So, water has a strong leaching effect on gypsum rock, in which large amounts of Ca$^{2+}$ and relatively few Mg$^{2+}$ are produced making the Mg$^{2+}$/Ca$^{2+}$ molar ratio decrease. Coming down to the partition of waters, Na$^+$/Ca$^{2+}$ molar ratio of karst water in recharge areas is substantially less than 1, and Na$^+$/Ca$^{2+}$ molar ratios of karst water in runoff and discharge areas are mostly greater than 1. According to Mg$^{2+}$/Ca$^{2+}$ and Na$^+$/Ca$^{2+}$ molar ratios of different dissolution media, karst aquifers in recharge areas are mainly controlled by carbonate minerals (such as limestone and dolomite) dissolution, while runoff and discharge areas are mainly influenced by evaporite minerals.

**Dissolved Mineral Facies of Karst Groundwater**

Chemical composition of the groundwater reflects regional hydro geochemical processes and geological history (Shen, 1993; Pu et al., 2013), which ultimately boils down to the water-rock interaction between groundwater and the mineral facies of aquifers it flows through. Therefore, the dissolved mineral facies of groundwater can be explored by analyzing the chemical composition of groundwater.

Although the dynamics factors and common ion effect wasn’t considered, dissolution between mineral facies and groundwater could be proven to some extent. If [Ca$^{2+}$-SO$_4^{2-}$]/HCO$_3^-$=1:2, it indicates congruent dissolution of calcite; if [Ca$^{2+}$-SO$_4^{2-}$]/HCO$_3^-$=1:4, it shows congruent dissolution of dolomite (Wang et al., 2006). Samples from recharge areas plot basically along the relationship line near 1:4 and 1:6, indicating a common dissolved effect of calcite and dolomite. Most samples from runoff areas plot near the relationship line of 1:4, indicating that Ca$^{2+}$ and HCO$_3^-$ provided by dolomite dissolution are dominant. Samples from discharge areas basically plot below the relationship line of 1:4, indicating gypsum source of Ca$^{2+}$ in addition to calcite and dolomite according to the hydrogeological conditions of Liulin springs. If Mg$^{2+}$/HCO$_3^-$ = 1:4, the HCO$_3^-$ is completely provided by dolomite dissolution (Wang et al., 2006); If Mg$^{2+}$/HCO$_3^-$ = 1:6, the molar ratio of calcite and dolomite is 1:1 (CaCO$_3$ + CaMg(CO$_3$)$_2$ + 3CO$_2$ + 3H$_2$O = 2Ca$^{2+}$ + Mg$^{2+}$ + 6HCO$_3^-$). Samples from recharge areas plot near the 1:6 relationship line, indicating HCO$_3^-$ ion is provided jointly by dolomite and calcite dissolution. Samples from runoff areas plot between the 1:4 and 1:6 relationship lines, indicating HCO$_3^-$ ion content is provided by dolomite and calcite together, but dolomite contributed more. Samples from discharge areas plot mainly on the 1:4 relationship line, indicating the lead role dolomite dissolution plays. Above all, dolomite and calcite are the main dissolved minerals influencing chemical components of karst groundwater in Liulin springs, and the dissolution effect of dolomite is more pronounced than that of calcite.

The dissolution effect of gypsum can be obtained from Figure 4. Samples from recharge areas plot on the relationship line of 1:1 and most samples from runoff and discharge areas fall below the relationship line of 1:1. Therefore, it can be inferred that gypsum is involved in water-rock interaction of recharge areas and also dissolved in runoff and discharge areas, but contributed little.

In summary, the main dissolved mineral facies of karst groundwater in Liulin springs include dolomite, calcite and gypsum. Dolomite has become the most important dissolved mineral facie, followed by calcite and gypsum.

**Conclusions**

1. Stoichiometry showed that HCO$_3^-$ is always the dominant anion in chemical composition of karst groundwater, while there are differences in the secondary anion. The average concentration of SO$_4^{2-}$ is greater than that of Cl$^-$ in recharge areas, and it is opposite in runoff and discharge areas. Karst groundwater had the highest Ca$^{2+}$ and Na$^+$ as the major cations, accounting for more than 75% of the total cations, which also shows a certain difference among the zoning. The average concentration of Ca$^{2+}$ is greater than Na$^+$ in recharge areas, and it is opposite in runoff and discharge areas.

2. Karst groundwater in Liulin springs flowing through the aquifers was affected by dissolved carbonates and evaporites, of which carbonates dominate in recharge areas and both carbonates and evaporites dominate in runoff and discharge areas.

3. The main mineral facies of karst groundwater in Liulin springs include dolomite, calcite, gypsum and halite. Dolomite is the most important dissolved mineral, followed by calcite and gypsum.
in recharge areas as well as calcite, gypsum and halite in runoff and discharge areas.

**Acknowledgments**

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**References**


Pu T, He YQ, Zhang T. 2013. Isotopic and geochemical evolution of ground and river waters in a karst dominated geological setting: A case study from Lijiang basin, South-Asia monsoon region. Applied Geochemistry (33): 199-212.


KARST PALEO-COLLAPSES AND THEIR IMPACTS ON MINING AND THE ENVIRONMENT IN NORTHERN CHINA

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Abstract
Karst paleo-collapses are unique collapse structures widely found in the coal measures of northern China. Their geometric dimensions and internal properties indicate that a compound dissolution of carbonate and gypsum rocks may contribute to their formation. When these collapses are permeable to groundwater flow, they hydraulically connect the coal seams and the karst aquifers, which is a pre-requisite for water inrushes during coal mining. Over the last 40 years, water inrushes through these collapses have caused fatalities, economic losses, and degradation in the environment in northern China. Determination of locations and hydrogeological characteristics of the karst paleo-collapses are essential in preventing water inrush incidents through them. Advanced geophysical prospecting, aquifer testing and accompanied dye tracing are effective approaches to investigating these structures.

Introduction
Cover collapses and rock collapses are two basic types of collapse structures (Newton, 1987). Research of historical records shows that most collapses are cover collapses, leading to sinkholes (Beck, 1991), where unconsolidated sediments overlying carbonate bedrock move downward through dissolution openings into a network of dissolved void space or a single large cavity, capable of accommodating the sediments. The cover collapse sinkholes have been extensively studied and well documented due to their recognized sensitivity to human activities and their impacts on engineering works and the environment (Newton, 1987; Beck, 1984; Beck and Wilson, 1987).

The occurrence of bedrock collapses is rare compared with that of cover collapse sinkholes (Newton, 1987). Most of the human-induced rock collapses are related to dam construction in mountainous areas, where water potential differences between recharge and discharge zones are great (Yuan, 1987). Continuous and repeated change in air and water pressure (the hammer effects) in karst conduits are the main factors triggering rock collapses. In southern China, over 20% of reservoirs built in areas of bare karst failed to retain water due to the rock collapses at their bottoms (Zou, 1994). Rock collapse dolines are rarely seen in the act of collapsing (White, 1988). None of the reported 1,700 sinkholes in Florida, United States has been confirmed as cave roof collapse (Beck, 1991).

However, these sinkholes are newly developed; on a geologic time-scale, cave-roof collapses are widely distributed as well (Sangster, 1988; Li and Zhou, 1989; Vegter and Foster, 1992). Rock collapses developed in the past geological history are referred to as paleo-collapse structures or paleo-collapse breccia pipes or columns. Depending on present hydrogeological conditions in areas of paleo-collapses and the internal properties of these structures, they can function as weak geological media for preferential groundwater flow and contaminant transport (Zhou, 1997; Benson and others, 1991). Inactive paleo-collapse structures can be reactivated by human activities such as dam construction, mining underground minerals, pumping groundwater, and development of landfills. They may also be reactivated by natural events such as earthquakes and neo-tectonic activities.

Karst Paleo-Collapses in Northern China
Geology and climate are two major factors affecting the development of karst features. Usually, there have been several phases of karstification during geological history. In China, for example, the most striking paleo-karst is found in (1) the Caledonian Orogeny unconformity between the Ordovician limestone and middle carboniferous series in northern China; (2) the unconformity between the Lower Permian limestone and the Upper Permian Series in southern China; (3) the pre-
Jurassic karstification in the Sichuan basin; and (4) the pre-Tertiary buried karst in northern China.

The effects of paleo-karst can be observed in the various karst features of the carbonate aquifers and the yields of oil and gas reservoirs. They can also be seen in the interrelationships between important mineral deposits and karst aquifers which pose difficult problems in mining engineering. Paleo-collapse structures are one of the most important types of paleo-karst features (Zhou, 1997).

Paleo-collapses are widely distributed in northern China, especially in the provinces of Shanxi, Hena, and Hebei, as listed in Table 1. They have been found in over 50 coalfields and their total number exceeds 3,000 with an intensity of up to 70 collapses/km². In some stopes paleo-collapse structures comprise 30% of the total mined areas. They are recognizable in plan view as patches of breccia with miscellaneous lithological composition, generally derived from overlying strata and completely enclosed in lower bedrock. Diameters range from tens to hundreds of meters with the largest measuring 1,050 m. In profile, they take the form of vertical cylinders several hundred meters deep. No bedding is apparent inside these structures and the different rocks are intermixed and poorly sorted. They generally contain higher proportions of displaced blocks and the adjoining strata are offset as a result of dissolution-collapse. Fragments tend to be sharply angular, typically rotated, show little sign of wear and appear to have dropped from their original stratigraphic position. Figure 1 shows some common profiles of paleo-collapses reported in northern China. While the majority of the paleo-collapses are buried underground, some are exposed to surface and expressed as depressions.

Usually the infill materials in the paleo-collapse structures are tabular 5–40 cm angular fragments which display random orientation. Sides are subparallel and contacts between host and fills are sharp and irregular. In most cases, the matrix consists of clastic sediments without cement or mineralization. These structures are generally perpendicular to the ground surface. However, they can become inclined as a result of tectonic movements but remain perpendicular to the surrounding strata. Voids may be present at the top of the structures and drill bits can drop noticeably during borehole drilling. Closed depressions sometimes form in the surficial sediments without any apparent fluctuation in water level or any construction works taking place.

### Table 1. Paleo-collapses reported in northern China’s coalfields.

<table>
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<tr>
<th>Province</th>
<th>Location</th>
<th>Number of paleo-collapses</th>
<th>Max. size of collapses (m)</th>
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148  NCKRI SYMPOSIUM 5  14TH SINKHOLE CONFERENCE
in the underlying karstified Ordovician limestone, and the lithological character of the breccia gives a strong impression of collapse of upper strata. Gypsum in the Ordovician limestone is recognized as playing an important role in paleo-collapse formation (Li and Zhou, 1990). Sulfate-reducing bacteria (SRBs) may help initiate and accelerate the dissolution process of the

The paleo-collapses found in northern China are of different hydrogeological types depending primarily upon the lithology of their internal rock blocks, extents of weathering and cementation, and the secondary structures associated with the collapses. Based on the exposed paleo-collapses from drillings and excavations, the paleo-collapses can be permeable, impermeable and poorly permeable. In different paleo-collapses or different locations of the same paleo-collapse, the rock blocks may have different hydrogeological properties. The permeable paleo-collapses consist of weathered rock blocks but they are typically unconsolidated and not cemented. The impermeable paleo-collapses consist of weathered rock blocks that are cemented by weathered shale and mudstone. The poorly permeable paleo-collapses consist of partially cemented rock blocks with secondary fractures around the border of the collapses.

Table 2 summarizes the hydrogeological properties of some of the paleo-collapses.

Figure 2 shows profiles of the three types of paleo-collapses. All of the three paleo-collapses are exposed in the Fangezhuang mine. Paleo-collapse No. 9 is permeable; paleo-collapse No. 2 is impermeable; and paleo-collapse No. 1 poorly permeable.

Formation Mechanisms of Paleo-Collapses

The origin of this type of collapse is not fully understood, but the bottom of a paleo-collapse structure is usually
yielded a temperate, but dry forest/grassland assemblage when the collapses occurred (Lu and Copper, 1997). In active karstification situations gypsum can dissolve very rapidly and caves can quickly enlarge, or amalgamate, become unstable and collapse. When this occurs collapse columns or breccia pipes can propagate upwards and cause surface collapse. The sizes of the breccia pipes and collapses relates directly to the thickness and strength of the gypsum deposits. The more massive, thick and homogeneous the deposit, the larger are the caves that can be supported and consequently the larger the collapses and breccia pipes. When the gypsum between the collapse columns also dissolves the whole sequence effectively founders onto an irregular bed of breccia.

The nature of the underlying and overlying strata adjacent to the gypsum also has a profound influence on the way gypsum karst develops. In sequences where the adjacent rocks are largely impervious the karst can only develop along the joints in the gypsum. In phreatic situations, where the gypsum is underlain, or overlain, by porous dolomites or sandstones, water enters the gypsum in a diffuse way. When this happens dissolution progresses on many fronts and maze cave systems can develop, such as those observed in the Ukraine (Andrajchouk and Klimchouk, 1993). In this situation much of the dissolution occurs at the contacts with the adjacent strata which may also be karstified. In the maze caves the passages range from joint-controlled networks of small stable caves with little dissolution, to networks of actively dissolving conduits with large and unstable cavities; these commonly have subsidence associated with them.

Where thin beds of gypsum are interbedded with porous beds of limestone or dolomite (or sandstone), the potential for the hydration of anhydrite and the dissolution of the resultant gypsum is large. When anhydrite hydrates to gypsum, the expansion can cause brecciation of the mineral, brecciation of the adjacent rock, and injection of fibrous gypsum veins; such deposits are present in the Ordovician Fengfeng Formation. When dissolution of the gypsum occurs, cavities result followed by collapse and the formation of layered breccia deposits. Because dissolution is not uniform throughout the deposits, caves can also develop allowing further erosion of the collapsed material. Breccia pipes or collapse columns

Shanxi province is in a semi-arid environment undergoing uplift. Palaeokarst, such as that originally developed under phreatic conditions in the Ordovician Fengfeng Formation, has been uplifted above the water table and preserved. In the Fengfeng Formation the dissolution of the gypsum has caused the formation of extensive breccia beds with associated breccia pipes or collapse columns that have propagated upwards through the active coalfields (Li and Zhou, 2006). Preservation of gypsum karst is likely when climatic conditions change causing a reduction in the availability of water to the natural systems. In this way the gypsum karst of southwest Oklahoma (Johnson, 1990) was preserved. Pollen/spore samples from underground deposits have
can develop in this type of strata, but may be of smaller dimensions than those developed in the more massive gypsum rocks. This is due to the presence of numerous discontinuities in the bedded sequence and the piecemeal failure of the thin-bedded rock, which will span smaller distances. Where the sequence comprises interbedded gypsum and insoluble, or less soluble, rocks a breccia, largely composed of the insoluble components, can develop along the level of the former gypsum beds. This is the situation that formed breccias at numerous levels, and collapse columns of breccia penetrating much of the sequence, in the Fengfeng Formation of Shanxi Province.

Where gypsum is in contact with adjacent carbonate rocks, the water chemistry associated with the dissolution phenomenon is different to that for gypsum alone, or for limestone alone. The presence of gypsum together with carbonate rocks will usually cause a compound karstification effect. Waters rich in calcium carbonate can aggressively dissolve gypsum and simultaneously deposit calcite (Wigley, 1973). When this occurs the breccias caused by the dissolution of the gypsum may be cemented. Carbonate-cemented breccias, possibly formed by this mechanism, occur in the Fengfeng Formation. Gypsum dissolves easily in flowing water and increases the amount of sulphate in the water. However, the solubility of gypsum in calcium carbonate-rich waters may be decreased by the common ion effect. Conversely, the presence of sodium, magnesium and chloride ions (possibly derived from interbedded dolomites and halite beds) can enhance the dissolution of gypsum. Experimental work shows that the presence of sulphate in the water increases the dissolution rate for dolomite. For water with an $SO_4^{2-}$ content of 1 mg/l the dissolution amount for dolomite was 1.67 mg/l while that for limestone was only 0.94 mg/l (Lu and Cooper, 1997). The result of this groundwater chemistry on karstification is that very intense and pervasive leaching of the carbonate deposits, especially dolomites, can occur resulting in a honeycomb structure of very little strength. The dissolution of the gypsum, when it is complete, can leave just an insoluble residue, commonly of silt or clay.

After the gypsum dissolved and was removed by groundwater, numerous cavities or fractures were left behind. Solution of the gypsum layers and continuous dissolution of the limestone eventually led to the collapse of the overlying strata. Development of a paleo-collapse is envisaged as a more or less continuous process which progresses upward from an initial conduit (cave) until an equilibrium pressure arch configuration is attained. This occurs when a collapse reaches a lithologic unit of sufficient strength, or the cavity is completely filled with breccia and thus becomes self-supporting.

**Impacts of Paleo-Collapses on Mining and the Environment**

The absence of coal seams and the sudden inrush of karst water from the Ordovician limestone have been encountered in the mines of the Permo-Carboniferous coalfields of northern China. These events are due to the presence of paleo-collapse structures. Table 2 lists 15 water inrushes reported from the mines, including the largest inrush in the world, which occurred in Fangezhuang Mine in 1984. Karst water gushed into the mine at a flow rate of 34 m$^3$/s at a depth of 313 m below sea level (bsl). The surface level is 27 m above sea level (asl). The whole mine was flooded within 21 h and as a result, the regional water table in the Ordovician limestone dropped from 5.94 m asl to 111.09 m bsl. The cone of water depression covered 84 km$^2$ with a north-south axis of 25 km. The fall in the level of the water table in the Ordovician limestone caused serious problems for local residents. These included the drying up of their water supply wells, contamination of the groundwater and the formation of new sinkholes. The water inrush led to the development of 17 cover collapses, with resulting sinkholes ranging in diameter from 2.5 to 3 m and with depths of 3 to 12 m.

Three basic conditions are required to cause a water inrush. These are the presence of:

- water-bearing karst conduits or a nearby water body;
- water-permeable internal structures in the paleo-collapses; and
- water pressure differences between the karst aquifer and the working area.

The relative location of a mine to the active flow zone or karst conduits in karst aquifers determines the amount of water that can flow into the mine. In the presence of a large water-pressure difference, hydrofracturing will facilitate the upward flow of Ordovician karst
water into mines. Apparently inactive paleo-collapse structures or those which have been cemented can be reactivated by activities such as mining, pumping water, dam construction, and landfill development. They may also be triggered by natural events such as neotectonic movements and earthquakes. Table 3 provides some examples and Figure 4 shows three scenarios where paleo-collapse structures can lead to water inrushes. Mining drifts do not have to intercept paleo-collapse structures directly to cause a geohazard but may instead intercept faults or fractures connected to them. However, once a water inrush occurs and significant water flows into the workings, the whole mine or quarry may become flooded. In cases where different aquifers, several hundred meters apart, become hydraulically connected, the impacts on safety, economy, and the environment can be alarming.

Detection and Remediation of Hazardous Paleo-Collapses

Because the impacts of paleo-collapse structures on the environment and engineering works are serious, it is essential to locate their positions before they are actually exposed. A variety of methods, including geophysical and geochemical ones, are used; exactly which depends on site conditions. In northern China, a combination of pumping tests and dye-tracing tests proved effective in locating two paleo-collapse structures in Fengfeng Mine No.4. The major geological strata in Fengfeng Mine No.4 are shown in Figure 5.

The Carboniferous thin-bedded limestone (Daqing limestone) and the Ordovician limestone are the two major aquifers. The average thickness of the Daqing aquifer is some 5.5 m. Sixty-nine percent of the boreholes sunk in this aquifer, within the mining area, discharge over 0.017 m$^3$/s. Apparently inactive paleo-collapse structures or those which have been cemented can be reactivated by activities such as mining, pumping water, dam construction, and landfill development. They may also be triggered by natural events such as neotectonic movements and earthquakes. Table 3 provides some examples and Figure 4 shows three scenarios where paleo-collapse structures can lead to water inrushes. Mining drifts do not have to intercept paleo-collapse structures directly to cause a geohazard but may instead intercept faults or fractures connected to them. However, once a water inrush occurs and significant water flows into the workings, the whole mine or quarry may become flooded. In cases where different aquifers, several hundred meters apart, become hydraulically connected, the impacts on safety, economy, and the environment can be alarming.

Table 3. Case histories of water inrushes through paleo-collapses in northern China.

<table>
<thead>
<tr>
<th>Mine</th>
<th>Date</th>
<th>Flow rate (m$^3$/s)</th>
<th>Description</th>
<th>Hazard</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tongyie Mine, Anyuang</td>
<td>1965</td>
<td>0.39</td>
<td>A water inspection borehole drilled into a paleo-collapse structure from a drift. The initial water flow rate was 0.5 m$^3$/min-1 and water flow rate increased to 23.3 m$^3$/min-1. An exploration borehole revealed 17 cavities within 50 m of the collapse with the maximum bit-drop of 2.59 m.</td>
<td>The whole mine was flooded</td>
</tr>
<tr>
<td>Lifeng Mine, Jiaozuo</td>
<td>1967</td>
<td>2</td>
<td>Karst developed very well in the area. Intensive mine water drainage reactivated the paleo-collapse structure.</td>
<td>A working stope was flooded</td>
</tr>
<tr>
<td>Fagezhuang Mine, Kailuan</td>
<td>1978</td>
<td>1</td>
<td>Water flowed into the mine from the sandstone, which is 160 m above the underlying Ordovician limestone. A sluice gate was constructed to isolate the water inflow area and a 0.2-m fracture was revealed, connected with a paleo-collapse structure.</td>
<td>Part of a drift and a working stope (70,188 m$^3$) was flooded</td>
</tr>
<tr>
<td>Fagezhuang Mine, Kailuan</td>
<td>1983</td>
<td>0.23</td>
<td>A small fault with displacement of 0.2–0.5 m was intercepted by a working stope. Karst water flowed through a paleo-collapse structure into the fault and then to the working stope.</td>
<td>The working stope was flooded</td>
</tr>
<tr>
<td>Fagezhuang Mine, Kailuan</td>
<td>1984</td>
<td>34</td>
<td>This is the biggest water inflow incident in the world. The mining coal seam was 180 m above the Ordovician limestone but they are connected by a paleo-collapse structure. The reactivation of the paleo-collapse may be associated with a recent earthquake in this area. Grouting boreholes revealed that the top of the sinkhole was unfilled with sediments. Due to the water inrush, 17 cover sinkholes were induced on the surface.</td>
<td>The whole mine was flooded and the adjacent three mines were threatened</td>
</tr>
<tr>
<td>Huoxian Mine</td>
<td>1967</td>
<td>0.13</td>
<td>Karst water flowed into an excavating drift through a paleo-collapse structure and a connecting fault.</td>
<td>The drift was abandoned</td>
</tr>
<tr>
<td>Huaxian Mine</td>
<td>1984</td>
<td>0.06</td>
<td>Karst water from a paleo-collapse structure flowed into surrounding fractures and then into the horizontal drift in the Ordovician limestone.</td>
<td>The drift was abandoned</td>
</tr>
<tr>
<td>Wucun, Huixian</td>
<td>1999</td>
<td>0.67</td>
<td>Karst water from a paleo-collapse structure flowed into the mining area.</td>
<td>The panel was flooded</td>
</tr>
<tr>
<td>Dongpang, Xingtai</td>
<td>2003</td>
<td>19.5</td>
<td>Karst water from a paleo-collapse structure flowed into the mining area.</td>
<td>The mine was flooded</td>
</tr>
</tbody>
</table>
adsorbed electric-magnetic waves. Pumping test results together with dye tracing confirmed the findings by the well-logs and enabled the paleo-collapse structures to be located. Four pumping tests were conducted in the Daqing aquifer at the rates of 0.2, 0.42, 0.33, and 0.44 m$^3$/s, respectively. Piezometric pressures in both aquifers showed corresponding fluctuations. The pumping in the Daqing aquifer caused a cone of depression in the Ordovician limestone. During each pumping test, a conservative fluorescent dye was introduced into a borehole in the Ordovician limestone and was collected at all accessible boreholes and discharge points in the Daqing aquifer. The results of the dye tracing are shown in Figure 5. In all cases, the dye first appeared in boreholes in areas A and B. The straight-line velocities were calculated based on the distances and the dye travel times. Dye traces 1, 2, 3, and 4 had velocities to area A of 0.62, 0.43, 0.08, and 6.5 m/min, respectively; and to area B of 0.007, 0.0087, 0.0017, and 0.19 m/s, respectively. The high but contrasting velocities from different traces highlight the rapid flow and strong heterogeneity in the Ordovician limestone aquifer. This in turn assisted in delineation of the possible locations of water-conducting paleo-collapse structures. Subsequently, 16 grouted boreholes were drilled into the suspected paleo-collapse structures and the adjoining Daqing aquifer so as to seal-off and plug the vertical passageways (Figure 6). As a result, the amount of water flowing into the mine decreased from 41 to 3 m$^3$/min, and the piezometric pressure in the Daqing aquifer dropped significantly.

Figure 5. Locating paleo-collapse structures by dye tracing in Fengfeng Mine No.4 (Zhou, 1997).

Figure 4a–c. Scenarios where karst water from Ordovician limestone flows into workings through paleo-collapse structures. 

a Through paleo-collapse-connected fissures; 
b through paleo-collapse-connected faults; 
c through paleo-collapses (Zhou, 1997).

m$^3$/s with the maximum discharge of 0.25 m$^3$/s. Forty days’ continuous pumping at 0.68 m$^3$/s did not drain the aquifer; instead, its piezometric pressure remained stable. Observation wells in the Ordovician limestone recorded a drop in piezometric level. Therefore, it was concluded that the Daqing aquifer was receiving water from the Ordovician limestone aquifer and subsequent analyses indicated both aquifers to have the same chemistry. Geologic structure analysis and exploratory drilling suggested that lateral inflow from out-side the mine was impossible due to impervious faults (Figure 5). Thus, the connecting link between the two aquifers must be vertical. Paleo-collapse structures are one possible explanation for this, and indeed over twenty such structures have been intersected in the mined area. Surface geophysical methods were unsuccessful in locating the paleo-collapse structures due to the big depths involved (Li and Zhou, 1990). However, the well logs indicated that boreholes at positions A and B in Figure 5 strongly
as shown in Figure 6. The investment into the grouting operation was recovered from the savings in dewatering cost within 8 months.

Conclusions
Paleo-collapse structures are common karst features when viewed on a geologic time-scale. They can act as passage-way for groundwater flow and contaminant transport. In northern China, significant damage has been caused in mines as a result of these structures, including the largest water inrush in mining history. Three factors determine whether a paleo-collapse structure will become a water passageway and thus a geohazard:

- The paleo-collapse structure intercepts water-bearing strata;
- The water table in the water-bearing strata is higher than the bottom of the mine working;
- The internal structure of the paleo-collapse favors water flow.

The presence of paleo-collapse structures indicates a strong groundwater flow (conduit flow) in the areas in the past and does not necessarily reflect present hydrogeological conditions. Only those paleo-collapse structures that are within present conduit flow zones or in the vicinity of a water body have the potential to cause geohazards. Paleo-collapse structures may be reactivated by human activities such as dam construction, landfill development, mining and quarrying, intensive water pumping, as well as natural events such as earthquakes or neo-tectonic movements. A multidisciplinary approach including geo-physical prospecting, geochemical analyses, test drilling, pumping tests, and dye tracing is required to locate these structures. Grouting is the most effective and perhaps the only possible way of preventing paleo-collapse structures from becoming geohazards. Since grouting is expensive, a cost benefit analysis should be undertaken beforehand.

References

Figure 6. Plugging the paleo-collapse by grouting to prevent water inrush in Fengfeng Mine No.4 (Zhou, 1997).


THE SANDSTONE KARST OF PINE COUNTY, MINNESOTA

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Abstract
The glaciated, forested landscape of central Pine County in east-central Minnesota contains a series of sinkholes, stream sinks, springs and caves. The features are formed in Precambrian Hinckley Sandstone and overlying unconsolidated glacial deposits. This is a sandstone karst. The features serve the same function as in carbonate karst terrains: sinkholes and caves focus recharge into a heterogeneous subterranean flow system that discharges into springs. The Hinckley Sandstone is a quartz arenite. No carbonate grains or cements have been found in sandstone samples from the sinkhole area, nor is there evidence that calcite solution controls bedrock permeability. Three parameters appear to control the distribution of sinkholes: depth to bedrock, type of underlying bedrock, and meter-scale heterogeneity of surface sediments. The permeability structure of the Hinckley Sandstone appears to be controlled by fractures and depositional features at centimeter to meter scale. Field mapping in the area has revealed 309 karst features: 237 sinkholes, 25 stream sinks, 32 springs and 15 caves. Recent LiDAR coverage indicates that there are many more sinkholes and other karst features than the original mapping was able to locate. Interpretation of the LiDAR images is challenging because karst processes, glacial processes and human activity have all produced natural and anthropogenic closed depressions of a variety of sizes and shapes in this landscape.

Introduction
Definitions of Karst
According to White (1988), karst can be defined as a combination of process and form, where process is dominance of chemical over mechanical erosion and form is a set of distinctive geomorphic features. When solution and mechanical erosion compete, solution must be the more dominant process in order for the resulting landforms to be considered karst. Similar landforms that owe their development to other causes are often termed ‘pseudokarst’.

Others have developed broader definitions. In his treatment of some puzzling features on sandstone, Jennings (1983) defines karst as the result of “the process, solution, which is thought to be critical (but not necessarily dominant) in the development of the landforms and drainage characteristics of karst.” He defines pseudokarst as “country with resemblances to karst, which are due to other processes.”

Jennings’ definitions acknowledge that karst is more than a geomorphic landscape; it is more useful to consider karst as a system, with a definition based on universal characteristics instead of lithology. These characteristics include interconnected networks of flow paths capable of high-velocity and turbulent flow. Karst flow systems have “a specific type of fluid circulation capable of self-development and self-organization” (Klimchouk and Ford, 2000). Considering karst as a flow system focuses on the essence of what distinguishes karst flow from porous media flow.

Global Distribution of Silicate Karst
While karst and pseudokarst occur in a variety of lithologies, our focus is on nearly monomineralic silicates such as quartz arenite or orthoquartzite. Solutional karst features in these lithologies have been recognized on all continents, in a range of climates. Wray’s (1997) review paper notes silicate karst in Venezuela, Brazil, the United States, Morocco, Chad, Niger, Nigeria, Zimbabwe, South Africa, Thailand, Australia, the United Kingdom, Poland, the former Czechoslovakia, and scattered sites.
Chalcraft and Pye (1984) found the solubility of geologic opal to be up to four times lower than that of experimental amorphous SiO$_2$, so that opal hydration is not a necessary step in the solution of quartz. They also note that the solubility of all forms of SiO$_2$ can be enhanced by alginic and amino acids produced by algae. The tops of the Roraima tepuis have widespread colonies of algae and accumulations of organic material in bogs, unlike the surrounding lowlands. They present SEM (scanning electron microscope) images with features (etching of quartz grains and cements) that they attribute to solution. They argue that solution is critical to the development of the Roraima karst.

In northwestern Australia, Young (1986; 1987) describes the geomorphic features of several sandstone karsts. He addresses speleogenesis through petrographic studies to look for microscopic solution features, as well as variation (such as primary porosity at the onset of dissolution, or structural and depositional controls) that could explain the distribution of landforms. Young (1988) reports extensive etching of grains and optically continuous overgrowths in SEM images. He documents two main etching textures. The first are v-shaped notches that show strong crystallographic control. These are surface reaction controlled features, indicative of slow fluid movement. The other group is that of embayments; a single embayment can cross a grain-overgrowth boundary, which does not happen with v-notches. Embayments are flow-controlled features, indicative of higher fluid velocities (Young, 1988).

**Geological Setting of the Field Area**

The study area lies in central Pine County, which is located in east-central Minnesota. Mesoproterozoic basalts are overlain by the Fond du Lac Sandstone, which is overlain by the Hinckley Sandstone (Figure 1). Two large northeast-southwest trending reverse faults (the Douglas Fault and Hinckley Fault) have uplifted terrain to the east: in the east part of the study area, basalts are exposed at the surface, in the central part of the study area the Fond du Lac Sandstone is covered by a veneer of Hinckley Sandstone, and the west part of the study area is covered by a layer of Hinckley Sandstone up to 150 meters thick (Boerboom et al., 2002; Mooney et al., 1970). The stratigraphy and faults are associated with the northern part of the western limb of the failed Midcontinent Rift System across western Europe. Caves with depths of nearly 400 meters and lengths up to several kilometers have been documented in silicate karst regions, demonstrating a local significance simply in terms of pervasiveness and scale (Truluck, 1991; Wray, 1997). Not included in Wray’s list are sinkhole-topped mesas in northwestern New Mexico (Wright, 1964), solution pans in the Gobi Desert of Mongolia (Dzulynski and Kotarba, 1979), and the sinkholes of northeastern Minnesota discussed in this paper (Shade, 2002a).
The basins adjacent to the Midcontinent Rift System contain thick sequences of sedimentary rocks. The Fond du Lac Sandstone it may be as thick as 2 – 3 km at its eastern boundary, and thins westward, away from the rift zone. The Hinckley Sandstone is up to 500 m thick at the Hinckley Fault, and thins westward (Mooney et al., 1970). Morey (1972) interpreted the lithology and sedimentary structures of the Fond du Lac as shallow.
water delta deposits. The Hinckley Sandstone appears to be primarily composed of reworked Fond du Lac materials (Morey, 1972). It is described by Tryhorn and Ojakangas (1972) as a tan to orange, fine- to medium-grained quartz arenite that is typically about 96% quartz, well sorted and well rounded. The base of the Hinckley Sandstone is exposed in a thin layer between the Hinckley and Douglas Faults, where it contains small quantities of other lithic elements and is classified as feldspathic arenite (Boerboom et al., 2002).

The surficial geology of Pine County is dominated by glacial deposits and landforms sculpted by glacial advance and retreat. During the Wisconsinan Glaciation of the Mid- to Late Pleistocene, the Superior Lobe moved into the county from the north and at its maximum covered the entire county. The Superior Lobe retreated and advanced across Pine County at least twice (Knaeble et al., 2001).

Karst features occur over five different surficial map units (Patterson and Knaeble, 2001). All are unconsolidated Pleistocene deposits in the Sandstone and Aksov Lookout Tower phases (or ice margins) of the Superior Lobe. Surficial units where sinkholes have been mapped include Qsgs, which is described as “sandy glacial sediment”, Qsgf, which is described as “silty and clayey glacial sediment”, Qssi, which is dominantly sand and gravel, described as “sorted sediment proximate to ice”, and Qssm, described as “sandy glacial sediment” (Patterson and Knaeble, 2001).

**Methods**

**Field Mapping**

In the first phase of this work karst features were found by field mapping, both by systematic walking surveys and by talking to local residents. All reported features were field verified. Locations were mapped by GPS, and entered into a GIS database. The karst survey initially focused on Partridge Township, and later extended into the townships of Bruno, Finlayson, and Sandstone. The entire area underlain by Hinckley Sandstone was not surveyed. Every sinkhole in the surveyed areas has not been located; additional features will certainly be discovered both in these area and other parts of Pine County.

**Sinkhole Excavations**

Several sinkholes were excavated in order to understand how they formed. The features were selected based on setting, size and morphology. Three were dug open with hand tools and two were opened by backhoe. For all five features, a trench with vertical sidewalls was excavated perpendicular to the most likely bedrock fracture orientations. Surface profile and stratigraphy were measured every 10 cm along the trench, using leveled strings as a vertical reference. Trenches were excavated as deep as possible within the constraints of safety, time, and the water table. In one case, we used a truck-mounted Giddings Soil Probe to make a series of cores in and around a sinkhole to better define stratigraphic relationships that were too deep for hand excavation.

**LiDAR Analysis**

A second phase of mapping is currently underway. The recent availability of 1m resolution LiDAR data for Pine County allows for more systematic and complete mapping of karst features. We are visually scanning shaded relief DEMs (Digital Elevation Models) at varying scales down to about 500:1. Locations of potential sinkholes are recorded in a GIS environment in the Minnesota Karst Features Database, maintained by the Minnesota Geological Survey (Tipping et al., 2015).

The locations of sinkholes identified by Shade (2002a) are compared to newly mapped potential karst features to help determine how to differentiate sinkholes from other closed depressions present in Pine County on the DEMs, such as glacial and anthropomorphic features.

**Petrology**

The permeability structure of the Hinckley Sandstone was investigated at the outcrop, hand sample, and microscopic scales. Outcrop observations were made by observing depositional features and measuring structural features in outcrops along the Kettle River. Hand samples were collected from one outcrop and oven dried for several days at 60°C. Cubes measuring approximately 4 cm per side were placed in a vacuum chamber and impregnated with colored thin section epoxy. After the epoxy cured, blocks were cut open to assess the penetration of epoxy. Finally, sand grains from the hand samples were imaged with a scanning electron microscope (SEM) to observe grain surface textures.

**Results**

**Features Found by Field Mapping**

In the initial phase of this project, field surveys located 309 karst features, which includes 237
sinkholes, 25 stream sinks, 32 springs, and 15 caves as shown in Figure 2 (Shade et al., 2002b, 2002c). The sharp boundary along the southeast margin of the sinkhole array appears to be an actual boundary of the occurrence of sinkholes. That boundary roughly corresponds to the Hinckley Fault. The edges of the array to the north and west are undefined, and reflect only the limit of field surveys.

**Excavation of Selected Sinkholes**

We excavated five features, three with hand tools and two by backhoe. Four of the excavated features were sinkholes, formed by active transport of material into the subsurface by water. The fifth feature appeared to be of glacial origin. The cross section diagrams from two of the sinkholes are presented here.

Sinkhole D144 is part of a cluster of sinkholes located in Banning State Park, on top of the bedrock surface on the east of the Kettle River gorge. When found, it was 2.5 by 2 m in diameter and 0.7 m deep. A 4-m-long by 2-m-deep trench in glacial till exposed a funnel-shaped deposit of organic-rich soil cutting vertically through the till. At the limit of excavation, a small open drain continued into the subsurface (Figure 3). Excavation by hand tools did not reach the bedrock surface, and the location of the sinkhole in the state park eliminated the possibility of mechanical excavation or soil boring as were used in three other excavations.

Sinkhole D222 is part of a cluster of sinkholes located 5 km further south along the Kettle River. This cluster is also located in Banning State Park, on top of the cliffs

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**Figure 2.** Karst feature distribution (Shade, 2002a) and bedrock faults in the study area.
east of the Kettle River. This cluster of sinkholes forms a linear northeast-southwest trend that aligns with a group of karst features on the opposite side of the river, and runs parallel to the adjacent reach of the Kettle River. When found, the sinkhole was about 6 m in diameter by 0.8 m deep. A 4.5-m-long by 2.25-m-deep trench in glacial loess and till exposed a central funnel-shaped deposit of sandstone boulders and leached organic material cutting vertically through the glacial deposits. At the limit of excavation, a small open drain continued into the subsurface. Excavation by hand tools did not reach the bedrock surface, and the location of the sinkhole in the state park precluded the use of mechanical excavation or soil boring as were used in three other excavations.

Sinkhole D355 is located south of the town of Askov, in a shallow stream bottom near municipal sewage lagoons. At the time of excavation, it was a large shallow sinkhole (10 by 12 meters in diameter by <1 meter deep) containing several smaller sinkholes. The largest was 1.5 by 2 meters in diameter by 0.5 meters deep, and was indicated by the landowner as the main drain for feature. A 9-m-long by 4.5-m-deep trench was excavated by backhoe. The trench exposed soil and stream deposits overlying glacial till, with a horizontal bedrock surface at 4.5 m depth. The stream sediments and till were cut by an irregularly shaped unit of organic material and reworked till with one branch coming from the main sinkhole drain and another issuing from the north wall of the trench. The two branches converged at a depth of about 3 meters and continued down toward an enlarged bedrock fracture in the base of the trench.

Sinkhole D127 is located in the northeastern part of the study area, in a field about 8 km east of the Kettle River. When found, it was a recent collapse where a 1.3 by 1.4 m plug of soil and grass had dropped 0.8 m. The topsoil was undercut on the north and west sides, so that the sinkhole had an actual diameter of 2 meters. A 3-m-long by 3.25-m-deep trench exposed horizontally layered glacial sediments of loess, sand, clay, and till (Figure 4). These sedimentary layers were cut by a vertical funnel-shaped deposit of loose sand similar to layer 3 in Figure 4. Excavation by hand tools did not reach the bedrock surface, and a series of soil borings indicate a weathered bedrock surface at a depth of about 4 m, and an intact bedrock surface at a depth of about 5 m. In the throat of the sinkhole the bedrock is about 0.5 m deeper, forming an asymmetrical funnel in the rock surface which narrows to a much small area, ostensibly an enlarged fracture. Here the soil boring reached 7 m depth without hitting bedrock.

Feature D326 is located in a field 0.7 km south of D127. At the time of excavation, it was a closed depression about 30 by 40 meters in diameter by 3 meters deep. A 13-m-long by 3-m-deep trench was excavated by backhoe. The trench was characterized on its north end by sand and gravel mixed with large sandstone boulders,
and hosting small deposits of sand, peat. The south end of the trench was composed of glacial till with a small deposit of clay and spare sandstone cobbles. This closed depression did not have cross-cutting stratigraphy, a collapse funnel, or an open drain as seen in the other excavated sinkholes, and it appears to be of entirely glacial origin.

**Sinkholes Found by LiDAR**

Figure 5 shows a part of the southeast quarter of section 3, T42N, R20W on the southeast side of the Kettle River in Banning State Park. This area contains a dense array of sinkholes. Forty-two sinkholes were field mapped by the first phase of this research (Shade, 2002a). About half of the sinkhole locations were adjusted to fit the sinkholes visible in the LiDAR DEM and are shown as the green triangles in Figure 5. Locations were adjusted by a few meters to tens of meters and reflect the lower accuracy of the original GPS locations in comparison with the LiDAR elevation model.

Figure 6 shows the same area with the new LiDAR located sinkholes as red triangles. Seven additional sinkholes are visible in the DEM that were missed in the field work. The smaller field-located sinkholes are not visible in the DEM.

Figure 7 shows the DEM of section 3 of T42N, R20W and includes the area shown in Figures 5 and 6. Field work reported in Shade (2002a) located 64 sinkholes in Section 3 but did not cover the entire section (the green triangles). Visual scanning of the LiDAR DEM identified 137 more potential sinkholes. Some of the new features are in areas previously surveyed, while others are outside of prior survey areas.

**Petrology**

The most available and extensive outcrops of Hinckley Sandstone are located in historical quarries and in the valley along the Kettle River. Some outcrops are formed by hard, well-indurated rock, while nearby outcrops may have rock that is soft and friable. There are noticeable variations of rock strength and weathering patterns within outcrops. Many outcrops showed enlarged fractures, enlarged bedding planes, and small (<1m

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*Figure 5.* Sinkholes from Shade (2002a) (the green triangles) superimposed on a 1m LiDAR DEM. Locations have been adjusted as per text description.

*Figure 6.* Sinkholes from Shade (2002a) (the green triangles) with potential sinkholes identified from the LiDAR DEM (the red triangles).

*Figure 7.* Sinkholes from Shade (2002a) with potential sinkholes identified from the LiDAR DEM. Sec 3 Findlayson Twp (T42N, R20W), Pine County.
diameter) conduits. The development of solutional features was not uniform in outcrop, but rather appeared to be controlled by lithology. Some sandstone beds or groups of beds had many solution features while those beds above and below had no visible solution features.

Saturation of hand samples with epoxy demonstrated a range of flow patterns. Some were uniform, with equal penetration by epoxy in all directions. However, most were not uniform, showing penetration that was retarded by a weathering crust, as well as sharply heterogeneous patterns following or avoiding bedding features, and following or avoiding tiny fractures.

Individual grains were imaged with SEM to observe grain surface textures. The SEM work was exploratory and does not represent an exhaustive survey of the Hinckley Sandstone. On the basis of preliminary investigation, sand grains from the surface of conduits seen in outcrop were characterized by extensive etching, pitting, and embayments. Grains from outcrops without obvious conduits, enlarged fractures, or bedding planes had the same types of solutional surface textures, but to a much lesser extent.

**Discussion**

Karst development in Pine County appears to be controlled by bedrock type and geologic structure. All mapped features occur over or in the thickest part of the Hinckley Sandstone, northwest of the Hinckley fault. Most features lie within five kilometers of the fault, and none have been mapped more than seven kilometers from it (Figure 2).

No sinkholes have been found in the thin layer of Hinckley Sandstone southeast of the fault, despite extensive field searching. Nor have any likely sinkholes been identified through LiDAR imagery mapping. To the northwest, there is not such a well-defined edge to the karst. This area has not been well searched, so the absence of sinkholes may be an artifact of field activity. Sinkhole formation would be less likely near the Hinckley–Fond du Lac contact northwest of the study area because the base of the Hinckley has a higher content of non-quartz material that could clog a developing flow system.

The thickness and composition of glacial sediments are also important in determining where sinkholes are likely to form. In places where the glacial drift is thin, surface water is able to move relatively quickly into the underlying bedrock. Areas of high transmissivity within the glacial drift such as sand lenses, gravel lenses or boulder concentrations also permit the surface water to move downward rapidly. Thus, areas with thin, coarse, and highly permeable drift underlain by fractured bedrock are most favorable for sinkhole development.

The sinkholes lie near the Askov Lookout Tower ice margins from the Superior Lobe of the Wisconsinan glaciation, but the causal relationship between the moraines and sinkhole distribution is uncertain. High volumes of unsaturated water resulting from glacial discharge may have enhanced solution of quartz. Discharge of glacial melt water off the front of the moraines may also have cleared fine-grained sediment from joints in the underlying bedrock. Such open joints could subsequently act as subsurface conduits over which the sinkholes could form. The movement of surface sediment into the underlying fractures/conduits is an ongoing process as demonstrated by the recent collapse of sinkhole D127 (Figure 4).

Excavation of closed depressions shows that some are true sinkholes, characterized by cross-cutting stratigraphy and active movement of surface material into the subsurface. The sinkholes occur in a range of sizes, from meters to tens of meters in diameter. In addition to sinkholes, this landscape also contains large depressions formed by glacial processes, and small depressions formed by stump holes. These features are distinguished from sinkholes because they are filled with sediments that are oldest on the bottom, youngest on the top, and without subsidence features or cross-cutting stratigraphy. This landscape also contains depressions of a range of sizes formed by human activity, such as gravel mining. These depressions are characterized by chaotic fill, and may not have any appreciable recent deposition.

Epoxy experiments indicate that the Hinckley Sandstone has a heterogeneous permeability structure at the hand sample scale. Outcrop observation shows weathering heterogeneity at the meter scale. Karst feature distribution indicates discrete high-velocity flow paths with strong structural control. SEM imaging of sand grains show surface textures attributed to solution in previous work.

Although scanning of the LiDAR DEM is just beginning, it is clear that the sandstone karst of Pine County is more
Conclusions
Central Pine County hosts a karst system developed in sandstone. Surface karst features include sinkholes, streamsinks, and springs. These surface features are connected by a system of high-velocity flow paths seen in caves and enlarged bedrock fractures. Previous field mapping was limited by access to private property, as well as the scale of area to be surveyed. New LiDAR imaging could be a powerful tool to improve karst mapping by identifying potential features. Given that sinkholes are overprinted on a glaciated terrain, features identified by LiDAR must be field verified.

Acknowledgments
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A PROPOSED HYPOGENIC ORIGIN OF IRON ORE DEPOSITS IN SOUTHEAST MINNESOTA KARST

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Abstract
From 1942 through 1968 there was an active iron ore mining industry in western Fillmore, eastern Mower and southern Olmsted Counties of Minnesota. This iron mining district was 250 miles south of, and the ores were a billion years younger than, the ores of the classic iron mining districts in northern Minnesota. The high grade iron ore was mostly goethite and hematite and occurred as near-surface relatively small pods which unconformably filled paleokarst depressions in the Devonian Spillville Formation and the Ordovician Stewartville Formation.

The source of the iron has long been cryptic. The available field and textural evidence is consistent with a hypogenic origin of these iron deposits. Before the current Mississippi River drainage system was incised, regional ground water flow systems could have emerged through the karst conduits in the Paleozoic carbonates. The waters in the deeply buried aquifers underlying this area currently are anoxic and enriched in dissolved ferrous iron and would have been more so before the entrenchment of the Mississippi River reorganized the regional ground water flow system. When that water emerged into the atmosphere the ferrous iron would have quickly been oxidized by a combination of biotic and abiotic processes producing the ferric oxide ore at the spring orifices. Numerous springs and seeps in Minnesota are currently building iron oxide deposits at their orifices.

Introduction
The presence of iron ore deposits in southern Minnesota has been recognized since Winchell and Upham’s (1884) report. These deposits occur on top of Paleozoic sedimentary rocks and are often covered by Pleistocene glacial deposits. They are distinctly separated in time and space from the major Precambrian iron ore deposits in northern Minnesota (Morey, 1998). The high grade iron ore was mostly goethite and hematite and occurred as near-surface, relatively small pods which unconformably filled paleokarst depressions in the Devonian Spillville Formation and the Ordovician Stewartville Formation. “Following extensive exploration work that was conducted in the 1930s, two companies carried out mining operations in the Fillmore County district from 1942 to 1968. Cumulative production was 8.1 million tons of iron ore” (Bleifuss, 1972 p.498).

The deposits were often adjacent to or cementing discontinuous bodies of the nominally Cretaceous Ostrander Gravels. The ore bodies were covered with a few meters of unconsolidated Pleistocene glacial drift and loess and Holocene sediments. The mining was accomplished with bulldozers, front end loaders and dump trucks. The ore was shipped by rail mainly to mills in the St. Louis area.

The iron ores are conventionally mapped as the Iron Hill Member of the Windrow Formation (Andrews, 1958). Andrews’ (1958) stratigraphic study of the Windrow Formation in the Upper Mississippi Valley, mainly southern Minnesota, southwestern Wisconsin and in northern Iowa, reviewed the literature up to 1958. Based on the literature and his own extensive work, Andrews (1958, p. 597) concluded “It seems probable that the Iron Hill member was deposited as a result of reaction of iron-charged waters with carbonate bedrock.”

Rodney Bleifuss’ PhD thesis (Bleifuss, 1966) and subsequent publication (Bleifuss, 1972) are the most definitive works on the origin of the iron ores of southeastern Minnesota. Bleifuss’ thesis work was conducted during the active phase of the iron mining. He observed, studied, and documented many of the iron ore
The Ore Bodies

Location

Figure 1 shows the locations of the iron ore leases (MDM, 1941-1970) plotted on top of the bedrock geology of the mining district in western Fillmore (Mossler, 1995), eastern Mower (Mossler, 1998) and southern Olmsted (Olson, 1988) Counties in Minnesota. Figure 1 is an updating of Figure VI-43 in Bleifuss (1972, p. 499). Figure 1 is different from Bleifuss’ Figure VI-43 only in the bedrock geology, which has been significantly updated. All of the ore bodies were located on what is now interpreted as either the Devonian Spillville Formation or the Ordovician Stewartville Formation. The Spillville is a subdivision of Bleifuss’ (1972) Cedar Valley Formation. The Stewartville Formation is a subdivision of Bleifuss’ (1972) Galena Formation.

Based on the more recent geologic mapping shown in Figure 1, for the rest of this paper we will update the formation names from Bleifuss (1966, 1972) by substituting “Spillville” for “Cedar Valley” and “Stewartville” for “Galena”.

The Stewartville and Spillville Formations have the greatest secondary karst transmissivity of the geologic units shown on this map. All of the geologic units on Figure 1 regionally dip at a few feet per mile to the southwest. The iron ores are conspicuously not present on the Maquoketa and Dubuque Formations which are stratigraphically between the Stewartville and Spillville Formations.

Figure 2 is modified from Andrews’ (1958) Figure 2 with the names of the geologic units updated to current nomenclature. Although Andrews did not use the word “karst”, he recognized that “solution activity” was an important part of the process. Andrews (1958, p. 614-615) reasoned, based on the work of Krumbein and Garrels (1952), “that the iron was transported in an acidic solution (pH less than 7) in the ferrous state and that deposition resulted from an increase in pH of the solution. This increase in pH may be logically attributed to the reaction of the acidic solution with carbonate bedrock and resulted in precipitation of ferric oxide from this neutralized solution. It is thought that the ferric oxide could be precipitated in this manner both at the surface and by downward-percolating waters (emphasis added) in fissures of the underlying carbonate bedrock.”

Bleifuss (1972, p. 498) argued to the contrary that his observations and data indicated “the ores are Tertiary in age, and that they were developed from the oxidation of a primary marine siderite faces of the Cedar Valley Formation.”

The origin of the southeastern Minnesota iron deposits has long been cryptic and controversial and remains so. The fundamental issue, on which there is no consensus answer or model, can be summarized in simple questions. What was the source of the iron? How did that iron accumulate into mineable ore bodies in the Fillmore County district?

The thesis of this paper is that available field and textural evidence is consistent with a hypogenic origin of these iron deposits. Before the current Mississippi River drainage system developed, regional ground water flow systems could have emerged through the karst conduits in the Paleozoic carbonates. The waters in the deeply buried aquifers underlying this area currently are anoxic and enriched in dissolved ferrous iron and would have been more so before the entrenchment of the Mississippi River reorganized the regional ground water flow system. When that water emerged into the atmosphere the ferrous iron would have quickly been oxidized by a combination of biotic and abiotic processes producing the ferric oxide ores at the spring orifices.

Bleifuss (1972, p. 498) summarized the previous conceptual model as:

1. “The ores were formed by weathering of the underlying limestone units;
2. The development of the ore bodies required some supplementary process of concentration, involving migration and local concentration of iron during the weathering cycle;
3. The age of the Windrow Formation is Cretaceous, and the deposits in the Fillmore County district are correlative with similar lithologic units of known Cretaceous age in other parts of the region;
4. Fossil evidence that would positively date the Windrow Formation is absent in the district; and
5. The most likely age of the iron-rich residuum and associated iron ores is Cretaceous.”

Bleifuss (1972, p. 498) argued to the contrary that his observations and data indicated “the ores are Tertiary in age, and that they were developed from the oxidation of a primary marine siderite faces of the Cedar Valley Formation.”

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Figure 1. Iron mine parcels of the SE Minnesota iron mining district superimposed on the bedrock geology. Modified from Figure VI-43 in Bleifuss (1972).
Description of the Ore Bodies

Figure 3 is a part of Plate 5 from Bleifuss (1966) showing a plan view of a cluster of ore bodies in sinkholes on the Stewartville Formation. Figure 4 is Plate 6 from Bleifuss (1966) showing three cross sections through one of the ore bodies.

The following descriptions of ore bodies are repeated here because the original exposures no longer exist.

“The ore bodies overlie either the Spillville or the Stewartville Formations, and range in thickness from 3 to 30 feet. An under clay which ranges in thickness from a few tenths of an inch to more than two feet is developed between the ore and the underlying carbonate rocks. The ore is locally overlain by decomposed Spillville Formation, residual clays, or sediments of the Ostrander Member of the Windrow Formation. Both the Spillville Formation and the Stewartville Formation beneath the ore generally are fresh, although they may have been changed to a sandy dolomite ranging in thickness from a fraction of an inch to several feet.…

Although the ore bodies developed on the Spillville and Stewartville Formations are chemically and physically similar, they differ in size and shape. The ore bodies on the Spillville Formation generally have a greater areal extent, are more uniform in thickness, and have less relief than those on the Galena Formation.… Deposits containing more than 50,000 tons of ore were common.

In contrast, the ore bodies on the Stewartville Formation are isolated and generally contain much smaller tonnages. Generally, the upper surface of the ore is quite smooth, has a few closed depressions, and a relief rarely exceeding 10 feet. On a large scale, it is somewhat convex beneath the overlying unconsolidated materials.…

The relief on the carbonate bedrock surface beneath the ore on the Spillville Formation is small,… In contrast, the relief beneath the ore on the Stewartville Formation is much greater, and most of the mines show prominent bedrock ‘horses,’ some of which are more than 30 feet high.” (Bleifuss, 1972, p. 501.)
Figure 3. Plate 5 (cropped) from Bleifuss (1966). Plan view of iron ore bodies on the Stewartville Formation.

Figure 4. Plate 6 from Bleifuss (1966), cross sections of iron ore body shown in Figure 3.
**Description of the Iron Ores**

“The ore is composed predominantly of the mineral goethite and has minor amounts of hematite. The major gangue constituents are silt-size quartz and minor amounts of illitic clay. Two types of ore are readily identifiable in the field – ‘hard ore’ and ‘soft ore.’ The term ‘hard ore’ is applied to that material in which the principal ore mineral is dense, hard, crystalline goethite. Its most striking physical characteristic in place is its coarse, broken rubbly appearance. In typical exposures, it is composed of a mass of broken, closely-packed, angular fragments, one half to two inches across, that are intermixed with nodular masses of goethite as much as 10 inches in maximum dimension. A distinct horizontal layering is visible in some exposures, with individual beds being as much as six inches thick…..” (Bleifuss, 1972, p. 499-500).

“The soft ore, in contrast, appears rather massive and structureless in the field, and lacks the rubbly or nodular structure characteristic of the hard ore. In hand specimen, it has a soft punky texture and can be carved easily with a knife. The ore has a high porosity and a low bulk specific gravity. The principal ore mineral is goethite that shows a wide range of color from the bright yellow of ocherous goethite through shades of tan, brown and dark brown, to the brilliant crimson of ocherous hematite…. the dark brown ore varieties have much more manganese (about 2.0 percent) than the yellow varieties (about 0.5 percent).” (Bleifuss, 1972, p. 499-500).

Figure 5 and Figure 6 are black and white images of samples of the hard iron ore. Figure 5 is from Stauffer and Theil (1944, Fig. 6) and Figure 6 is from Andrews (1958, Plate 1A). Both images show the layers of iron ore deposited concentrically around fragments of the limestone bedrock. Both samples are consistent with what would be expected when the iron oxides had been deposited from fluids, which flowed around and reacted with the limestone bedrock.

**Summary of Relevant Literature Observations**

1. Early work on the Fillmore District iron ore deposits viewed the ores as straightforward weathering residues from the underlying country rocks.

2. Andrews (1958, p. 597) argues that the iron ores were “deposited as a result of reaction of [acidic] iron-charged waters with carbonate bedrock” but doesn’t suggest a source of the acidic, iron rich waters.

3. Sloan (1964, p.18) considered the iron ores and associated Ostrander Gravels of Fillmore County to be Cretaceous in age. He observed that the
iron ores “typically occur on a karst topography, primarily as fillings in enlarged joints and sinkholes or caves.”

4. Bleifuss (1972, p. 498) argues that the ores were developed from the oxidation of a “primary marine siderite faces of the Cedar Valley [Spillville] Formation” but doesn’t explain the textural evidence that the deposition involved flowing water.

5. The ore deposits are developed only on the Stewartville and Spillville Formations and not on the Maquoketa and Dubuque Formations.

6. The iron ores are in and associated with karst sinkholes and solutionally enlarged fractures and caves.

7. Mystery Cave, the largest cave in Minnesota, is developed in the Stewartville and Dubuque Formations, contains evidence of hypogenic speleogenesis (Klimchouk, 2007) and is overlain by one of the iron ore mines.

**A Hypogenic Source of the Iron Ores**

In other papers at this conference and in this paper, we are proposing that hypogenic regional groundwater flow systems have operated, and continue to operate, in southeastern Minnesota’s bedrock aquifer systems. The current surface and groundwater flow systems drain to the Mississippi River and its tributaries. Older regional groundwater and surface water drainage patterns, before the current Mississippi River drainage developed, were from east to west and potentially may have been much longer.

Deep wells in southeastern Minnesota often produce waters that are very anoxic, enriched in dissolved ferrous iron, with near neutral pHs. Some of the deep wells produce brackish to saline waters which are anoxic and iron rich.

The Stewartville and Spillville Formations in Minnesota have high secondary porosity and permeability and are regional aquifer systems. The Decorah Shale aquitard constrains the bottom of the aquifers. The Pinicon Ridge Formation aquitard constrains the top. The Maquoketa and Dubuque Formations act as aquitards to separate the two regional aquifer systems. The Stewartville and Spillville Formations are the natural discharge points, where they reach the surface, for regional groundwater flow systems.

To our knowledge none of the numerous springs issuing from the Spillville and Stewartville Formations in the iron ore district of western Fillmore County are currently depositing iron oxides. However, about 45 km west, in western Mower County near Austin, Minnesota, the Cedar River has eroded the thick glacial sediments of central Mower County. The first bedrock there is the Spillville Formation. There are springs in those areas which are currently depositing iron oxides (Green and others, 2002).

The sandstone karst of north-central Minnesota (Shade, 2002, Shade and others, 2015) has many springs and seeps that are currently depositing significant amounts of iron oxyhydroxides.

Figure 7 is a recent photograph of one such spring. This spring issues from an enlarged joint in the Hinckley Sandstone. When sampled on June 14, 2001, (Shade, 2002) the water was a low TDS, Ca (16.5 ppm), Fe (11.6 ppm), Mg (6.5 ppm), Na (2.3 ppm)/bicarbonate (alkalinity = 74 as ppm CaCO$_3$) water. The SO$_4$ (0.52...
ppm) and Cl (0.43 ppm) were very low. The pH of the water was 6.3. The leaf-covered mound in front of the lady is brown iron oxide that is several feet thick. This area was glacially scoured at the end of the Wisconsinan. The entire accumulation of material therefore must be less than about 10,000 years old.

When the anoxic, ferrous iron-enriched groundwaters discharge to the surface, both abiotic and biological processes rapidly oxidize the soluble ferrous iron to insoluble ferric iron and precipitate iron oxyhydroxides.

The oxidation of ferrous to ferric iron releases hydrogen ions which rapidly lower the pH and acidify the waters near the surface and at the surface.

The acidic waters aggressively react with and dissolve the carbonate bedrock at and near the surface. This enlarges the near surface fractures and creates the karst depressions that fill with the iron ore bodies. The karst depressions are enlarging as they fill with iron ore so the ores collapse on a local scale and produce the ore breccias seen in the mines.

If and when a particular hypogenic flow path becomes clogged with iron oxides, the flow will find other nearby paths to the surface and create new iron ore accumulations. Depending on the local geometry, these accumulations of iron oxides can build mounds and/or coalesce to form larger structures. The insoluble iron oxides will then tend to armor the carbonates they cover against dissolution by surface precipitation and localized topography reversals can occur.

The age of the Fillmore County iron deposits is very poorly constrained. The ores are pre-Pleistocene and are underlain by Devonian and Ordovician carbonates. Sloan (1964) concluded that the ores were Cretaceous. Bleifuss (1966, 1972) argued the ores are Cenozoic aged. In either case, there are 10s of millions, if not a 100 million years available for their formation. Nor is there any necessity that the iron ores all formed at the same. Iron depositing springs would likely migrate across the landscape, as surface erosion and regional groundwater flow systems evolve.

Estimating the flow of the Gushing Orange Spring shown in Figure 9 at 500 liters/minute and using Shades’ (2002) chemistry, we calculate that this spring discharges about 3 metric tons of iron per year. The entire Fillmore County iron ore district could easily have been produced in the available time by similar springs.

Acknowledgments
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References


DOWN THE RABBIT HOLE: IDENTIFYING PHYSICAL CAUSES OF SINKHOLE FORMATION IN THE UK

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Abstract
Heavy precipitation in the UK in February 2014 induced ground subsidence and consequently a rapid increase in the frequency of sinkhole occurrences. These new sinkhole collapses emphasize the need to further analyze the causes of the increased occurrence by investigating the relative importance of various surficial factors.

Malham and the Mendips are two areas of particular interest, since both are underlain by limestone bedrock and are susceptible to subsidence. This is due to limestone being primarily permeable in joints, and so it dissolves to form an extensive network of karstic caves. It was therefore useful to compare two sites of similar geology, both from the Triassic and Jurassic periods, as this controlled the amount of presently exposed limestone from past glacial retreat, for accurate comparison of susceptibility.

Susceptibility maps of the two areas were created by integrating GIS application and statistical methods to develop algorithms to address the issue of dissolution. The maps aim to identify the physical surficial conditions, in addition to heavy precipitation that exacerbates subsidence development.

Statistical testing of the GIS data indicated that in Malham, slope is the most significant parameter (Kruskal-Wallis, H=29.36, p<0.001; H=14.55, p=0.006, respectively) in sinkhole formation; while in the Mendips altitude is the most significant parameter (Kruskal-Wallis, H= 20.44, p<0.001; H= 86.51, p<0.001, respectively). Curvature appeared less statistically significant with fewer values reported from post-hoc Mann-Whitney U tests. This integrated geological mapping and statistical approach will prove useful in delineating susceptibility zones in areas within the UK.

Introduction
This study investigates the physical and surficial causes of sinkhole formation in both Malham and the Mendips, in the UK. These are two areas underlain predominantly by limestone bedrock, and are highly prone to dissolution, due to the highly soluble nature of the limestone (Waltham et al., 1997). This study explores how surficial features exacerbate dissolution, and aims to demonstrate that bedrock characteristics are not the most important factors. In addition, the aim is to assess the relative importance of the surficial features. Base map data and geological data were obtained from EDINA DigiMap and the British Geological Survey respectively (BGS License number 2014/143 ED British Geological Survey© NERC. All rights reserved. Edina DigiMap© Crown Copyright/database rights 2015. An Ordnance Survey/ (Datacentre) supplied service). This data, together with surface feature data from ArcGIS, enabled the creation of zonation maps with high, considerate, moderate and low susceptibility areas.

Within this report, predictions are estimated on a spatial scale, following the sensitivity associated with temporal prediction. Therefore, chronological data is not accounted for, and the maps only present visual future susceptibility on a two-dimensional level.

This report analyzes the role that surficial factors of slope, curvature, and altitude within the two areas play in exacerbating subsidence, and whether one factor in particular may be more critical. This analysis assesses the relative importance of each variable, to ultimately create a reliable spatial sinkhole susceptibility map.

Due to the topical, public, and media interest in sinkhole collapse, this research is of significant importance in today’s environment, economy, and society. Sinkholes affect 15% of the world’s surface today (Wilson and Beck, 1992). Most relevant papers date back to the early 1900s (Elrod, 1898; Vineyard and Williams, 1967; Purdue, 1907). These predominantly focused on pre-existing cavities, where the limestone cave systems that were once mining sites, initiated subsidence. Recent literature now focuses more upon the range of external factors that exert pressure on these vulnerable locations, due to the increasing availability of modern technological equipment, allowing more in-depth analysis (Sass, 2007; Cooper, 2008; Stecchi et al., 2009).
This paper aims to narrow down the importance of particular surficial factors. Based on a range of literature (Waltham, 2008; Parise et al., 2009; Parise, 2010) it is clear that this hazardous phenomenon has the ability to destroy lives and local communities. The creation of any susceptibility zonation maps based on the current ambiguity of such existing surficial causes will provide insight into avoiding a potentially unsafe environment. Though many variables involving subsidence formation have been previously investigated in research, no definitive answers have been concluded following the arbitrary nature of sinkholes (Upchurch and Littlefield, 1988; Florea et al., 2002) and the fairly novel area of sinkhole research. This paper determines the importance of each factor, rather than concluding the generic causation of multiple factors. Furthermore, many studies focus on evaporite karst areas (Johnson, 1997; Cooper, 2008; Gutiérrez et al., 2008; Galve et al., 2009) due to its higher susceptibility to dissolution, though carbonate karst is more common (Gutiérrez et al., 2008).

This study focuses on limestone carbonate karst areas from the Triassic and Jurassic geological time periods; these two areas have similar geology, with only surficial differences for susceptibility mapping. Though it can be hard to justify the specific causes following such apparent spatial dichotomies even within the UK, it was necessary to focus on a local scale, in order to identify detailed causes, rather than wider, regional causes.

This study identifies gaps in literature by creating a susceptibility map comparing two areas on a local scale, using purely surficial and physical factors in order to obtain as much detail and understanding as possible. Though zonation maps have been previously created, they have been predominantly single-site based and scale-specific (Kaufmann and Quinif, 2002; Stecchi et al., 2009). This paper aims to further this research by creating a comparative map of two areas based upon multiple surficial factors.

This study is meant to be useful in mapping the safety zonation of areas for future building (Gutiérrez et al., 2008), and therefore aims to implement a preventative measure, and create local awareness of specific conditions that may aggravate subsidence (Farrant and Cooper, 2008).

**Study Site**

Two 25km² areas consisting of pre-existing doline points were extracted from the BGS GIS database for each study site; East of Settle around the Craven District in Malham, and North-west Mendip Hills (Figures 5 and 6). Larger sized areas are also used however, to assess future susceptibility based on the slope, aspect and curvature of the surrounding areas, determined by the values as grouped in Table 1. These sites were selected based on the predominant presence of limestone bedrock defining these two areas, in addition to known sinkhole activity.

**Methods of Study**

This research relied purely on secondary data obtained from the British Geological Survey, in order to analyze the physical formation, and spatial distribution of sinkholes. A 1:50,000-resolution, 50m grid cell size digital elevation model and BGS doline data for each area was imported into ArcGIS. The doline points provided extensive information based on count, type, shape and distribution of the points. Distance between the points was calculated using ‘point cluster analysis’, and each point was then corresponded to its bedrock class that it was underlain by. The ‘identity’ tool further enabled the partnering of each point with its related topographical slope, curvature and altitude values, and enabled the integration of statistical testing and GIS. Curvature can be defined as the degree to which a surface is curved, and can be strongly linked with trends of faults and topographical fractures (Stecchi et al., 2007). The curvature is the second derivative of the elevation surface, which was run on a 3x3 cell scale determined by the DEM grid size resolution. The layer was filtered

| Table 1. Susceptibility key for mapping areas. |

<table>
<thead>
<tr>
<th>Level</th>
<th>Description</th>
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</thead>
<tbody>
<tr>
<td>High</td>
<td>Flat slope &lt;3°</td>
</tr>
<tr>
<td></td>
<td>Linear curvature ≈ 0</td>
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<tr>
<td></td>
<td>On limestone</td>
</tr>
<tr>
<td></td>
<td>High altitude</td>
</tr>
<tr>
<td>Considerate</td>
<td>Gently sloping &lt;6°</td>
</tr>
<tr>
<td></td>
<td>Near linear curvature ≈-1 to 1</td>
</tr>
<tr>
<td></td>
<td>Mudstone</td>
</tr>
<tr>
<td></td>
<td>Intermediate altitude</td>
</tr>
<tr>
<td>Moderate</td>
<td>Slopes &lt;9°</td>
</tr>
<tr>
<td></td>
<td>Linear curvature ≈-2.5 to 2.5</td>
</tr>
<tr>
<td></td>
<td>Siltstone and interbedded rocks</td>
</tr>
<tr>
<td></td>
<td>Intermediate altitude</td>
</tr>
<tr>
<td>Low</td>
<td>Steep slope &gt;9°</td>
</tr>
<tr>
<td></td>
<td>Extremely convex or concave curvature ≈-5 to 5</td>
</tr>
<tr>
<td></td>
<td>Non-porous bedrock e.g. sandstone</td>
</tr>
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<td></td>
<td>Low altitude</td>
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to remove any minor topographical hollows or peaks (Sullivan et al., 2007; Stecchi et al., 2009). Each surficial factor was reclassified into low-high susceptibility groups based on natural breaks, which was used to highlight the values where points naturally clustered. The susceptibility map was consequently created through using the ‘raster calculator’ tool, multiplying each surficial layer together. The output was reclassified into four levels of susceptibility.

**Results of Study**

Normality tests were executed on each variable; every variable returned as not normal, so non-parametric tests were performed throughout. All Mann-Whitney U tests were carried out with 95% confidence. It is clear from Figures 1 and 2 that there is a higher frequency of sinkholes on flatter slopes than steep slopes, and on linear curvatures than concave or convex. There is less of a correlation present based on altitudinal values. In Malham, the highest frequency of 64 sinkholes occurs at 510m above sea level, though this is not the highest elevation. The second highest peak consists of 52 sinkholes at 380m above sea level. In the Mendips, the frequency also varies, with a peak count of 39 sinkholes at 290m above sea level.

Malham bedrock clustering did however return with a statistical significance (Kruskal-Wallis, H=23.82, p<0.001) (Figure 3). Bedrock clustering in the Mendips returned with no statistical difference across different bedrock classes (Kruskal-Wallis, H=8.39, p=0.078) (Figure 4). Slope, altitude and curvature each presented a difference with Kruskal-Wallis testing and a further post-hoc Mann-Whitney U for individual bedrock class pairing. The results were more varied for Malham, with less of a pattern presented; limestone consistently

![Figure 1. Frequency bar charts in Malham (left) presenting the number of sinkholes (total: 400) on each driver of slope, curvature and altitude respectively.](image)

![Figure 2. Frequency bar charts in the Mendips (right) presenting the number of sinkholes (total: 161) on each driver of slope, curvature and altitude respectively.](image)
presented a difference. In the Mendips, conglomerate appeared to show the strongest difference between each other class within all tests for the surficial factors. Chert also showed significance when tested for altitudinal difference (Figure 4).

Sandstone presents the widest range of curvature values, ranging from -0.8 to 0.5 1/100 z-units. In contrast, limestone presents the smallest mean range, with 121 sinkholes at 0 1/100 z-units, though with multiple outliers on extreme curvature values, accounting for the large number of sinkholes apparent on limestone bedrock. More outliers are evident in Malham (Figure 3) than the Mendips (Figure 4) due to the wider range of sinkholes that exist outside the predicted common thresholds.

The map above (Figure 5) was created using the raster calculator by combining slope, curvature and altitude. The high-susceptibility values were defined as <5.7° for slope, >413m above sea level for altitude and -0.34 to 0.12 1/100 z-units for curvature; this was based on reclassified natural data breaks with sinkhole frequency. Each reclassified layer was input into the raster calculator where they were combined to provide an output layer. This was reclassified again into the four classes.

Figure 3. Malham doline point clustering on each bedrock class presented by boxplots on each surficial factor: slope, curvature and altitude respectively.

Figure 4. The Mendips doline point clustering on each bedrock class presented by boxplots on each surficial factor: slope, curvature and altitude respectively.
The high-susceptibility values for slope were defined as <3.7° for slope, -0.37 to 0.16 1/100 z-units for curvature and >176m above sea level for altitude. The same method as stated above was used here also.

**Discussion**

**Susceptibility Maps**

The models above (Figures 5 and 6) present high to low susceptibility of sinkhole formation in Malham and the Mendips. It is encouraging that the high frequency of pre-
existing sinkholes is mapped on the highly susceptible red areas, whilst the orange areas have few sinkhole densities, and green and yellow areas have few to none. Based on pre-existing locations of current sinkholes and local topographical features of slope, altitude and curvature, the different areas demonstrate potential wider locations of future sinkhole development. Thus, demonstrating that such local scale features clearly exacerbate sinkhole existence. However it is also clear that some sinkhole points occur on the considerate susceptibility areas. This is because the threshold of the slope, curvature and altitude values do not define the exact points that sinkholes can only occur on, but instead provide an indication of the most vulnerable
areas, though considerate to low areas still need to be considered regardless. The threshold values also largely vary on a spatial scale, even within two similar geological areas within the UK, and so these varying boundaries need to be site-specific. Results statistically proved local factors of flat slopes, higher altitudes and linear curvatures to be more susceptible to sinkhole formation, with the exacerbation of precipitation. Though it was expected that concave curvatures would be highly susceptible, my statistical results found linear curvatures to be more so. However, difficulty does arise with curvature analysis through the necessary filtering. These results were also consistent with findings from Farrant and Cooper (2008), Simms and Ruffell, (1989) and Sánchez et al. (2007). The basic theory that these local factors exacerbate subsidence can however provide an approximate and valuable insight into potentially vulnerable locations, with the suitable underground conditions.

Furthermore, in standardising approximations, the high-susceptibility percentage of the total slope and altitudinal values can be quantified. The areas combined present susceptible slopes to occur at 9-11% of the total slope, whilst highly vulnerable elevations occur at 46-64% of the total range.

Slope
A difference in sinkhole frequency was clearly evident, with flat slopes containing more sinkholes than steep slopes. This finding was expected. Farrant and Cooper (2008) suggested that dissolution pipes and irregular rockhead form on flat slopes, ideal for karst formation, whereas steep slopes promote erosion. This finding is additionally supported in the context of landslides in Glade (2005) and Cooper (2008), where steep slopes induce ground instability and consequential rockfall, thus also causing erosion indirectly. It is also interesting to note that Farrant and Cooper (2008) claim slope to be an irrelevant factor when considering evaporite subsidence on gypsum and salt bedrock, as the karstic rock is rarely exposed to the surface.

In contrast, Santo et al. (2007) highlights the critical relationship between carbonate karst and local slope stability. This is an issue where widespread presence of carbonate karst and a high availability of dissolution to the slopes through hydrogeology induce subsidence. This is therefore a relevant study, as sinkhole formation can be directly correlated with slope where instability occurs in carbonate karst. Although it is evident gentler slopes are more susceptible than steeper slopes, perhaps the characteristics relating to carbonate karst slopes exacerbate this, in addition to the physical slope angle.

Though my results demonstrated gentle slopes to be more prone to sinkholes, Stecchi et al. (2009) reported building destruction initiated by ground subsidence on these “gentle” slopes. It is interesting to define the thresholds of “gentle” and “steep” slopes; from this study, it is clear that slope boundaries are highly subjective, even on the small-scale, local analysis that this study is built on. For example, sinkhole frequency is high on slopes ranging from 0-6° in Malham, but only 0-2° in the Mendips, thus the “gentle” boundary could not be equally applied. Furthermore, Glade (2005) presents slopes >2° as the most active due to high erosion and weathering processes, though this can only be site-specific to his study.

A critical and potentially useful theory investigated by Sass (2007) looks into the surface depth to bedrock measurements on slopes. This is interesting as although bedrock depth is not considered in this report, it could explain the presence of certain bedrock types in particular locations; if a steep slope is heavily eroded, it would reveal deeper underlying bedrock layers, than a flat slope comprised of the original first layer bedrock, that is a target for further deposited material. This could therefore explain in the Mendips for example, why bedrock such as sandstone, a predominantly low permeability rock (Ward and Morrow, 1987), is more prominent on steep slopes; mudstone and siltstone, a more resistance clayey material, (Franklin and Chandra, 1972) is more prominent on flat slopes, if sandstone was originally at deeper depths than other bedrock.

Stecchi et al. (2009) also present slope to only be connected with ground movements, and not topography, however multiple papers (Doctor and Young, 2013; Rahimi and Alexander, 2013) note that the visual surface depressions and hollows indicate subsidence and sinkhole development.

Curvature
It was anticipated that sinkhole frequency would be higher on concave curvatures than linear or convex curvatures, due to the heavy pooling of precipitation.
It is interesting to note the disparities among theories involving the promotion and prohibition of sinkhole development. UWSP (n.d.) explains how higher altitudes promote stronger weathering processes, and therefore expose the karstic bedrock further. In contrast to this, Tharp (2002) similarly presents findings of low curvature being most prone to sinkhole formation due to hydraulic fracturing. In contrast to this, Stecchi et al. (2009) found high curvature values to correlate with fractures and fault lines. This therefore highlights the difficulty in obtaining accurate curvature analysis, based on the high level of filtering generally needed to take into account the wider landscape, rather than minor topographical changes. Though Stecchi et al. (2009) claimed that no filtering was needed, due to the smoothness of the raw data, Sullivan et al. (2007) and Bergbauer and Pollard (2003) state the necessity of filtering data, in order to avoid any problems relating to the dependence on the sample grid, and consequent focus on minor topographical disparities, as opposed to wider changes.

A further notable link with curvature, are fault lines and fracturing (Murray, 1968; Vendeville, 1991; Tharp, 2002). This is due to the ability of curvature to predict the distribution of deformation (Bergbauer and Pollard, 2003), and its close relationship to geology. Vendeville (1991) points out the criticality of curvature analysis, in how it varies significantly with geology; this reason underlies the study’s choice to compare two sites of similar geology from the Jurassic/Triassic geological time periods.

Malham’s range of concave curvature values, in contrast to the Mendips could therefore present a general sinking in the surface; Stecchi et al. (2009) report the negative values to be indicative of sinking.

**Height**

Although some high altitudinal values correlated with high sinkhole frequency, there was a wide range of variance in the data. Whilst a pattern is less notable in the Malham (Figure 1), this is most likely due to the wider range of data in the Mendips, and overall higher number of sinkholes present in Malham (Figure 1), so the spread is wider. In the Mendips, the distinct drop in frequency at its highest altitude is questionable; this is still lower than Malham’s lowest altitude, which highlights issues of changeability across spatial scales.

It was expected that limestone would comprise the flattest slopes, highest altitudes and most linear curvatures. However, limestone did not show as much of a significant difference as expected at either site (Figures 3 and 4), in contrast to that of conglomerate in the Mendips (Figure 4), which appeared to be the most statistically different
to all other bedrocks. However conglomerate consists of limestone fragments (BGSd, n.d.), and so can be said to hold similar characteristics to pure limestone, so the theory is not wholly disproved.

In Malham, slope appears to present a stronger statistical difference across each bedrock class than any other factor; altitude also portrays strong importance. Alternatively, curvature values across both sites appear to have the least importance in determining sinkhole formation, with the least values presenting a statistical significance. This could be explained by the ease of error created, and difficulty in analysing curvature, as noted in Bergbauer and Pollard (2003) and Sullivan et al. (2007).

In the Mendips, the most important driver of subsidence appears to be altitude, presenting the strongest statistical difference across bedrock. Though there is much conflict over whether high altitude influences sinkhole development or not (Simms and Ruffell, 1989; Santo et al., 2007; UWSP, n.d.), perhaps it is more directly the difference in altitude across the different bedrock classes that define development.

Overall, it can be said that slope and altitude have a mutual importance as drivers influencing subsidence, though each portrays different significance within each site. This is due to the change in driver importance across varying spatial gradients, highlighting the need to adapt to the change in site-specific drivers. Though even within the UK Malham and the Mendips present a difference in driver importance, confident conclusions can be drawn, based on the similarities.

**Conclusions**

Three conclusions can be drawn from this investigation:

1. Slope and curvature appear to be the most significant drivers influencing sinkhole formation; given the difficulty in curvature analysis however, slope is the most reliable and critical factor.

2. Spatial zonation maps are still valid even without the detail of temporal data. However, this research has presented the dichotomies present even across a small spatial gradient within the UK, which still creates some ambiguity in the surficial drivers.

3. Susceptibility mapping has proved useful for the future, though studies must be aware of the spatial inconsistencies present. This therefore calls for a wider-scale hazard susceptibility map encompassing larger areas and temporal data.

**References**


RELAY RAMP STRUCTURES AND THEIR INFLUENCE ON GROUNDWATER FLOW IN THE EDWARDS AND TRINITY AQUIFERS, HAYS AND TRAVIS COUNTIES, CENTRAL TEXAS

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Abstract
The Cretaceous Edwards and Middle Trinity Aquifers of central Texas are critical groundwater resources for human and ecological needs. These two major karst aquifers are stratigraphically stacked (Edwards over Trinity) and structurally juxtaposed (normal faulting) in the Balcones Fault Zone (BFZ). Studies have long recognized the importance of faulting on the development of the karstic Edwards Aquifer. However, the influence of these structures on groundwater flow is unclear as groundwater flow appears to cross some faults, but not others. This study combines structural and hydrological data to help characterize the potential influence of faults and relay ramps on groundwater flow within the karstic Edwards and Middle Trinity Aquifers. Detailed structure contour maps of the top of Walnut Formation in the study area were created from a geologic database (n=380) comprised of primarily geophysical and driller’s logs. The data were then contoured in Surfer® (Kriging) with no faults. Structure contour surfaces revealed detailed structural geometries including linear zones of steep gradients (interpreted as faults) with northeast dipping zones of low gradients (interpreted to be ramps) between faults. Hydrologic data (heads, dye trace, geochemistry) were overlaid onto the structure contour maps in GIS. Results for the Middle Trinity Aquifer suggest relay ramps provide a mechanism for lateral continuity of geologic units and therefore groundwater flow from the Hill Country (recharge area) eastward into the BFZ. Faults with significant displacement (>100 m) can provide a barrier to groundwater flow by juxtaposition of contrasting permeabilities, yet flow continues across fault zones where ramps exist, or where permeable units are juxtaposed with other permeable units. In the Barton Springs segment of the Edwards Aquifer the primary flow path defined by dye tracing and heads is coincident with the Onion Creek relay ramp dipping to the northeast. This work addresses the lateral continuity (intra-aquifer flow) of the Edwards and Trinity Aquifer systems, which has importance for conceptual models and ultimately resource management.

Introduction
The Cretaceous Edwards and Middle Trinity Aquifers of central Texas are critical groundwater resources for human and ecological needs (Figure 1). These two major karst aquifers are stratigraphically stacked and structurally juxtaposed in the Balcones Fault Zone (BFZ) (Figure 2). However, the role of faulting and related structures on groundwater flow is not clearly understood due to the stratigraphic and structural complexity.

The purpose of this paper is to describe the influence of faults and related structures called relay ramps on
Figure 1. Simplified geologic map with potentiometric surfaces of the Middle Trinity and Edwards Aquifers. Two major faults, the Mount Bonnell and the San Marcos faults, are shown with the relay ramp structure proposed by Grimshaw and Woodruff (1986) and Collins and Hovorka (1997). Figure modified from Smith et al. (2015).
groundwater flow within the Edwards and Trinity Aquifers in a portion of the BFZ in central Texas.

Studies have long recognized the importance of faulting for the development of the Edwards Aquifer (Hill and Vaughan, 1898; DeCook, 1963; Sharp, 1990). More recently, studies have addressed the hydrologic connection within the Edwards and Trinity Aquifers (Smith and Hunt, 2010; Gary et al., 2011; Wong et al., 2014; Smith et al., 2015) with some studies focusing on structure (Ferrill et al., 2008). However, the influence of faults and the related “relay ramp” structures on groundwater flow have not been fully characterized. Cross sections through the BFZ generally show vertical offset and suggest lateral discontinuity, which may or may not occur in three dimensions (Figure 2). Yet recent studies of groundwater flow suggest lateral continuity of flow across or around faults in the BFZ (Figures 1 and 2; Hauwert et al., 2004; Smith et al., 2015). This paper will explore the mechanism for lateral continuity of flow in a karst setting with complex structures. Implications of this work address the lateral continuity of units and therefore intra-aquifer flow. This has great importance for conceptual models and ultimately, resource management.

**Structural Setting**

A series of complex tectonic cycles have strongly influenced the hydrogeology of central Texas. The tectonic events or cycles are described in detail in Ewing (1991), and are composed of the Grenville (pre-Cambrian), Ouachita (late Paleozoic), and Gulfian (Triassic to present) cycles. The Llano Uplift is a structural dome in central Texas which is related to the formation of the San Marcos Arch. These features influenced Cretaceous deposition and subsequent structures, such as the BFZ (Figure 1).

![Figure 2](image-url). Geologic cross section along the Blanco River showing the geologic and hydrogeologic units. The faults shown are normal faults of the BFZ. Note the Edwards and Trinity Aquifer are both stratigraphically stacked and structurally juxtaposed. Groundwater flow is schematically shown to move across faults. Line of section A to A’ is down the Blanco River in Figure 1. Figure modified from Smith et al. (2015). Vertical Exaggeration is ~100x.
**Balcones Fault Zone: A Review**

The BFZ produces the prominent physiographic feature known as the Balcones Escarpment in central Texas. The BFZ is a dominant structural feature extending in an arcuate pattern from Del Rio along the border with Mexico, toward Dallas in north Texas. The BFZ trend changes from W to NNE (Figure 1). The BFZ is a fault system consisting of numerous normal faults with hanging walls generally dropping down toward the Gulf of Mexico with displacements ranging from 30 to 260 meters. There are up to 365 meters of total displacement across the BFZ. Faults are generally steeply dipping (45-85 degrees) with stratigraphy a fundamental control on the geometries and dips (Ferrill and Morris, 2007). Faults generally trend to the NE (N40 to 70 E) and dip to the southeast (Collins and Hovorka, 1997). The faults are described as “en echelon,” which indicates closely-spaced, overlapping and subparallel. Depending on location, the faults can occur at oblique angles to the overall regional structural trend. The BFZ is characterized by numerous structures including horsts, grabens, and relay ramps (the focus of this paper). The BFZ generally follows the strike of the Cretaceous units and the trend of the Paleozoic-age Ouachita front (Sellards and Baker, 1934; Grimshaw and Woodruff, 1986; Ewing, 1991; Barker and Ardis, 1996; Collins and Hovorka, 1997; Collins 2004). The faults extend down into the Ouachita rocks and may also pass into extensionally reactivated Ouachita faults (Ewing, 1991); but they may also have listric geometries that terminate or sole out into shales at depth (Collins and Hovorka, 1997).

The BFZ is Tertiary in age, but the exact period or epoch of faulting is uncertain—the youngest sediments to be faulted are late Paleogene (Eocene-age ~55 Ma; Sellards and Baker, 1934). However, most of the fault movement is thought to have occurred during the early Neogene (late Oligocene ~30 Ma or early Miocene ~15 Ma). This timing is also coincident with regional uplift centered on the Colorado Plateau and extensional Basin and Range province which extends into west Texas. Although the BFZ is located at the boundary between the uplifting plateau area and the subsiding Gulf Coast Basin, it is unknown if the uplift and extension of the Basin and Range is related to the BFZ (Ewing, 1991; Collins, 2004). Instead, the BFZ may have formed as a result of the sedimentary loading and extension of the Gulf Coastal Plain toward the Gulf of Mexico Basin (Collins, 2004). Ewing (2004) describes the formation of the BFZ as the differential subsidence and slippage along the old Ouachita lines of weakness.

**Relay Ramps**

Normal faults are inclined dip-slip faults in which the hanging wall moves down compared to the footwall. They generally have steep dips of 60 degrees or greater, depending upon the stratigraphic unit (Ferrill and Morris, 2007). Where the offset along a fault decreases along its strike to zero, the extension is taken up by adjacent sub-parallel (en echelon) faults. Between these faults (that dip in the same direction) there is often a “transfer zone” where deformation is accommodated by folding, faulting, and fracturing (Twiss and Moores, 1992). These are the structures described as “relay ramps” (Figure 3; Grimshaw and Woodruff, 1986; Collins and Hovorka, 1997).

Relay ramps of different scales are described as occurring in the BFZ (Collins, 1995; Collins, 2004). Grimshaw and Woodruff (1986) describe two en echelon faults and an associated relay ramp structure in the San Marcos area that they hypothesize influenced the geomorphology and groundwater flow—namely the location of the Blanco River and San Marcos Springs. This same structure (Figure 1) is also mapped by Collins and Hovorka (1997).

Figure 3. Schematic diagram of a relay ramp structure and its influence on groundwater flow. Two major faults transfer the displacement from one to the other resulting in folding, fracturing and faulting (not shown) along the ramp structure. These structures were proposed and mapped by Grimshaw and Woodruff (1986) and Collins and Hovorka (1997). Figure modified from Grimshaw and Woodruff (1986).
**Hydrogeology and the Balcones Fault Zone**

The Trinity Aquifer is a sole-source supply for much of the central Texas Hill Country—its springs (Jacob’s Well and Pleasant Valley Springs, among others) provide baseflows that ultimately recharge the Edwards Aquifer down gradient (Figure 2; Smith et al., 2015). The Edwards Aquifer is also a significant sole-source supply for hundreds of thousands of people in central Texas and its renowned springs such as Comal, San Marcos, and Barton Springs provide habitat for a variety of endangered species.

The BFZ was critical to the hydrogeologic evolution of the Edwards and Middle Trinity Aquifers. Faulting provided the hydrogeologic architecture (e.g. recharge areas vs. confined aquifers) and the initiation point for karst processes (DeCook, 1963; Slade et al., 1986; Sharp, 1990; Ferrill et al., 2004). Structures such as joints and fractures influence the location and development of karst recharge features. These features often are located within stream channels and are capable of high rates of groundwater recharge (up to about 3,000 liters per second, or 100 cubic feet per second). Antioch Cave in Onion Creek which recharges the Edwards Aquifer and Saunder’s Swallet in the Blanco River which recharges the Middle Trinity Aquifer are examples of such features (Figure 1). Both aquifer systems contain joint-controlled conduits that transmit large amounts of water. The conduits are documented by cave maps, dye tracing, aquifer tests, and potentiometric surfaces (Wierman et al., 2010 and references therein). The structural influence on flow is more pronounced in the Edwards Aquifer at the regional scale as the aquifer is located entirely within the BFZ, while only the eastern portion of the Trinity is strongly influenced by the BFZ. However, major springs in both the Edwards and Trinity Aquifer systems are strongly influenced by structure as evidenced by faults at Barton Springs in the Edwards Aquifer and visible fractures or faults at Pleasant Valley Spring and Jacob’s Well Spring in the Middle Trinity Aquifer.

**Approach**

The approach to evaluate the influence of relay ramp structures on groundwater flow was to construct a detailed geologic contour surface in the BFZ (Figures 4 and 5). Different types of hydrologic data from the Edwards and also the Middle Trinity Aquifers were overlain onto this structure contour map.

The Walnut Formation (also known as the Basal Nodular Member) below the Edwards Group was the stratigraphic layer selected as the primary contour mapping horizon. The formation’s subsurface characteristics based on geophysical logs of wells are described in Hunt et al. (2011). It was selected as the primary mapping horizon for the following reasons: 1) The relative ease in identifying it in outcrop and geophysical logs (fossil assemblages and lithology in outcrop, high gamma ray signature on geophysical logs), 2) it represents the base of the Edwards Aquifer and, as such, many wells penetrate it, and 3) it has relatively consistent thickness through the study area. Geologic data used to construct the detailed structure contour surfaces were derived from an unpublished database maintained at the Barton Springs/Edwards Aquifer Conservation District. The database consists of well information and the tops of geologic formations or units primarily based upon geophysical logs plus driller’s logs, outcrops, and cuttings. A few contacts were derived from published geologic maps to fill in data gaps. Data for the top of the Walnut Formation were gridded and contoured in Surfer® using a Kriging algorithm—no faults were used in the gridding process. A total of 379 data points were used consisting of 45% geophysical logs, 42% driller’s logs, and 13% outcrops.

The gridded and contoured structural data was intentionally done without reference to faulting or structural domains. The authors believe the data reflect the overall geometry of the unit, without introducing the bias of mapped faults that in fact represent a spectrum of geometries from wide zones of dipping beds, to discrete offsets with variable throw. This approach is an obvious simplification of the structural surface, but allows for the significant geometries (major faults, and ramps) to be highlighted.

Simplified faults were drawn over the contour map from the Geologic Atlas of Texas (Stoeser, 2005) where gradients were steep and supported the presence of significant relatively discrete faults. These faults generally fall into two classes, those that have greater than 150 m displacement, such as the Mount Bonnell and the San Marcos Faults, and those with intermediate displacements up to 60 m. It is assumed that the structures mapped in the Walnut Formation persist at depth into the Middle Trinity Aquifer, about 150 m below the Walnut Formation.

Hydrogeologic data used in this evaluation are potentiometric data from Hunt and Gary (2014) consisting of a synoptic event during drought conditions (February-March 2009) in both the Edwards and Middle Trinity Aquifers (Figures 1, 5 and 6); dye-tracing data during low-flow conditions were summarized from the work of Hauwert et al., 2004 and Johnson et al., 2012 (Figure 5); geochemical data compiled from the Texas Water Development Board database and modified data from Wierman et al., 2010 (Figure 6).
Figure 4. Structure contour map of the top of the Walnut Formation—the base of the Edwards Aquifer. Two relay ramp structures are drawn where gradients flatten out between large faults. The two ramps are named Onion Creek Ramp (OCR) and the Kyle Ramp (KR). Contouring was done without faults in Surfer®. To illustrate the geometry of the relay ramp, faults were drawn where contouring supported their presence—these generally coincide with faults mapped in the Geologic Atlas of Texas (Stoeser, 2005).
Aquifer permeability is reported to generally be enhanced parallel to faults and decreases perpendicular to faults in the Edwards Aquifer (Ferrill et al., 2004; Ferrill et al., 2008). However, is it possible that other structures such as relay ramps may also influence groundwater flow? These types of structures have been attributed to control groundwater flow paths in other parts of the Edwards Aquifer, such as the Knippa Gap west of San Antonio (Clark et al., 2013). We know that the fractures associated with faulting can be as significant as the faults themselves in terms of influencing groundwater recharge, flow, and discharge. We hypothesize that the structural dip, and associated faulting and fracturing in a relay ramp could be a significant factor in influencing groundwater flow.

Groundwater flow from the Blanco Watershed eastward into the BFZ must flow “across” some faults with significant displacements. In fact, relay ramps provide a mechanism or pathway for groundwater to flow around faults, as the head and geochemical data in Figure 6 suggest. Where displacements are minimal or where permeable units are juxtaposed against each other, flow can actually be across the faults. In addition, head data also suggest that faults with significant displacements are indeed barriers to flow as shown by the NE-trending flow along the Mount Bonnell Fault (Figures 1 and 6).

Groundwater flow paths in the Barton Springs segment of the Edwards Aquifer have been conceptualized to be focused along solutioned (karstic) NE-trending fractures and faults (Hunt et al., 2005). The primary flow path in the Barton Springs segment of the Edwards Aquifer, determined from potentiometric maps and dye tracing, is called the Manchaca Flow route (labeled “MF” on Figure 5; Hauwert et al., 2004). This flow route generally coincides with faults (in part) and the Onion Creek ramp structure of this study. The wide potentiometric trough and circuitous dye trace paths within the Onion Creek ramp suggest a wide area of elevated permeability, instead of a single discrete flow path along a single fault zone. Dye tracer tests have demonstrated that during high-flow conditions in Onion Creek, groundwater flow can reverse directions and flow to the SE (up structural dip) toward San Marcos Springs (Smith et al., 2012).

A zone of highly permeable Edwards Aquifer is inferred from an area of very low hydraulic gradient extending south of San Marcos Springs along the saline zone, to the north toward Kyle (Land et al., 2010). This area is coincident with the Kyle ramp structure, an area bound by a significant fault on the northwest and the saline zone boundary to the east (Figure 5). Where the Kyle ramp ends, the high permeability zone appears to also

Results

Figure 4 presents the results of the structure contour map of the top of the Walnut Formation. The map contains generalized faults where the contour gradients suggest their presence and also interpreted ramp structures.

Hydrologic data in the Edwards (Figure 5) and the Middle Trinity (Figure 6) are overlain on the major faults and the interpreted ramp structures for comparison.

Discussion

Structure is an important control on the location of recharge, flow paths, and spring discharge locations in carbonate aquifers (Sasowsky, 1999), and for the Edwards and Trinity Aquifers (Ferrill et al., 2004). Faults have generally been the primary structure cited in the literature to influence groundwater flow (Maclay and Small, 1986; Hunt et al., 2005).
Figure 6. Map showing hydrologic data (potentiometric and geochemistry) relative to the ramp structures and major faults. Flow paths of the Middle Trinity defined by potentiometric heads and the “tongue” of low TDS appear to flow to the east along the Onion Creek ramp. Note the flow to the northeast indicating the fault may be a barrier to flow in that area. Potentiometric data from Hunt and Gary, 2014. TDS data modified from Wierman et al., 2010.
Groundwater flow appears to be strongly influenced by faults and by the fault-bound ramp structures. This essentially agrees with previous hypotheses of Grimshaw and Woodruff (1986) and Collins and Hovorka (1997). Detailed structural data and hydrologic data presented in this study support those hypotheses. However, instead of one large single ramp structure (Figure 1), the data supports a more complex system of at least two smaller structures (Figure 4). The practical application of mapping relay ramp structures is that it provides a mechanism for predicting lateral continuity of geologic units, and therefore lateral groundwater flow paths. A better understanding of these flow paths will provide a better understanding of the connection between recharge areas and the deeper portions of these aquifers that are being used more extensively for water supply.

Conclusions
Groundwater flow appears to be strongly influenced by faults and by the fault-bound ramp structures first proposed by Grimshaw and Woodruff (1986). Conclusions from this study include:

- Relay ramps provide a continuity of geologic units for lateral groundwater flow within the Middle Trinity Aquifer from the Hill Country eastward into the BFZ. At least two relay ramps (Onion Creek Ramp and Kyle Ramp) were identified in the study area based on detailed structure contours of the Walnut Formation.
- Relay ramps may influence groundwater flow in the Barton Springs segment of the Edwards Aquifer due to their northeast structural dip superimposed with solution along fractures and faults.
- Faults with significant displacements (Mount Bonnell) can provide a barrier to flow and force flow around faults and along relay ramps. Flow across moderate to minor faults can occur due to the juxtaposition of permeable units. These conclusions should be incorporated into conceptual models of the Edwards and Trinity Aquifers—and considered in the management of aquifers that span structural and political regions.

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References


GOLIATH’S CAVE, MINNESOTA: EPIGENIC MODIFICATION AND EXTENSION OF PREEXISTING HYPOGENIC CONDUITS

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Deceased

Abstract
Goliath’s Cave is developed in the Ordovician Dubuque and Stewartville Formations of the Galena Group in Fillmore County, MN. The cave currently functions as an epigenic karst system with allogenic surface water sinking into the cave and a vadose stream running through the cave and resurging at springs a few kilometers away. Passages in the cave are locally controlled by vertical joints in the nearly flat-lying carbonate bedrock, but the water flow directions often do not correspond to the systematic joint directions. The cave contains straight, joint-controlled passages that appear to pre-date the current epigenic drainage systems. These old passages contain hypogenic features and are connected and modified by distinct, younger epigenic passages — often with very sharp transitions back and forth between the two passage types. The epigenic flow incises vadose canyons into the hypogenic passages. The hypogenic passages represent ancient, deep, compartmentalized flow systems that predate the present topography. The concept “ancient” is poorly constrained, however. These ancient cave passages are being reactivated by epigenic processes, while undergoing destruction by general erosion of the landscape.

Introduction
The Devonian, Ordovician, and Cambrian sedimentary rocks of southeast Minnesota (Mossler, 2008) host a variety of caves, sinkholes, sinking streams, blind valleys, large springs, and other karst features. The formations are relatively flat-lying, dip regionally to the southwest at a few m/km, and have been above sea level and subject to erosion since mid-Cretaceous time. All of southeast Minnesota has been glaciated several times during the Pleistocene but has not been covered with ice during the last two major glacial cycles. The highest concentration of karst features is in Fillmore County along the southern border of Minnesota. Fillmore County contains more mapped karst features than all of the rest of Minnesota combined (Gao et al., 2005).

The major cave systems in Fillmore County are concentrated in two different stratigraphic intervals within the Galena Group. Joint-controlled maze caves occur in the Dubuque and Stewartville Formations (the upper red-marked range in Figure 1). Many of these caves currently feed surface water into the subsurface, via sinking streams and infiltration through sinkholes and the soil, and function as flood water mazes. Examples of this group include Mystery Cave (the longest cave in Minnesota), Spring Valley Caverns,
Holy Grail Cave, and Goliath’s Cave. Caves of the second group consist of dendritic stream caves feeding major springs and are developed in the lower Cummingsville Formation, about 65 m stratigraphically below the first group (the lower red-marked range in Figure 1). The second group includes Cold Water Cave in Iowa (the longest cave in the Upper Mississippi Valley Karst) and Tyson Spring Cave, Bat River Cave, and Pine Cave in Minnesota. These caves function to transfer groundwater back to surface streams via springs.

Both types of Galena Group caves currently function as epigenetic systems, contain clear vadose passages and features, and their formation has been explained by previous workers as examples of epigenic speleogenesis. Prompted by Alexander Klimchouk’s hypogenic speleogenesis models (Klimchouk, 2000, 2007), we are in the process of reexamining the first type of Galena caves for evidence of early hypogenic speleogenesis. This work presents our initial observations from Goliath’s Cave.

Goliath’s Cave

Goliath’s Cave is located in S½, N½ of section 3 of York Twp. (T101N, R12W) in Fillmore County, Minnesota. The western portion of the known cave is under the Cherry Grove Blind Valley Scientific and Natural Area (SNA) in the SW¼, NW¼ of section 3 and is owned by the State of Minnesota (see Figure 2). The Cherry Grove Blind Valley accepts drainage from about 2 km² of crop land and pasture via Jessie’s Kill, a first order stream that sinks in the southwest corner of the SNA, via four tile drain outlets along the north edge and northwest corner of the SNA, and via surface sheet flow from all sides. The natural sinkhole entrance to the cave is gated, locked, and is located at 4,825,889 m N, 559,133 m E, (UTM, zone 15). The subsurface drainage in Goliath’s Cave is easterly. The eastern portion of the cave is owned by John Ackerman’s Minnesota Cave Preserve and is accessible via a locked, 76 cm diameter drilled shaft (David’s Entrance) at 4,825,750 m N, 559,452 m E, (UTM, Zone 15). The current mapped extent of the cave is 3.62 km (2.25 miles) and exploration is ongoing.

Figure 1. Stratigraphic column for southeast Minnesota. The red bar labeled 1 is the stratigraphic range of joint-controlled maze caves such as Goliath’s Cave in the Dubuque and Stewartville Formations. The red bar labeled 2 is the stratigraphic range of the dendritic stream caves in the Cummingsville Formation. The cave streams in Goliath’s Cave resurge in Canfield Big Spring in the lower Cummingsville Formation. Modified from Mossler (2008).
The passages in the northern SNA portion of Goliath’s Cave are developed mainly in the Dubuque Formation. The entrance passage from the bottom of the entrance sinkhole (MN23:D04986) is a ~40 m crawl way that can fill rapidly with water and is closed for several months of each year. The walls of many of the passages are broken, angular Dubuque Formation, formed by breakdown. The passage floors are breakdown blocks, secondary sediments, or flowstone. The ceilings of some of the large passages are smoothly sculpted surfaces formed by solution (see Figure 3). At the eastern edge of the SNA, the northern passages descend to stream level, the Rubicon, which is in the top of the Stewartville Formation. Dye traces have demonstrated that the Rubicon drains the northern SNA passages of Goliath’s Cave and the surface water inputs to that part of the cave system.

The passages in the Minnesota Cave Preserve in the eastern portions of Goliath’s Cave are developed in the Stewartville Formation. Most passages have solution-sculpted ceilings and walls, and they either have solution-sculpted bedrock floors or floors covered with thin layers of mud, sand, gravel, and cobbles. An active vadose stream flows easterly along the Rubicon passage (see Figure 4).

The Rubicon extends about a kilometer. A major tributary stream joins the Rubicon from the southwest a few meters downstream of David’s Entrance. Under normal flow conditions that tributary is the primary source of flow in the Rubicon. Dye tracing tests have demonstrated that the water from the sinks of Jessie’s Kill, sinkholes, and surface infiltration in the southern SNA branch of Goliath’s Cave drain via this feeder stream. The side tributary is the last significant input of water to the Rubicon.
A few tens of meters downstream from David’s Entrance the straight Rubicon Passage ends in a bedrock wall. A low oval passage with a sediment floor continues to the northeast and after a few meters the passage abruptly changes back to an easterly heading. From there a 2+ m high walking passage continues with a bedrock floor. The walking passage extends for about a kilometer in a joint-controlled passage (see Figure 5). The stream in the bottom of the passage progressively incises a sinuous vadose channel into the floor. The gradient of the stream increases steadily eastward.

Near the middle of the Rubicon, the vadose stream has incised to the point that the Rubicon becomes two levels vertically. The stream in the lower vadose channel drops over a 1-2 m cascade, steepens further, is actively carving several 0.5-1 m diameter potholes, and then drops over a 5 m waterfall. From there the lower passage becomes a wide stream tube that is filled almost to the ceiling with water. Downstream from the potholes an up-climb of a few meters leads to a well-developed 2+ m high upper level that roughly parallels the lower stream passage. Open crevices in the floor of the upper level connect to the lower stream level at several points.

The ceiling of the upper hypogenic passage has a complex solution-sculpted ceiling, containing closed cupolas that remained dry in the June 2008 flood (Figures 9 and 10). The vertical ceiling joint in this part of the passage contains secondary deposits of chert, which are rare in vertical joints in the Stewartville. A few tens of meters past the waterfall in the lower stream, the upper level passage abruptly ends, and a parallel, smaller, 1-2 m high passage continues eastward. This section of passages does not have a solution-sculpted ceiling. The passage ends in some cross joints near the edge of a surface valley wall.

**Cave Stream Monitoring**

In March 2008 a Campbell CR10 data logger was installed at Goliath’s Cave with a pressure transducer.
and a conductivity and temperature probe in the Rubicon Passage at the bottom of David’s Entrance. A tipping bucket rain gauge was installed on the surface adjacent to the shaft entrance. Data was recorded every half-hour. The data logger monitored conditions in the cave stream, with some gaps due to sensor failure, until March 2011. Hydrograph, thermograph, chemograph, and precipitation records are shown in Figure 6.

The data logger recorded the effects of four spring snow melts in 2008-2011, the spring, summer and fall rains, and a relatively stable period from the summer of 2008 to the snow melt season of 2009. Recharge events, snow melts and precipitation events, cause rapid changes in water flow, chemistry and temperature. All of the recharge events send low conductivity, aggressive waters through the cave system in discrete pulses. The spring snow melts send pulses of cold water through the cave streams. These pulses are only a few hours wide and can lower the cave stream water temperature down to about 1 °C. The base flow stream temperature is about 9 °C. Warm summer rains raise the temperature in the cave streams up to about 17 °C, again in short duration pulses.

In late May 2008 a warm rainy period began that culminated in a 7-8 June 2008 deluge that deposited 35 cm of precipitation in 24 hours onto the blind valley and its surface watershed. Mud and debris lines in the cave indicate that the cave filled to the ceiling all the way to the eastern end of the upper level Rubicon Passage. Mark Seeley (2009, written communication), the Minnesota State Climatologist, stated that this storm has a probability of <0.1%, corresponding to a probable recurrence interval of 1000 to 2000 years. The resulting major flood exceeded the Cherry Grove Blind Valley’s water intake capacity and overtopped the Blind Valley. The pressure transducer over-ranged, but when the site was visited a couple of days later a clear “debris ring” was visible in the entrance shaft 11.5 m above the bottom. The stream passage at the bottom of David’s Entrance is ~2 m high. The rapid response of the cave stream in Goliath’s Cave to surface runoff and shallow vadose flow through field tile drains that empty into the blind valley is illustrated in Figure 6. The flow, temperature and conductivity of the cave waters all record the strong epigenic function of the current water flow system.

![Figure 6](image-url)  
*Figure 6. Stage, temperature, conductivity and rainfall, David’s Entrance, Goliath’s Cave.*
**Cave Morphology**

It is the morphology of the cave passages that preserves the original hypogenic speleogenesis of Goliath’s Cave. There are two end-member types of cave passages which are most evident in the Stewartville passages in Goliath’s Cave. Both types are joint-controlled, but their morphologies are different. The transitions between the two types are often abrupt and not related to any visible changes in bedrock lithology, water flow, or breakdown.

The apparently older passages are discontinuous and end in blank bedrock walls. The old, large passages have complex ceiling features, many of which did not fill with water during the June 2008 flood. The old passages often have solutionally enlarged cross joints. The complex ceiling features and cross joints include many of Klimchouk’s (2007, p. 36) “morphologic suite of rising flow” (MSRF) indicators. There are rising chains of cupolas (both in ceilings and in side joints), complex half-tubes (some of which display cross-cutting relationships at joint intersections) and ceiling outlet features (see Figures 7 and 8). These passages appear to be hypogenic with epigenic overprints.

The apparently younger passages are typically smaller than the older passages and connect the older passage segments into the current, integrated epigenic flow system. These younger passages empty into the older passages both at floor level and from perched openings in the walls of the passages. Where the present cave stream flows through older passages, the stream frequently is eroding a sinuous vadose canyon in the bottom of the older, straighter passages to form a passage with hypogenic features on the ceiling and upper passage walls; and epigenic features on the lower walls and floor. Many of the floors of the epigenic passages are bare, solution-sculpted Stewartville bedrock. The cross joints in the young passages display little if any secondary solution enlargement. In the eastern end of the Rubicon there are several small potholes actively eroding into the epigenic floor of this mixed passage.

In contrast, the floors of the older passages are often armored (covered) with sediment. The sediment ranges from fine-grained muds to coarse-grained gravels and cobbles (Figures 4 and 5). The cross joints in the older passages are often enlarged by solution.
“Knobblies”
When the Stewartville Formation was being deposited the original Ordovician carbonate sediment was pervasively burrowed. The burrows were backfilled with carbonates with different dolomite contents than the original sediment. Those fossil burrows can be seen in freshly-quarried, unweathered surfaces of the Stewartville as a gray mottling. Differential solution (Palmer, 2007, p.155, Fig 6.44) produces a very rough wall surface in the Stewartville Formation. Palmer notes that Wyoming cavers refer to such surfaces as “velcro”. Minnesota cavers refer to such surfaces as “knobblies” or “Stewartville knobblies”. They are an unambiguous field marker for the Stewartville Formation.

As (Palmer, 2007, p. 155) noted, the knobblies occur in two different apparent morphologies: 1) as smooth, solution-sculpted surfaces with cm to mm holes, or 2) as solution-sculpted surfaces with a jagged array of sharp, cm-scale projections. Palmer (2007, p. 155) suggests that “burrows project outward where they were formerly buried beneath cave sediment, but they are indented where they have been exposed to the cave air for a long time.”

In Goliath’s Cave the smoother version of the knobblies are found on the walls of the older Stewartville passages. Palmer’s suggested mechanism, condensation corrosion, is certainly possible, but several other mechanisms are also possible. The relatively smooth, pitted version of the knobblies could have been produced by slow, phreatic, laminar flows or by repeated wetting during flood events.

We do not see evidence in Goliath’s Cave for Palmer’s mechanism for the formation of the jagged, projecting

**Figure 9.** Ceiling bells and channels that remained dry in the June 2008 flood. Width of view is ~ 3 m. Photo by Matt Covington.

**Figure 10.** Solution enlarged side joint with rising cupolas. Note flood line. Width of view is ~ 2 m. Photo by Matt Covington.
knobblies under sediments. We find that the sharp, jagged knobblies characterize the walls, ceilings and floors of the Stewartville passages occupied by currently active streams. In many passages the knobblies grade from projecting to smooth from the bottom to the top of a given passage cross section (for example, see Figure 4). We note that the scale of the burrows is roughly the same as the scale of eddies that form in the turbulent flow through the passages. Thus the jagged knobblies have a similar scale to the scallops expected to form in the passages. Furthermore, many of the jagged knobblies appear very similar to scallops. Might the walls be “resonating” at the worm burrow scale length to produce the jagged projections? That is, given that the scale of variation in the rock and the scale of turbulence in the stream are roughly the same, it is possible that the turbulent flow becomes coupled to the rock variation scale. Independent of the specific mechanisms, our basic hypothesis is that the smooth knobblies record laminar flow conditions while the jagged knobblies record turbulent flow conditions. Note that slow flow conditions would be expected for both the hypogenic formation hypothesis and the floodwater formation hypothesis, since both hypogenic water and backed-up floodwater would typically have low velocities. However, we also note that smooth knobblies are found in ceiling bells that remained dry during the June 2008 flood.

Results
Using the logic and observations above, we have assigned many passages in Goliath’s Cave to one of three categories: 1) Hypogenic passages, 2) Hypogenic passages significantly modified by epigenic processes and 3) Epigenic passages. The spatial relations of these passages are shown in Figure 11 and the “Fig.” labels show where Figures 3-5 and 7-10 are located. The undesignated passages are ones that the authors have not yet visited.

Karst processes in southeast Minnesota have a long geologic history starting syndepositionally in these Ordovician carbonates (Alexander, et al., 2013). During most of these intervening 400+ million years, the carbonate rock was above sea level and subject to groundwater circulation. The current surface and subsurface drainage to the Mississippi River system did not exist for most of that geologic history. The groundwater flow systems before the development of the Mississippi River drainage system unavoidably produced karst features. It is not surprising that some of those old features may still be observed in southeast Minnesota caves.

Acknowledgments
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Goliath’s Cave
MN23-C0085
Fillmore County Minnesota
Map composite of several surveys
Surveyed length - 2.25 miles

Figure 11. Map of speleogenesis passage type in Goliath’s Cave.
Cave and Dawn Ryan and John Lovaas for mapping the CGBV SNA portion of Goliath’s Cave. Betty Wheeler meticulously proofed this manuscript.

References
CREATION OF A MAP OF PALEozoic BEDROCK SPRINGSHEDS IN SOUTHEAST MINNESOTA

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Abstract
Springs are groundwater discharge points that serve as vital coldwater sources for streams in southeast Minnesota. The springs generally emanate from Paleozoic carbonate and siliciclastic bedrock aquifers. Use of systematic dye tracing began in the 1970s and continues through the present as a standard method for investigating karst hydrology and to map springsheds. The work was accelerated in 2007 because of increased funding from the State of Minnesota’s Environment and Natural Resources Trust Fund. A compilation springshed map of dye traces conducted over the last several decades has been assembled for the region.

In southeast Minnesota, the springs are the outlets of conduit flow systems in both carbonate and siliciclastic bedrock aquifers. Conduit flow dominates groundwater transport in carbonate aquifers and is an important component of groundwater flow in siliciclastic aquifers. Conduit flow in aquifers occurs independently of the presence or absence of surface karst features.

The springsheds of these springs have three interacting components: Groundwater Springsheds (analogous to classic karst autogenic recharge areas), Surface Water Springsheds (analogous to classic karst allogenic surface runoff areas), and Regional Groundwater Springsheds.

Surface Water Springsheds can be up to several orders-of-magnitude larger than the Groundwater Springsheds to which they contribute water. Surface Water Springsheds can feed surface flow into one or several stream sinks. Those multiple stream sinks may be in one or more Groundwater Springsheds.

The leading edges of dye tracing breakthrough curves typically show groundwater flow velocities in the hundreds-of-meters to kilometers-per-day range in all of the bedrock aquifers tested. The width and duration of tails of breakthrough curves in these conduit flow systems vary with the bedrock aquifers. The Galena Group has Full Widths at Half Maximums (FWHMs) of a few hours and tails that are down to background in a few days. The Prairie du Chien Group also has FWHMs of hours but has tails that continue for weeks. The St. Lawrence and Lone Rock Formations have FWHMs of months to years.

Introduction
Springs are natural discharge points for groundwater systems. They provide baseflow for streams and are critical sources of cold, relatively constant-temperature water for trout streams. In southeast Minnesota (Figure 1) springs are commonly found emerging from the Paleozoic sedimentary rocks where river valleys deeply incise the water-bearing bedrock layers. In this region, many springs flow from carbonate (limestone, dolostone) and carbonate-cemented siliciclastic layers. These formations dissolve in slightly acidic groundwater and have developed integrated systems of conduits that allow groundwater to flow quickly through the aquifers.

Figure 1. Location map of the state of Minnesota. The study area is indicated by black shading in the inset.
and emerge in the springs. Many other springs emerge from siliciclastic units which contain no significant carbonates. Springs emanating from siliciclastic bedrock aquifers share some of the key hydrologic properties while not exhibiting all of the characteristics of traditional carbonate karst.

In order to conserve and protect springs and the surface water bodies they supply, it is necessary to understand the geologic setting and from where the water is derived. The University of Minnesota (U of M), the Minnesota Department of Natural Resources (DNR), and a group of experienced local cavers have been mapping springsheds in southeast Minnesota since the 1970s. The working hypothesis of these groups is that dye traces provide fundamentally valuable scientific information about how groundwater moves in the subsurface. This can ultimately provide a better understanding of these systems, which is needed to formulate public policies and guide professional decisions. Funding from the State of Minnesota Environment and Natural Resources Trust Fund (ENRTF) has allowed these researchers to accelerate and formalize efforts to delineate springsheds by releasing fluorescent organic dyes into sinkholes or sinking streams to determine the general flow paths to springs. This information was assembled to produce a new map, “Mapped Paleozoic Bedrock Springsheds in Southeast Minnesota” (Green and Alexander, 2014). This map displays the springsheds which were mapped and verified as of June 2014 at a 1:150,000 scale. It also incorporates work done as part of the Fillmore County Geologic Atlas (Alexander et al., 1996). Green et al. (2014) presented the methodology used to define springsheds shown on the map. This paper discusses some of the results of that effort.

A springshed can be defined as “those areas within ground- and surface-water basins that contribute to the discharge of a spring” (Copeland, 2003). Dye tracing is the preferred method for documenting karst springsheds. Kingston (1943) reported the earliest known dye traces in Minnesota karst, which were conducted in 1941 in response to a typhoid fever outbreak in 1940 in Harmony, Fillmore County, Minnesota. When co-author E. Calvin Alexander, Jr. began tracer studies in Fillmore County in the late 1970s, the theoretical model was that of the classic autogenic and allogenic recharge components to karst water systems. This model was reasonable in the well-developed Galena karst of Fillmore County, but there were caveats. Within the county, there are allogenic areas where perennial surface streams flow and then sink to become sinking streams, but they are either: 1) perched on patches of relatively impermeable glacial deposits on top of the Paleozoic carbonates and sandstones; 2) in areas where relatively impermeable shale is the first bedrock; or 3) in areas where the first water table is at or near the surface. There are autogenic areas where surface water flows only for short times after major precipitation events or spring snow melts, and allogenic flows that sink in sinking streams or sinkholes. But in these areas most of the recharge was found to be through the thin soils rather than into sinkholes or sinking streams (e.g., Hallberg et al., 1984). Large areas exist over shallow carbonate bedrock with no perennial surface water flow, but where there are few if any visible surface karst features. Inappropriate projects proposed for such areas prompted innumerable, incredibly unproductive debates and negotiations over how many sinkholes are necessary in a given area to define it as a “karst”.

As we expanded dye tracing work into adjoining counties east and north of Fillmore County in stratigraphically lower carbonates, sandstones and siliciclastic rocks, the allogenic and autogenic models became increasingly untenable. In addition, a third major component of spring flows became more apparent. The Paleozoic aquifers extend westward far beyond the extent of the allogenic and autogenic basins. Regional groundwater from these areas was shown to contain increasingly larger components of flow from the stratigraphically lower springs.

In Green and Alexander (2014) and Green et al. (2014) our conceptual model is described to include three sources of spring flow, i.e., three (at least conceptually) mappable components to each springshed. These three components are shown diagrammatically in Figure 2.

**Groundwater Springsheds (GwS):** Areas where precipitation infiltrates from the land surface to the first conduit-flow dominated aquifer and flows to the spring. GwSs are mapped mainly based on groundwater flow connections demonstrated by tracing techniques. These flow velocities are much faster than typically expected based on porous media models. Our GwS can include classic autogenic basins in addition to areas that don’t fit various definitions of autogenic basins.
The boundaries of these three types of springsheds do not necessarily correspond to those of overlying surface watersheds. In addition, the groundwater springshed boundaries are dynamic, changing as groundwater levels rise and fall, in response to wetter and drier cycles of precipitation.

**Hydrostratigraphic Background**

The flat-lying Paleozoic sedimentary bedrock of southeast Minnesota is Cambrian to Devonian in age (Mossler, 2008) and was deposited as a series of shallow, fluctuating seas advanced and retreated across what is now southeast Minnesota (Figure 3). The Cambrian-age rocks are primarily sandstone, siltstone and shale while the Ordovician and Devonian rocks are primarily carbonate (limestone and dolostone) and shale.

Bedrock layers that serve as the source of water for the many springs found in southeast Minnesota are displayed on the stratigraphic column (Figure 3), The Decorah Shale (dark blue-gray in Figure 4 and 5) is a regional aquitard and is the boundary between the Galena limestone karst system and the Prairie du Chien karst system (Runkel et al., 2014). Many springs discharge...
from the lower Galena because the water cannot move through the relatively impervious, underlying Decorah shale. The St. Lawrence Formation (green in Figure 6) is an important regional bedrock unit from which many springs discharge and where disappearing streams (stream sinks) contribute water to springs (Green et al., 2008; Green et al., 2012).

These rock units are covered by varying thicknesses of regolith (mixtures of tills, outwash deposits and loess). Areas with less than 15 meters of regolith have been the focus of springshed mapping efforts because sinkholes and disappearing streams are primarily found in those areas. Surface karst features are not common where the regolith is thicker than 15 meters (brown areas in Figures 4, 5, and 6). The areas where the thickness of regolith is greater than 15 meters are typically in three settings: isolated areas in the bluffs regions, alluvium deposits along major rivers, and under deep glacial deposits to the west.

The hydrostratigraphy of southeastern Minnesota is further complicated by the extensive development of subvertical joints and fractures, subhorizontal bedding plane parallel fractures, and subtle structural features (Runkel et al., 2014). In addition, Alexander et al. (2013) have emphasized the role that the numerous, long, subaerial unconformities in this section have played in the development of the extensive, pervasive secondary porosity and permeability in these rocks.

**Methods**

The GwSs were defined primarily from sinking stream or sinkhole to spring connections established by tracing techniques. Tracing technology and techniques have evolved significantly from the late 1970s to the present. Tracing agents have included cations, anions, conductivity, environmental stable and radioactive isotopes, injected stable isotopes, lycopodium spores, heat, flow pulses and sediments. The bulk of the tracing work, however, has used fluorescent dyes as the tracers. Several water tracing fluorescent dyes have been used, but our standard dyes were Rhodamine WT, sulforhodamine B, eosin and fluorescein/Uranine.

Initial fluorescent dye traces in the 1970s were conducted using visual dye detection. In the 1980s and early 1990s, Turner Designs Model 10-005 Field Filter Fluorometers were used to analyze the
water and bug samples. In 1993 a Shimadzu RF5000U Scanning Spectrofluorophotometer became our standard dye analysis instrument and that continues to be used at the present (Alexander, 2005).

The dye traces have used a mixture of sample collection methods, such as integrating charcoal detectors (bugs) and direct water samples. The major technological change in direct water sampling was the availability of auto-sampling devices for timed direct water sampling, starting in the 1980s. Both direct water sample and bug trace techniques are still used.

Multiple dyes have been used since the 1990s to trace from different sinkholes or stream sinks at the same time. The dyes were introduced into the groundwater systems through sinking streams, snow melt running into sinkholes, and dry sinkholes. In the latter case, the dyes were flushed with water from a tanker truck (typically 1800–7500 liters). Successful dye traces were conducted in the Lithograph City Formation, Spillville Formation, Galena Group, Prairie du Chien Group, St. Lawrence Formation, and the Lone Rock Formation (Figure 3).

The SWSs were mapped topographically where they contribute surface runoff to a mapped stream sink. The upstream boundaries of surface water basins were identified using watershed maps derived from digital elevation models (DEM) created with Light Detection and Ranging (LiDAR) data or topographic maps (Minnesota Department of Natural Resources, 2014). The catchment area upstream of the stream sink was identified using an ArcGIS tool that selects the surface watersheds upstream of a given point.

**Results**

The Green et al. (2014) map displays GwSs and SWSs identified across southeast Minnesota. The GwS boundaries are outlined in red and record conditions at the time that the dye traces were conducted. SWSs are outlined in yellow. Common boundaries of neighboring springsheds represent surface-water or groundwater divides. Sinkholes or stream sinks that were used as dye-trace input points are symbolized differently depending on whether or not dye was later detected in the springs that were being monitored. The dye-trace vectors (black arrows) are the diagrammatic depiction of the groundwater flow routes.

**Galena Group Springsheds**

Figure 4 is reproduced from Inset 2, an enlarged portion of the Green and Alexander (2014) map showing GwSs and SWSs in the well-developed Galena Group karst of west central Fillmore County. More dye traces have been conducted in the Galena than in all of the other bedrock units combined. Mohring and Alexander (1986) summarized the early dye traces in this area. Many additional traces were included in the Alexander et al.(1996) map which was a precursor of the Green and Alexander (2014) map. Luhmann et al. (2012) report multi-tracer experiments in the area shown in Figure 4. Many, but not all, of the flow vectors in Figure 4 enter the subsurface at sinking streams or sinkholes in the Dubuque and Stewartville Formations and their spring resurgence are in the lower Cummingsville Formation above the Decorah Shale.

The Stewartville and Prosser formations of the Galena Group exhibit the densest sinkhole development in Minnesota. Groundwater flow through these units is dominantly through conduits, solution-enlarged joints, and fractures. Groundwater flow paths from sinkholes and stream sinks to springs in the Galena Group often cross surface watershed boundaries. The low permeability of the rock matrix generally provides negligible flow. Multiple dye-tracing investigations have demonstrated breakthrough travel velocities of 1.6 to 4.8 km per day (Green et al., 2005).

Tracer breakthrough curves through the Galena Group are only slightly asymmetric with small, exponentially decreasing tails. The breakthrough curves typically have Full Widths at Half Maximums (FWHM=the time duration of the breakthrough curve measured between the points on the concentration axis which are at half of the maximum concentration) of a few hours; the tails are often back to background levels in a few days.

Dye traces from sinkholes and stream sinks on the boundaries of multiple groundwater springsheds have demonstrated that those boundaries are dynamic: groundwater flow directions can change in response to recharge events. The springsheds vary greatly in size ranging from several hundred hectares to many square kilometers (Green et al., 2005), have thin glacial cover over bedrock, and commonly have sinkholes. Stream sinks and sinkholes serve as direct recharge points for surface water to enter the limestone aquifer and are good
Figure 4. Map of springsheds in the Galena Group limestone karst. (Inset 2 from Green and Alexander, 2014). The inferred groundwater flow route vectors are diagrammatic depictions of the connections discovered by dye tracing from a dye introduction point (sinkhole or stream sink) to a spring. The dye traces are used to delineate the boundaries of Groundwater Springsheds. While most of the sinkholes and stream sinks transmit water to a single spring, there are several that connect to multiple springs in different directions. These traces mark the boundaries between Groundwater Springsheds under the hydrologic conditions at the time of the dye trace.
indicators of conduit flow in the subsurface. Recharge water also infiltrates through the sediment cover and into the carbonate bedrock. The landscapes between the sinkholes or stream sinks and the springs to which they are connected may lack surface-karst features but are characterized by high velocity conduit flow in the subsurface.

**Prairie du Chien Group Springsheds**

Figure 5 is an enlarged portion of the Green and Alexander (2014) map showing the Prairie du Chien Group GwSs and SWSs of the West Branch of Duschee Creek Valley in east central Fillmore County, Minnesota. The Prairie du Chien Group is stratigraphically lower than the Galena Group, and is the first bedrock in a broad band stretching east and north of the Galena Group subcrop. It extends from the Iowa border northwestward and underlies much of the Twin Cities Metropolitan Area at a relatively shallow depth. The stratigraphically intervening Platteville, Glenwood and St. Peter Formations tend to be present as the first bedrock in only narrow bands which are not extensive at the surface. Typically they are primarily seen along incised stream valleys where a portion of the Galena Group remains as a cap rock.

The Prairie du Chien Group is a major aquifer across much of southeast Minnesota. The visible surface karst features are sparse and are typically further apart than in the Galena karst; large portions of the Prairie du Chien subcrop have little if any perennial surface water flow and groundwater contamination issues are widely present. Alexander et al. (2013) describe the karst conditions that led to the catastrophic sinkhole collapse of three Municipal Waste Water Treatment Lagoons on the Prairie du Chien.

The Duschee Creek Valley area in Figure 5 is of interest because the two large springs at the north end of the valley are the water sources of the Lanesboro State Fish Hatchery, which produces brown and rainbow trout for stocking programs in Minnesota. Thus, these springs provide the water source for an enormously important statewide economic resource.

Dye traces from a large sinkhole 3.4 km south southeast of the Hatchery Spring demonstrated rapid breakthroughs (flow velocity of several km/day) at the Hatchery Springs with FWHMs of hours to a day. But the tails of the breakthrough curves extended for weeks (Wheeler, 1983). This pattern is one of four flow patterns we have observed in traces in the Prairie du Chien Group and is the most common pattern in this hydrologic system.

In 1991, co-author Jeff Green located a critical stream sink in the West Branch of Duschee Creek Valley, 3.7 km south of the Hatchery Spring. Dye traces in 1991 and 1992 documented that the stream sink was a major source of storm water contaminants to the Hatchery Springs. An engineered sealing of that stream sink and rerouting of the stream away from the stream sink eliminated the storm water contamination at the Hatchery Spring (Kalmes and Mohring, 1995).

A second pattern was observed in a 1984 well-to-well trace in the Shakopee Formation of the Prairie du Chien Group (Alexander and Milske, 1986). A breakthrough flow velocity of approximately 2 km/year was observed with a broad, smooth peak persisting about two months, followed a second two-month smaller peak.

A third pattern was observed in dye traces conducted at the Oronoco Landfill in Olmsted County (Alexander et al., 1991). We observed an initial rapid breakthrough spike at monitoring wells, followed by an exponential drop off with a half-life of a few days. The initial spike was followed by a series of precipitation-driven, decreasing spikes over the next couple of years.

The fourth “pattern” is no pattern. About 15 percent of the dye traces attempted in the Prairie du Chien were never detected at any monitoring point (Green and Alexander, 2011).

**St. Lawrence and Lone Rock Formation Springsheds**

Figure 6 is reproduced from Inset 4, an enlarged portion of the Green and Alexander (2014) map showing the three adjacent GwSs and SWSs mapped to date in eastern Fillmore and western Houston Counties, Minnesota.

Groundwater flow through the St. Lawrence is through conduits that include modified bedding-parallel fractures, nonsystematic vertical fractures, and through the bedrock matrix (Runkel et al., 2003). Dye tracing has shown that, while not exhibiting all of the characteristics of traditional carbonate karst, the St. Lawrence Formation has a karst-like conduit flow component. Multiple dye-
Figure 5. Map of Springshed in the Prairie du Chien Group carbonate karst. An enlarged section of Green and Alexander (2014). This figure shows the GwS and SWSs of the two large springs that are the water sources of the Lanesboro State Fish Hatchery. The two southern dye input points (a sinkhole and a stream sink) have been shown by dye tracing to be connected to both springs. The northernmost dye point is a well that had dye introduced below the surface in the annulus prior to grouting. That dye was only detected at the west spring.
tracing investigations in the St. Lawrence Formation have demonstrated breakthrough travel velocities of 90 to 600 meters per day (Green et al., 2008, 2012). These are about an order of magnitude lower than the observed breakthrough travel velocities in the Galena limestone, the best-developed carbonate karst in Minnesota (Green et al., 2005). These velocities, however, are a few orders of magnitude faster than velocities expected for aquifers dominated by porous media flow, and many orders of magnitude faster than expected flow velocities of rocks mapped as aquitards. The breakthrough curves of the St. Lawrence traces rise rapidly to a peak and then fall with FWHMs of months. The dye has been consistently detected at springs for one to three years after it was

Figure 6. Map of Springsheds in the St. Lawrence and Lone Rock pseudokarst, (Inset 4 from Green and Alexander, 2014). The vertical dashed line is the Fillmore County/Houston County boundary. The streams in the Surface Water Springsheds are perennial that sink into stream sinks in the St. Lawrence Formation. Dye traces from these stream sinks have been detected at St. Lawrence and Lone Rock Formation springs. The Groundwater Springsheds boundaries are based on the dye traces and on field observations that indicate the hillsides along the valleys contribute water to the flow system.
GwS boundaries are extended into the upland interfluvies between the stream sinks and springs.

The lack of large conduit networks in outcrop and borehole observations makes classifying the St. Lawrence and Lone Rock as karst problematic. Flow characterization through dye tracing and breakthrough curves for these units is more consistent with their being classified as pseudokarst. Pseudokarst has been described as “landscapes with morphologies resembling karst, and/or may have a predominance of subsurface drainage through conduit type voids, but lack the element of long-term evolution by solution and physical erosion” (Kempe and Halliday, 1997).

Recent work in the siliciclastic bedrock of southeast Minnesota has found that millimeter-to-centimeter-sized, bedding-parallel apertures are connected through an anastomosing network of apertures, clustered along discrete (<2m) stratigraphic intervals and are found at depths exceeding 200 meters (Runkel et al., 2015). The apertures are more limited in size than in karsted carbonate rock. They are commonly associated with distorted bedding which is interpreted to reflect dewatering features that occurred shortly after burial when the rock was only partially lithified. These voids, then, are unlikely to primarily be the result of solution as karst has been traditionally defined (Stewart et al., 2012; Runkel et al., 2015).

Summary

Springs are the discharge points for groundwater systems. Dye tracing to delineate springsheds has been an ongoing activity since the 1970s in southeast Minnesota. This tracing work has been compiled into a new map entitled, “Mapped Paleozoic Bedrock Springsheds in Southeast Minnesota” (Green and Alexander, 2014). This map displays both delineated groundwater and surface water springsheds in carbonate karst bedrock formations, and also in pseudokarst bedrock formations. The map is available on the Minnesota Department of Natural Resources website.

These dye traces collectively demonstrate that in southeast Minnesota:

1. Springs are the outlets of conduit flow systems.

2. Conduit flow dominates groundwater transport in these carbonate and siliciclastic bed rock aquifers independent of the presence or absence of surface karst features.
3. The springsheds of these springs have three components: Groundwater springsheds (analogous to classic karst autogenic recharge areas), surface water springsheds (analogous to classic karst allogenic surface runoff areas) and regional groundwater springsheds.

4. Surface water springsheds can be up to several orders-of-magnitude larger than the groundwater springsheds to which they contribute water. A surface water springshed can feed surface flow into one or several stream sinks. Those multiple stream sinks may be in one or more groundwater springsheds.

5. Breakthrough curves typically show groundwater flow velocities in the hundred-of-meters to kilometers per day range in all of the bedrock aquifers tested.

6. The widths and tails of breakthrough curves in these conduit flow systems vary with the bedrock aquifers. The breakthrough curves of Galena Group have FWHMs of a few hours and tails that are down to background in a few days. Breakthrough curves of the Prairie du Chien Group in the Duschee Creek Springshed also have FWHMs of hours but have tails that continue for weeks. St. Lawrence and Lone Rock Formations breakthrough curves have FWHMs of months to years.

Acknowledgements
The authors gratefully acknowledge very constructive critical reviews of this paper by Chris Smart and two anonymous reviewers. We also acknowledge and thank hundreds of colleagues, co-workers, students, friends and land owners who have participated in the dye traces over the decades.

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References


MEDIA, SINKHOLES, AND THE UK NATIONAL KARST DATABASE

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Abstract
The British Geological Survey (BGS) maintains a number of databases that feed into hazard susceptibility assessments, including karst, landslide, and mining susceptibility. The winter period from December 2013 to January 2014 was one of, if not the most, exceptional periods of winter rainfall in the last 248 years for England and Wales. During this period the Jet Stream diverted easterly tracking cyclones along a more southerly route than is usual across the United Kingdom (UK). This resulted in south-east and central-southern England experiencing total rainfall values of 372.2mm for this period, which was the wettest two-month period since 1910. This period was associated with extensive flooding and increased numbers of slope failures, landslides, and sinkholes, which affected transport routes into and out of London, thereby generating considerable media attention. In addition to government and stakeholder requirements, the BGS experienced an unusually high level of enquiries from the public and the media pertaining to sinkholes, which put an additional strain on resources, but is an acknowledged component of the BGS remit. During February alone, the BGS received reports of 19 sinkholes. The majority of these occurred in the Cretaceous Chalk of southern England. Approximately half were not naturally occurring sinkholes, but were due to the collapse of anthropogenic features. Typically, the anthropogenic subsidence collapse features included: the collapse of chalk shafts associated with historic extraction of chalk for brickworks; the collapse of deneholes (medieval chalk workings for chalk for ground improvement), and chalk mine shaft collapses. This paper describes the National Karst Database, stakeholder requirements, and how the BGS has responded with new and improved mechanisms for data collection, storage and dissemination.

Introduction
Shallow geohazard and risk research at the British Geological Survey (BGS) is focused at a range of scales, from site-specific process understanding to national scale hazard susceptibility assessments. Results of these
assessments are delivered through the product GeoSure. This product was driven by stakeholders (primarily the home insurance-sector) and comprises a series of national datasets providing geological information about potential ground movement or subsidence. More specifically, GeoSure comprises national scale (1:50,000) GIS-derived susceptibility maps for collapsible deposits, compressible ground, landslides, running sand, shrink-swell sediments, and soluble rocks (Figure 1). These maps are aimed at the non-scientific community and incorporate a simple hazard rating (A=low to E=high susceptibility). The main soluble rock types that are encountered in the UK comprise (in order of decreasing solubility and dissolution rate): Triassic salt, Permian gypsum and dolomite, Triassic gypsiferous rocks, Cretaceous chalk, Jurassic limestone, and Palaeozoic limestone. The susceptibility map was built by scoring each of these rock types based on differential weighting of lithology, topography, geomorphology, superficial cover type and superficial cover thickness (Farrant and Cooper, 2008). The knowledge behind the scoring methodology is partly derived from the understanding that comes from maintaining the National Karst Database. It was also enhanced by local knowledge and manual subdivisions. In particular, the evaporites have a slightly different set of determinative parameters that are calculated separately (Farrant and Cooper, 2008).

The National Karst Database, which is being populated and maintained by the BGS, aims to provide data pertaining to the distribution of dolines, stream sinks, caves, springs and incidences of building damage. This data is being retrieved from field slips and paper maps (with additional fine-scale data and ground-truthing derived from fieldwork), from remote sensing techniques such as Light Detection and Ranging (LiDAR) and from existing documentary data sources such as historical and modern Ordnance Survey maps, cave surveys, academic papers and historical documents. Other data such as a legacy “Applied Geology” database (more recently held by Peter Brett Associates) that was originally funded by the Department of the Environment during the 1990s, and the Mendip Cave Registry data are also being used to identify relevant features. This is a time-intensive data gathering exercise, which has been driven by the karst research interests of the scientists involved. Consequently, the data gathering has been very detailed; for example, there are seventeen attribute fields associated with springs.

Both GeoSure and the National Karst Database were designed to capture information on the distribution of naturally occurring karst features. However, during the winter of 2013-2014 an exceptional number of sinkholes were reported in the UK. A significant number of these were found to be anthropogenically derived. They had a significant impact on infrastructure and property, which captured the attention of the media, and consequently, the imagination of the public. This resulted in demands for data delivery comparable with that of the BGS National Landslide Database (Pennington et al., 2015), which was also adapted to incorporate anthropogenic slope failures as well as the core of natural landslides. Here we describe the triggering factors for the unusually high number of sinkhole events, the nature of the events, the demand on resources and how BGS has responded by developing new collaborations and a new portal for data delivery.

Figure 1. Soluble rock susceptibility map with February 2014 “sinkholes” superimposed. BGS@NERC. Contains OS Open data®Crown Copyright and database rights, 2014.
The Unusual Weather Conditions During Winter 2013-2014

On 21 February 2014, the UK Met Office reported on the winter storms of December 2013 to January 2014. They advised that this was one of, if not the most, exceptional periods of winter rainfall in the last 248 years for England and Wales. The rainfall total for December and January was 287.6mm (119.9 and 167.7mm respectively) for the south-east and central-southern England region, which was the wettest two-month period in their records from 1910. During February, there was a further 126.1mm for the south and south-east. Between late January and mid-February, twelve storms hit the south and south-east making it the stormiest period of weather that the UK has experienced for at least twenty years (http://www.metoffice.gov.uk/research/news/2015/uk-winter-storms-one-year-on). The four storms from early to mid-February were particularly severe giving rise to strong winds and huge waves as well as significant rainfall. The sequence of storms was attributed to the position and strength of the Jet Stream, driving a succession of low pressure systems across the Atlantic in a marginally more southerly route than usual. The impact of the rapid succession of events was significant coastal erosion and extensive flooding, particularly of the Somerset Levels. Inevitably this resulted in self-questioning with respect to UK resilience. It is this sustained period of wet weather that is suspected to be the trigger for the spate of sinkholes and collapse subsidence features that were reported, primarily occurring in the south and south-east (Table 1; Figure 1).

Table 1. Sinkhole occurrences in February 2014.

<table>
<thead>
<tr>
<th>Date in 2014</th>
<th>Location</th>
<th>National Grid Reference</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>28 Jan</td>
<td>Walkern Croft, Benington Road, Walkern, Hertfordshire</td>
<td>5285 2250</td>
<td>Collapse of karst flow path</td>
</tr>
<tr>
<td>2 Feb</td>
<td>Walter’s Ash, High Wycombe</td>
<td>4839 1978</td>
<td>Suspected chalk mine shaft associated with former brickworks. Car collapse into subsidence feature</td>
</tr>
<tr>
<td>6 Feb</td>
<td>Duddleswell, East Sussex</td>
<td>5467 1285</td>
<td>Collapse associated with disused soakaway</td>
</tr>
<tr>
<td>11 Feb</td>
<td>M2 between junctions 5 and 6</td>
<td>5942 1590</td>
<td>Suspected denehole. Considerable disruption to motorway traffic.</td>
</tr>
<tr>
<td>12 Feb</td>
<td>Rainham Mark Grammar School, Gillingham, Kent</td>
<td>5806 1668</td>
<td>Collapse of brick capped denehole.</td>
</tr>
<tr>
<td></td>
<td>Upper Basildon</td>
<td>4586 1763</td>
<td>Associated with former brickworks and chalk mine.</td>
</tr>
<tr>
<td>13 Feb</td>
<td>Gillingham Anchorian Rugby Club</td>
<td>5786 1664</td>
<td>Doline associated with feather edge of Thanet Beds over Chalk</td>
</tr>
<tr>
<td></td>
<td>Oakridge Gardens, Wood Lane End, Hemel Hempstead</td>
<td>5075 2078</td>
<td>Associated with former brickworks</td>
</tr>
<tr>
<td></td>
<td>Turner’s Hill</td>
<td>5353 1360</td>
<td>Sand piping</td>
</tr>
<tr>
<td></td>
<td>Warren Farm, Finmere, Buckinghamshire</td>
<td>4624 2332</td>
<td>Collapse of karst flow path</td>
</tr>
<tr>
<td></td>
<td>Holmesdale Grove, Barnehurst, Bexley</td>
<td>5513 1760</td>
<td>Doline associated with feather edge of Thanet Beds over Chalk</td>
</tr>
<tr>
<td>16 Feb</td>
<td>Croxley Green, Hertfordshire</td>
<td>5071 1963</td>
<td>Doline associated with Glacial sand and gravel over Chalk</td>
</tr>
<tr>
<td></td>
<td>Nettlebed, Oxfordshire</td>
<td>4703 1863</td>
<td>Chalk mines associated with former brickworks</td>
</tr>
<tr>
<td></td>
<td>Windermere Road, St Albans</td>
<td>5169 2065</td>
<td>Doline in chalk</td>
</tr>
<tr>
<td>17 Feb</td>
<td>Magdalens’s Road, Ripon</td>
<td>4316 4719</td>
<td>Doline formation due to gypsum dissolution below ground</td>
</tr>
<tr>
<td>20 Feb</td>
<td>Devizes Road, Salisbury</td>
<td>4134 1307</td>
<td>Doline in Chalk</td>
</tr>
<tr>
<td>21 Feb</td>
<td>Wilmington, Dartford, Kent</td>
<td>5509 1707</td>
<td>Doline</td>
</tr>
<tr>
<td>21 Feb</td>
<td>Wilmington, Dartford, Kent</td>
<td>5510 1703</td>
<td>Doline</td>
</tr>
<tr>
<td>22 Feb</td>
<td>Victoria Avenue, Broadstairs, Kent</td>
<td>6380 1696</td>
<td>Associated with brickworks</td>
</tr>
<tr>
<td>22 Feb</td>
<td>Percy Avenue, Broadstairs, Kent</td>
<td>6389 1707</td>
<td>Associated with brickworks</td>
</tr>
<tr>
<td>26 Feb</td>
<td>Bridge House, High Street, Dartford, Kent</td>
<td>5543 1739</td>
<td>3 voids in garden area; suggested possibility of this being related to former leats. Otherwise suspected natural dolines.</td>
</tr>
<tr>
<td>1 March</td>
<td>Chilcompton, Somerset</td>
<td>5160 2065</td>
<td>Collapse of karst flow path</td>
</tr>
</tbody>
</table>
“Sinkhole” Occurrences

 Whilst the true definition of a sinkhole describes a naturally occurring closed depression, the media uses this term primarily to describe any sudden opening of a cavity in the ground. It soon became clear that there was a demand for statistics regarding the extent of media interpreted “sinkhole” occurrences. A new Access database was initiated to serve this need. Twenty-four “sinkholes” were captured in this database for February 2014. The events were divided into eight categories (Figure 2), which can be grouped as 1) naturally occurring pipes in the chalk; 2) natural gypsum-related collapse; 3) shaft or mine collapses associated with former chalk workings (medieval deneholes or more recent workings associated with brick making), and 4) piping-related events on non-soluble bedrock. The naturally occurring dolines in chalk commonly occur close to the feather edge of the overlying Palaeogene strata or where the chalk is overlain by glacial or glacio-fluvial sands and gravel (suffosion dolines). It is suspected that water migrating through the capping materials reacts with pyrite (iron sulfide) in the granular strata, which causes a reduction in pH and therefore greater potential for dissolution of the chalk bedrock (Edmonds, 1983). The high percentage of collapse features associated with former brickworks is attributed to the historic practice of using locally-won wash mill ground chalk as an additive (up to 15%) in brick making. The addition of chalk improved the brick firing process and gave a yellow hue to the bricks.

Media Interest

For BGS the media created a storm of a different kind. The “sinkholes” started to appear in early February and the resultant newspaper headlines captured in Table 2 are typical. Interest seemed relentless, evolving from sensationalism to concerns about the cost implications for individuals. Requests for radio and television interviews of BGS staff were equally numerous. Whilst there was some benefit for BGS from the publicity, the requests placed a huge demand on already stretched resources.

Enquiries and the Public

Sensational headlines triggered the imagination of the British public, the consequence of which was a “flood” of enquiries (Figure 3), each one requiring an individual response. Many enquiries related directly to the reported incidents, requesting information on how to manage and remediate sinkholes, a number related to hazard susceptibility in specific areas underlain by soluble bedrock, some related to other types of foundation problem or subsidence and others were requests for further descriptions of how and where sinkholes occur. Unsurprisingly, there was increased interest from researchers, particularly undergraduate students undertaking independent research projects.

The increased number of enquiries to BGS appears to have stemmed from the amount of media coverage that the BGS was exposed to during this time and also from other public sector organisations providing BGS contact details. As with the media, public enquiries presented an extra strain on resources.

State of Emergency

The UK Civil Contingencies Act dictates what happens during emergencies with a requirement to ensure that relevant organisations collaborate. A state of emergency was declared and meetings of COBR (Cabinet Office Briefing Rooms), the emergencies committee, which includes the key people that are able to make things happen (ministers, senior officials and people from outside government, in this case including representatives from the Environment Agency, Ministry of Defence and Met Office), took place. Between Christmas and 5 February 2014, twenty-one COBR meetings had taken place, primarily to respond to the threat of floods. During February 2014 COBR was being informed by the Scientific Advisory Group for Emergencies (SAGE), which is responsible for coordinating and peer reviewing, as far as possible, scientific and technical advice to inform decision-making. A number of senior BGS staff were called upon to advise SAGE.
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The Government Chief Scientific Adviser particularly required information on hazard susceptibility, the numbers and distribution of “sinkholes” and landslides, and the processes associated with their occurrence to be made available to infrastructure managers.

**How BGS Has Responded**

In order to try and address the multiple needs of stakeholders BGS responded with updated web pages to provide information on what to do in the event of a sinkhole, case study reports on “sinkholes”, and an updated map showing the distribution of sinkholes during February 2014. This enabled the provision of relevant information for the public and media enquiries. Case studies comprised reports on site visits to exemplify the different types of feature. For example, deneholes are medieval chalk extraction pits; characteristically they comprise a narrow shaft with a number of chambers radiating from the base. Chalk was extracted for soil improvement and was usually applied directly to the field, although sometimes it was first burnt in lime kilns to produce quicklime (calcium oxide). The addition of water to quicklime forms slaked lime (calcium hydroxide) a white powder, which can also be used in the preparation of mortar. Once they had reached their limits, the deneholes were commonly capped. A variety of capping techniques were used. Rainham Mark Grammar School, Gillingham, Kent provided an example of a denehole shaft collapse (Figure 4). Interestingly, this shaft was situated on relatively high ground, possibly reflecting the location of a former field boundary in an area with a relatively thin cover over the chalk. Figure 4 also shows the development of small dissolution pipes close to ground surface. Bricks in the base of the hole suggest that this feature may once have been capped with a brick arch. At one of the two northerly occurrences of sinkholes in February 2014, in Ripon, Yorkshire, a naturally occurring sinkhole, which

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**Table 2. “Sinkhole” related newspaper headlines February 2014.**

<table>
<thead>
<tr>
<th>Date</th>
<th>Newspaper</th>
<th>Headline</th>
</tr>
</thead>
<tbody>
<tr>
<td>2 February 2014</td>
<td>Bucks Free Press</td>
<td>Family evacuated from home after sinkhole swallows car</td>
</tr>
<tr>
<td>3 February 2014</td>
<td>Guardian</td>
<td>Sinkhole swallows car in High Wycombe</td>
</tr>
<tr>
<td>12 February 2014</td>
<td>Mail</td>
<td>The Home Counties sinkhole that swallowed a CAR! Family flee village home after their VW disappears into 30ft-deep crater which appeared overnight.</td>
</tr>
<tr>
<td>14 February 2014</td>
<td>News Shopper</td>
<td>Giant sinkhole opens in Barnehurst back garden.</td>
</tr>
<tr>
<td>16 February 2014</td>
<td>Your Maidstone</td>
<td>Two more “sink holes” open in Medway</td>
</tr>
<tr>
<td>17 February 2014</td>
<td>Daily Mail</td>
<td>Terrifying sinkholes that are opening up all over Britain</td>
</tr>
<tr>
<td>17 February 2014</td>
<td>Huffington Post</td>
<td>UK Weather: Homes evacuated as yet another giant sinkhole appears, this time in Hemel Hempstead</td>
</tr>
<tr>
<td>17 February 2014</td>
<td>International Business Times</td>
<td>Sinkholes in the UK: Why is the ground collapsing?</td>
</tr>
<tr>
<td>18 February 2014</td>
<td>The Telegraph</td>
<td>Why sinkholes are swallowing Britain</td>
</tr>
<tr>
<td>18 February 2014</td>
<td>The Guardian</td>
<td>North Yorkshire home wrecked by ninth sinkhole in a month</td>
</tr>
<tr>
<td>18 February 2014</td>
<td>Express</td>
<td>Sinking Britain! Now sinkholes even threaten the dead as they appear in a cemetery</td>
</tr>
<tr>
<td>18 February 2014</td>
<td>Independent</td>
<td>What are sinkholes, how do they form and why are we seeing so many?</td>
</tr>
<tr>
<td>19 February 2014</td>
<td>Metro</td>
<td>Sinkholes: Why are they here and how do they happen?</td>
</tr>
<tr>
<td>20 February 2014</td>
<td>Mirror</td>
<td>Why are so many sinkholes appearing in the UK? All you need to know about the phenomenon</td>
</tr>
<tr>
<td>21 February 2014</td>
<td>The Guardian</td>
<td>Sinkholes on the increase after UK’s wet weather</td>
</tr>
<tr>
<td>24 February 2014</td>
<td>Live Science</td>
<td>Sinkholes swallow more after UK’s big storms</td>
</tr>
<tr>
<td>26 February 2014</td>
<td>Construction News</td>
<td>Sinkholes: can they be stopped?</td>
</tr>
<tr>
<td>4 March 2014</td>
<td>Guardian</td>
<td>Sinkholes: there’s good and bad news</td>
</tr>
</tbody>
</table>
The demands of the media, public and SAGE for “sinkhole statistics” has led to a more proactive and systematic approach to the collection of “sinkhole data” through media searches.

There is a greater urgency to complete the population of the karst database. Further, there has been a focus on developing and adapting a chalk cavity database as an Oracle delivery portal for information on sinkholes and also the karst features captured in the karst database. This will ensure consistent delivery of quality-assured data to all stakeholders, including infrastructure managers, researchers and commercial organisations, or other landowners with geohazard interests. The Oracle database has been adapted to incorporate a greater number of feature types, such as shafts, boreholes, dolines and springs, with a rapid search facility for data extraction purposes.

**Lessons Learned**

Some key lessons have been learned from February 2014:

1. The increasing storminess and reduced predictability of weather patterns has the potential to impact the occurrence of geohazards, including collapse subsidence features.

2. The 24-hour media underpinned by modern technologies further increases the demands on resources at times of increased incidences of geohazards.

3. The data requirements of our stakeholders extended beyond susceptibility mapping to database creation of actual events in order to...
4. Public awareness of BGS services has been heightened, as is evident in the ongoing number of “sinkhole” related enquiries. It is suspected that this is partly due to the reduction in local government funding, resulting in the public seeking new ways of addressing planning and building enquiries.

5. Increased interest from research students has continued, but is tending towards a focus on impact rather than process understanding, which will require further development of our datasets.

Acknowledgements
Banks, Reeves, Ward, Raycraft, Gow, Morgan and Cameron publish with the permission of the Executive Director of the British Geological Survey, NERC and extend our thanks to our colleagues A.H. Cooper and A.R. Farant who developed the karst database.

References


In this paper we present examples of each and discuss their constraints and evidence.

Introduction
As a result of Florida’s statutory requirement for sinkhole insurance coverage, much emphasis has been placed on identification of locations where sinkholes are developing and causing property damage. One issue related to identification of existing sinkholes deals with the origins of shallow, nearly circular to amoeboid depressions in the land surface (Figure 1), which abound in Florida. These features vary from seasonal wetlands, shallow ponds, to indistinct, dry depressions.

One issue that is widely debated is whether or not these shallow depressions in the land surface represent cover-subsidence sinkholes (White, 1988; Beck and Sinclair, 1986) [a.k.a. solution dolines, Field, 2002].

Figure 1. Closed depressions in the Big Cypress Swamp in Collier County, southwestern Florida. Depressions are seasonal wetlands.
Over the last few decades, the authors have conducted over 10,000 sinkhole investigations using surface geophysics, standard penetration testing (SPT), and cone penetrometer test (CPT) methods, studied cross sections of depressions exposed in quarries and borrow pits, and mapped these depressions. This paper presents our observations as to the origins of these depressions. Complete descriptions and data will be presented in a textbook, which is being written by the authors.

Overview
Based on our observations, it is apparent that closed depressions in Florida have a number of possible origins, some of which are karst-related while others can be considered pseudokarst (Field, 2002; Halladay, 2007). We have identified at least seven different mechanisms for formation of these depressions (Table 1), including

1. Cover-subsidence sinkholes over shallow limestone;
2. Suffosion sinkholes over shallow limestone;
3. Cover settlement over shallow shell beds;
4. Large, aeolian deflation areas that resemble “Carolina bays”;
5. Aeolian deflation depressions within dune trains;
6. Depressions that mimic landforms developed on a covered, shallow paleosol; and
7. Depressions created by pedodiagenesis (i.e., conversion of smectite to kaolinite) in a soil-forming environment.

Visual identification of the depression is insufficient to determine its cause. Only subsurface testing combined with petrographic examination of the carbonate fractions in the sediment can truly determine if the depression was caused by dissolution of limestone or shell material.

Furthermore, use of estimates of the rate of development of the depression are insufficient to determine if the cause of the depression is cover subsidence, a process wherein the rate of subsidence is governed by the rate of carbonate dissolution, which takes thousands to hundreds of thousands of years to create volume loss sufficient to cause a depression. Cover-collapse sinkholes are also common in Florida, and the rate of cover collapse can be sudden, occurring in minutes to hours, or slow if the void into which the cover collapses has limited volume or the collapsed materials undergo long-term consolidation.

Mechanisms of Depression Formation
The following sections present the authors’ opinions as to origins and examples of each of these depression-forming mechanisms.

Cover-Subsidence Sinkholes
True, cover-subsidence sinkholes (Table 1) are common in those areas of Florida where limestone is within a few meters of the land surface. They form as the upper surface of the limestone is dissolved away and the cover materials (sand- and clay-rich sediments) slowly subside to replace the volume lost to dissolution. The dissolution surface is often in the vadose zone, but evidence of cover-subsidence in shallow phreatic environments has been observed. Simple SPT boring observations do not allow for determination as to whether the phreatic dissolution surfaces are currently undergoing dissolution or not. A geochemical investigation is required to make this determination.

Even if dissolution is currently underway, the rate of subsidence is governed by the rate of dissolution of the carbonate rock, not by collapse mechanisms. This is of special interest in many areas of Florida where the limestone is overlain by Mio-Pliocene clayey sediments and/or Quaternary marine sand deposits. If the limestone is in contact with clay, dissolution may be limited because of permeability and groundwater flow-path limitations.

In addition to the relative depth of the water table and lithology of the cover material, there appears to be a cover thickness issue that limits the depth to which dissolution of the upper limestone surface can create a land-surface depression. In Florida’s predominantly sandy cover materials, small amounts of limestone dissolution and concomitant settlement of the sand causes dilatation and a slight increase in porosity of the sand cover. This loss of packing density and increase in porosity must be of sufficient magnitude to translate to the land surface and cause a depression to develop.

It is important to understand the difference between cover subsidence and cover collapse. This paper deals only with cover subsidence sinkholes, which form at the rate of dissolution of the underlying carbonate stratum and where the cover materials are sandy marine strata,
### Table 1. Summary of the types and properties of surficial depressions observed on the Florida coastal plain.

<table>
<thead>
<tr>
<th>Depression Type</th>
<th>Mode of Depression Development</th>
<th>Rate of Development</th>
<th>Probable Evidence</th>
<th>Predominant Scale and Shape</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cover-subsidence sinkhole [solution doline]</td>
<td>Soil/sediment subsidence as result of dissolution of limestone surface</td>
<td>Slow (thousands to hundreds of thousands of years)</td>
<td>1. Limestone fragments that show evidence of dissolution</td>
<td>&lt;100 m in diameter; relatively circular unless they intersect each other</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2. Leached sand and/or organics in depression sediments</td>
<td></td>
</tr>
<tr>
<td>Suffosion sinkhole</td>
<td>Raveling of non-cohesive sediment into pre-existing void space in limestone</td>
<td>Varies, may be rapid or moderately slow (decades to centuries) if void space volume limits ability of cover materials to ravel</td>
<td>1. Non-cohesive sediment directly overlies limestone</td>
<td>&lt;10 m in diameter; more or less circular</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2. Disruption of sediment structure in void fill and slopes of depression</td>
<td></td>
</tr>
<tr>
<td>Cover settlement over shell beds</td>
<td>1. Dissolution of shell</td>
<td>Slow (thousands to hundreds of thousands of years)</td>
<td>1. Shell fragments that show evidence of dissolution</td>
<td>&lt;100 m in diameter; more or less circular unless they intersect each other</td>
</tr>
<tr>
<td></td>
<td>2. Minor consolidation and sediment migration into primary porosity</td>
<td></td>
<td>2. Crushing of shells and traces of collapsed and dissolved shell</td>
<td></td>
</tr>
<tr>
<td>“Carolina bays”</td>
<td>1. Lowering of water table, possibly as a result of sinkhole development, allows for fine sand to be eroded by wind stress.</td>
<td>Apparently not forming today; assumed to be formed on the scale of decades</td>
<td>1. Lake, pond, or wetland depression within larger depression at point of maximum deflation</td>
<td>Long axis of large depression is up to 1,000 m. Smaller sinkhole-like depressions within the deflation zone are typically more or less circular and up to 100 m in diameter. Depressions are ovoid with long axis oriented northeast to southwest, the apparent prevailing wind direction;</td>
</tr>
<tr>
<td></td>
<td>2. Low basin forms at upwind end of depression and a parabolic dune train accumulates at the downwind end of the depression.</td>
<td></td>
<td>2. In Florida, the deepest area within the larger depression is located on the southwest end of the northeast to southwest oriented feature</td>
<td></td>
</tr>
<tr>
<td></td>
<td>3. Deflation within the “bay” reveals depressions and possible relict sinkholes in the bottom of the larger, deflation-derived depression</td>
<td></td>
<td>3. Parabolic dunes developed on upwind, northeastern quarter of the larger depression</td>
<td></td>
</tr>
<tr>
<td>Aeolian deflation depressions</td>
<td>Erosion by wind stresses within dune trains</td>
<td>May be rapid depending on wind stresses and vegetation cover</td>
<td>No subsurface expression; deflation zones typically parallel dune long axes</td>
<td>Depressions are complex and may be elongated, oriented parallel to the long axes of the dune train; the long axes of the depressions are typically less than 100 m</td>
</tr>
<tr>
<td>Depressions over paleosol and epikarst features</td>
<td>1. Fine-grained, marine terrace sand deposited over, and infilling, existing depressions developed on the late Miocene to early Pliocene paleosol</td>
<td>Slow (hundreds to thousands of years)</td>
<td>1. Depression floored by paleosol with no evidence of deeper limestone dissolution</td>
<td>Scale varies with circular depressions up to 200 m and streams kilometers in length. Depressions are circular to linear; infilled stream systems are often occupied by modern streams;</td>
</tr>
<tr>
<td></td>
<td>2. Consolidation and minor, early compaction of the relatively thicker sand within the infilled depressions causes development of depressions in the land surface</td>
<td></td>
<td>2. Infilling sediments down warped by compaction; wetland or stream sediments may be included within fill materials</td>
<td></td>
</tr>
<tr>
<td>Pedogenic depressions</td>
<td>Late Miocene to early Pliocene alteration of Miocene smectite to kaolinite, a pedogenetic process, causes volume reduction and land-surface depressions</td>
<td>Very slow (thousands to millions of years)</td>
<td>Clay flooring and bordering depression is kaolinite rich as compared to more distant, smectite-rich sediments</td>
<td>Scale unknown; presumed to be more or less circular</td>
</tr>
</tbody>
</table>
not weathering residue. Cover-subsidence sinkholes, therefore, develop over time frames of hundreds to thousands of years.

Cover collapse occurs as a result of failure of sediments that bridge voids. The collapse may be a result of piping failure or loss of resistance to bridging forces over a void. As a result, cover-collapse sinkholes in the Florida coastal plain develop rapidly (hours to years) in cover sediments that exhibit sufficient cohesion or structural strength to bridge a void. Because of their mode of development, cover-collapse sinkholes can be quite deep.

**Suffosion Sinkholes**

Suffosion or simple raveling without concomitant collapse of non-cohesive sediments into pre-existing void space can also cause small-scale depressions. These are common where the limestone is geologically young, near the land surface, and covered with sand, not insoluble residues created by limestone dissolution. The most notable locations are in the Miami and Big Cypress Swamp areas of southern Florida where limestone is within a meter of the land surface and the cover is non-cohesive sand. Figure 2 illustrates solution holes in the caprock of southern Florida. These solution holes and pipes are commonly sites where sand migrates downward creating small, suffosion-related depressions.

![Figure 2. Caprock, a sandy limestone formed by repeated wetting and drying of shelly sand, penetrated by solution channels through which suffosion of sand occurs. Rock has been turned vertically on edge to serve as "yard art."](image)

**Depressions Caused by Cover Settlement over Shell Beds**

Whether or not depressions caused by cover settlement over shell beds represent cover-subsidence sinkholes is problematic. These depressions are common on land surfaces underlain by late Neogene and Quaternary sand and shell strata. Where they exist and appear to be related to dissolution of shell material, the shell material is observed to be at a minimum of 50 percent of the sediment volume, and within a few meters of the land surface. Observations of hundreds of these features in SPT borings and borrow pits indicate that sediments with less than at least 50 percent shell material (Figure 3A) and deeper than about 3 m do not cause depressions or subsidence at the land surface because of lack of shell dissolution and the volume constraints mentioned under cover-subsidence sinkholes above.

White (1970) discussed the origins of these features in southern Florida and attributed them to both dissolution of carbonate sediments and differential settlement after oxidation and/or compression of organic sediments. He referred to these depressions as sag features. Schmidt and Scott (1984) referred to them as “karst depressions.”

Recognition of shallow subsidence features developed by dissolution of shallow shell beds requires petrographic examination of the shells immediately under these depressions in order to determine if the shells and/or shell fragments have undergone substantial dissolution. Note that a very large proportion of the shell (>>50%) must be removed in order to create sufficient volume reduction for a depression to form. Evidence of dissolution includes rounding of sharp corners, erosion of shell decorations, development of shiny surfaces as if the shells were dipped in acid, and/or development of “punky”, earthy surfaces as aragonitic components are selectively removed from the shell by dissolution. In many examples, the shell has been completely removed and only the “ghosts” of collapsed shells and/or molds of shells remain (Figure 3B).

Siliciclastic sediments in soil zones where the shell has been weathered and subjected to dissolution often contain abundant ferric hydroxide, which gives the sediment/soil a reddish hue, organics, and crude Liesegang banding is often present.
Depressions That Resemble “Carolina Bays”

In west-central Florida there are at least ten large, shallow depressions that resemble “Carolina bays.” Carolina bays are circular to oval wetland depressions that occur on the Atlantic Coastal Plain from New York to Florida. Swarms of the bays have a common orientation, which suggests a common origin, and they often have small, parabolic sand dunes at one end of the elliptical depressions (the presumed down-wind side; Figure 4). The common orientation and dunes suggest that wind, or some other unidirectional transport mechanism played a role in their formation.

The origin of the Carolina bays has been the subject of a long, and sometimes intense, debate. The origins of Carolina bays have been attributed to one, or a combination of, the following processes:

- Meteorite or comet impacts (Prouty, 1935, 1952; Wells and Boyce, 1953);
- Substrate dissolution and sinkhole development (LeGrand, 1953; May and Warne, 2004; Willoughby, 2007);
- Deflation as a result of eolian erosion and transport (Grant et al., 1998; Ivester et al., 2007); and

See also Eyton and Parkhurst (1975) for a summary of these diverse potential causes.

Carolina bays are thought to have formed sometime in the period from 10,000 to 100,000 years ago (Schalles et al., 1989; Brooks et al., 2001; Ivester et al., 2007).

There exist in Florida a number of circular depressions that bear some resemblance to Carolina bays. They are circular to elliptical in outline, occur in clusters, and the depressions show a common orientation from southwest to northeast. They have parabolic, aeolian dunes in the northeast quadrant, the apparent down-wind quadrant of the depressions. Most have small ephemeral lakes or wetlands near the center or southwest third of the depression. The interiors of the wetlands are dotted with what appear to be sinkholes, and SPT testing within the large depressions often presents evidence of sinkhole development and covered epikarst.

We hypothesize that these depressions were formed as a result of localized dewatering of sandy sediments, most likely as a result of sinkhole development during...
the Pleistocene. The dry, fine sand was then entrained by prevailing winds and accumulated in down-wind locations to form the parabolic dunes.

**Aeolian Deflation Depressions**

Florida has extensive, Pleistocene and Holocene aeolian dune fields. These are related to the modern coasts and ancient marine terraces. As is normal, these dunes are characterized by numerous inter-dune and dune-slope depressions. They have more or less circular to linear outlines and present on topographic maps and in the field as closed depressions, often with wetlands.

Because of the closed depressions on topographic maps, they have been mistaken for sinkholes, which are commonly interspersed with the deflation depressions. Only subsurface testing can differentiate them.

**Depressions over Paleosols**

One of the most remarkable features of west-central Florida’s Polk-Desoto Plain Physiographic Province can only be appreciated from the air. As one flies over the area, the large number of circular, wetland depressions and streams with trellis drainage patterns becomes evident (Figure 5).

Observation of these depressions in cross section in phosphate mine cuts and in SPT borings indicates that many of these features are associated with the late Miocene/early Pliocene weathering surface, which resulted in a paleosol that is locally termed the “leached zone” (Altschuler et al., 1951; Van Kauwenbergh et al., 1990).

Pleistocene marine terrace sand is slightly thicker in depressions on the top of the paleosol, and during early sediment compaction the thickness of the sediments dictates the amount of change in thickness of the sediment overlying the paleosol. While the amount of compaction is minor and only changes the relative density of the sediments by small amount, the result is a depression. For example, if the sand body were 1 m in thickness when it was deposited and shortly after deposition it compacted under its own weight, groundwater percolation, and bioturbation by five percent, the resulting thickness would be 0.95 m, which would likely not be visible. If, however, the sediment were 10 m in thickness when deposited, the post-settlement and compaction thickness would be 9.5 m and a 0.5 m depression would be visible. With a shallow water-table aquifer, this depression would likely be seasonally wet. This process has resulted in the patterns of drainage control and shallow depressions that are so common on the Polk-Desoto Plain Province (Figure 5).

This process was identified on the Polk-Desoto Plain by Cathcart (1963) who noted these features and stated that

“The subsurface topography of the Hawthorn is similar to the present surface; ancestors of the present surface streams flowed on the surface of the Hawthorn at or close to their present positions.” (Cathcart 1963, p.1).

In other words, the drainage and depressions on the modern land surface mimic the buried topography of the late Miocene/early Pliocene land surface. Post depositional settlement is greatest where the Plio-Pleistocene sand mantle is thickest. With settlement, new drainage systems and wetlands occupy the resulting depressions. When the drainage ways are rectilinear or trellis-like (Figure 5), they were probably developed on weak fractures or other nearly orthogonal features developed in the more cohesive and carbonate-rich sediments of the underlying Miocene Hawthorn Group.

The most comprehensive investigation of these depressions was conducted at the future site of the C.W. “Bill” Young Regional Reservoir in southeastern Hillsborough County, Florida. Numerous circular

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**Figure 5.** View of a portion of the Polk-Desoto Plain Physiographic Province in Hardee County, central Florida. Many of the wetlands and streams are developed over somewhat thicker sands in pre-existing low areas developed on a late Miocene/early Pliocene paleosol.
wetlands and several small streams crossed the site (Figure 6). A geophysical and stratigraphic study was used to determine the origins and risks of these features to the reservoir (Upchurch et al., 1999; Dobecki and Upchurch, 2010). Every photolinear intersection, each depression, and the stream beds were tested by ground penetrating radar, seismic shear-wave analysis, and seismic reflection and refraction and then drilled using SPT techniques.

As shown in Figure 6, one problematic cover-collapse sinkhole resulted in moving the berm to avoid the risk. All of the other low areas were determined to reflect depressions in the underlying “leached zone” (the late Miocene/early Pliocene paleosol). Two of the depressions, indicated by the ERM boring designations on Figure 6, were found to be over ancient sinkhole depressions. One of the relict sinkholes was filled with Miocene smectitic clay and the other with well compacted Pliocene sand. Neither showed evidence of modern activity. The streams and all other wetland depressions were developed over depressions in the upper surface of the paleosol but had no subsurface expression in the underlying Miocene and older strata.

**Pedogenetic Depressions**

Altschuler et al. (1956, 1963) and Isphording (1984) have suggested that sediment volume reductions accompanying alteration of smectite to kaolinite within the Miocene sediments of Florida have caused shallow land-surface depressions (Table 1). It is likely that these depressions were the precursors of the depressions discussed above since they would have developed during the late Miocene/early Pliocene pedogenetic event.

**Karst or Pseudokarst?**

The discussion above has cataloged seven different forms of shallow depressions that occur on the land surface in the Florida Coastal Plain. The discussion has purposefully omitted cover-collapse sinkholes, which are often relatively deeper and more easily identified than the shallow depressions discussed herein.

It is apparent that the many land-surface depressions in Florida have diverse origins. One cannot simply conclude that natural depressions represent karst conditions. Table 2 indicates which types of depressions are karst related and which are pseudokarst. Only detailed analysis of the sediments within and below the depression will reveal whether the depression has a karst origin or is pseudokarst.

**Recommendations and Conclusions**

Seven different origins of shallow depressions have been identified in Florida. The abundance of these shallow depressions in many areas of the modern Coastal Plain may be dramatic (Figure 1). Determining the origins of these depressions and the risk(s) they pose is often confusing to lay persons and professionals alike.

Based on the authors’ experiences, it is inappropriate to simply observe the field appearance of the depressions

**Table 2.** Types of depressions representing karstic or pseudokarstic processes.

<table>
<thead>
<tr>
<th>Depression Type</th>
<th>Karst or Pseudokarst?</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cover-subsidence sinkhole</td>
<td>Karst</td>
</tr>
<tr>
<td>Suffosion sinkhole</td>
<td>Karst</td>
</tr>
<tr>
<td>Cover settlement over shell</td>
<td>Both exist</td>
</tr>
<tr>
<td>Carolina bay-like depression</td>
<td>Pseudokarst*</td>
</tr>
<tr>
<td>Aeolian deflation area</td>
<td>Pseudokarst</td>
</tr>
<tr>
<td>Paleosol-related</td>
<td>Pseudokarst*</td>
</tr>
<tr>
<td>Pedogenetic</td>
<td>Pseudokarst</td>
</tr>
</tbody>
</table>

* May be associated with karst features such as sinkholes.
and conclude that they represent sinkholes or other karst features. It is strongly suggested that:

1. Only subsurface testing can determine their origin. Visually, the depressions appear similar and depressions with different origins are often mixed in an area.

2. Detailed analysis of the sediment stratigraphy, fabric, and texture is usually required to identify the origins of the depressions.

3. Petrographic, microscopic, and/or mineralogical analyses are often required to determine if shell and limestone has been subjected to dissolution or clays have been altered. Only a geochemical analysis of the pore water can determine if the dissolution is on-going.

References


SINKHOLE VULNERABILITY MAPPING: RESULTS FROM A PILOT STUDY IN NORTH CENTRAL FLORIDA

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Abstract
At the end of June in 2012, Tropical Storm Debby dropped a record amount of rainfall across Florida which triggered hundreds, if not thousands, of sinkholes to form which resulted in tremendous damage to property. The Florida Division of Emergency Management contracted with the Florida Department of Environmental Protection’s Florida Geological Survey to produce a map depicting the state’s vulnerability to sinkhole formation. The three-year project began with a pilot study in three northern Florida counties: Columbia, Hamilton and Suwannee. Utilizing the statistical modeling method Weights of Evidence, results from the pilot study yielded a 93 percent success rate of predicting areas where the geology is conducive to sinkhole formation. Lessons learned and field mapping techniques developed during the pilot study are now being applied to map the entire State’s vulnerability to sinkhole formation.

Introduction
Florida is underlain by several thousand feet of carbonate rock (limestone and dolostone) with a variably thick mixture of sands, clays, shells, and other near surface carbonate rock units. These several thousand feet of carbonate rocks are host to one of the world’s most productive aquifers, the Floridan aquifer system. Natural erosional processes, both physical and chemical, have acted upon these carbonate rocks as water flows through them, both horizontally and vertically, dissolves and physically erodes the rock. Those erosional processes create cavities within the rock. The dissolution and cavity collapse within the rocks has created Florida’s karst topography which is characterized by sinkholes, swallow, caves (wet and dry), springs, disappearing / reappearing streams, and subterranean groundwater flow.

The Florida Geological Survey (FGS) was contracted by the Florida Division of Emergency Management (FDEM) to produce a map depicting the State’s vulnerability to sinkhole formation following a mass sinkhole event triggered by record rainfall from Tropical Storm Debby in June of 2012. The three-year project began with a pilot study in three northern Florida counties: Columbia, Hamilton and Suwannee. Prior to Tropical Storm Debby’s record rainfall, the state had been experiencing a multi-year drought leading to reduced groundwater levels within the Floridan aquifer system. The leading hypothesis is that cavities which may have normally been water-filled had developed unsaturated air space. The lack of hydrostatic buoyancy meant the overburden (the sands and clays over the carbonate rocks) no longer had adequate support and collapsed when the record rainfall from TS Debby added increased hydrostatic loading and lubrication of overburden soils by rising groundwater levels over a very short time period (Figure 1).

Sinkholes are a geological hazard that place people’s property and even lives at risk. Vulnerability of an area to sinkhole formation is dependent upon both natural

Figure 1. Groundwater levels in Suwannee County (Live Oak, FL) prior to and after Tropical Storm Debby.
(geologic, hydrologic, and meteorologic) and human (water pumping, terraforming, ground loading) factors. As Florida’s population continues to increase, the potential for encountering a sinkhole hazard increases.

Current sinkhole hazard maps that are available to Florida Department of Emergency Management (FDEM) are insufficient and poorly substantiated by available geologic data. The FDEM presently relies on two sources of activity: 1) a non-scientific qualitative self-assessment of risk reported by each county, and 2) publically-available and statistically-biased subsidence incidence reports that are broadly generalized to the scale of entire counties without application of a scientifically defensible method.

The FGS maintains a database of voluntarily reported subsidence incidents, which are largely unverified reports of sinkholes; however, other subterranean events can cause holes, depressions or subsidence of the land surface that may mimic sinkhole activity. These include 1) subsurface expansive clay or organic layers which compress as water is removed, 2) collapsed or broken sewer and drain pipes or broken septic tanks, 3) improperly compacted soil after excavation work, and 4) buried trash, logs and other debris. Often a depression is not verified by a licensed professional geologist or engineer to be a true sinkhole, and the cause of subsidence is not known. As such, one of the primary goals of the pilot project was to map existing and recently formed sinkholes for usage as model training point sites.

Criteria for pilot study area site selection
Two important criteria for pilot area selection are geologic diversity and contrast. In order for a model to be tuned and validated using a pilot study in preparation for statewide application, the area chosen must contain a broad range of diversity and contrast.

Geologic diversity is defined as an area that contains multiple geomorphic terrains and districts. Terrains are small geomorphic areas which contain similar landforms formed under similar processes. A terrain is a sub-unit area to a larger area termed a district. Districts are larger generalized regional geomorphic areas which formed under similar processes. The pilot study area (Suwannee, Hamilton, and Columbia) (Figure 2) contains seven geomorphic terrains (Figure 3) and three geomorphic districts (Figure 4). In general, the greater the number of terrains and districts an area contains, the greater the underlying geologic diversity.

Geologic contrast is defined as an area which contains both variable overburden sediment thicknesses and content types and variable depths to carbonate rock. Two

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Figure 2. The pilot sinkhole study area consisted of 3 counties in northern Florida. The area was chosen for two important reasons: geologic diversity and an abundance of data relating to sinkholes that occurred during the approach and arrival of Tropical Storm Debby in June of 2012.

Figure 3. Pilot Study Area Geomorphic Districts: Geologic diversity is defined as an area that contains multiple geomorphic terrains and districts. Districts are larger generalized regional geomorphic areas which formed under similar processes. The pilot study area contains three geomorphic districts.
closed topographic depression targets to visit which may be sinkholes. Those identified sites were termed “points of interest” (POI). POIs were researched using GIS from which a POI GIS layer was created. GIS layers typically used during that process were: digital elevation models (DEM), LiDAR (light detection and ranging) high resolution elevation data, closed topographic depressions (CTDs) (DEM and LiDAR derived), streams, swallets, springs, geology, aerial imagery, the Florida National Hydrologic Dataset (NHD), and subsidence incident reports (SIRs) (a database of unverified sinkholes maintained by the FGS). The POIs serve two purposes. First, POIs may be used as training point sites for future modeling, provided that field investigations find them to be sinkholes. Second, the complete set of POIs serves as a planning tool that helps guide systematic and efficient navigation of the field area.

In order to assure adequate spatial coverage of the pilot study area, the study area was split into two primary grids: a 10 kilometer grid and a one kilometer grid. Within each 10 kilometer grid cell a minimum of four POIs were identified for onsite visitation by field staff. When possible, more POIs were identified within a 10 kilometer cell. The one kilometer grid cells were used as an arbitrary minimum spacing between each POI within a 10 kilometer grid cell. There was no limit to the number of sites documented, although effort was made to traverse at least a kilometer before documenting another site.

**Field Methods**

Fieldwork within the pilot study area was conducted over thirteen days from early November, 2013 through the end of March, 2014. While navigating from one POI to the next, field staff would visually survey both their physical surroundings and the GIS data looking for clues potentially indicating the presence of a sinkhole. If a potential sinkhole was identified or sighted, that site was added as a POI within the associated GIS POI dataset. Effort was then made to investigate that POI.

When on site, efforts were first made to determine whether or not the POI being observed was truly a sinkhole within the best professional judgment of field staff. If the POI was determined not to be a sinkhole, then notes were made in the comments field of the POI GIS layer indicating such. Identification of non-sinkhole features which mimic the topographic profile of a sinkhole was equally important to documenting actual sinkholes. Non-sinkhole features identified during fieldwork included: old rock quarries, old hard-rock phosphate mine pits, borrow pits, test pits, dug drainage ponds, decomposing tree roots and root mats, and cypress domes.

**Pre-fieldwork Site Reconnaissance**

Sinkholes in map view form closed topographic depressions. Therefore, elevation profiles indicating depressed topographic closure may be an indication of a sinkhole. Prior to fieldwork, time was spent researching potential areas that contain similar overburden sediments, such as sand, clay, and carbonate rock; however, sediment type alone is not enough to provide sufficient contrast. The overburden sediments need to vary in content, mixture and thickness across an area to provide sufficient contrast. In general, both the overburden thickness and the depth to top of carbonate rock have great range of variability in the pilot study area. The pilot study area has a broad range of overburden sediment content and mixture. Contrast is important to the Weights of Evidence modeling process as an overall measure of the spatial association (correlation) of an evidential theme with the training points.

**Figure 4. Pilot Study Area Geomorphic Terrains:**

Geologic diversity is defined as an area that contains multiple geomorphic terrains and districts. Terrains are small geomorphic areas which contain similar landforms formed under similar processes. A terrain is a sub-unit area to a larger area termed a district. The pilot study area contains seven geomorphic terrains.
When a POI was judged to be a sinkhole, data was collected and entered into a GIS shapefile. Those data included a GPS location, photos, general comments, and dimensions (which were recorded either via tape measure, laser range finder or measured in GIS using a DEM or LiDAR layer). In some instances, the size of a sinkhole was subjective because the sinkhole’s dimensional boundaries may have been 1) part of a nested cluster which had begun to coalesce, 2) was partly or completely within a stream channel, 3) was obscured by thick vegetation impeding measurement, or 4) infrastructure, such as roads, were built through it and made the sinkhole’s dimensional boundaries difficult to discern. In other instances, some measurements were not able to be made because safety was a concern due to the dangers of the sinkhole itself, livestock, or passing vehicles. When possible, in those instances diameters and/or depths were read from DEM or LiDAR datasets. At some sites multiple sinkholes were documented, and in those circumstances attempts were made to record a range of dimensions under a singular POI site. All measurements were recorded in feet. Distances measured via laser range finder registered in yards and were converted to feet.

**Documented Sinkholes**

Within the pilot study area a total 236 POI sites were visited. 207 of those sites were determined to be sinkholes. The remaining 29 were depressional features determined not to be sinkholes such as old hard-rock phosphate mine pits, borrow pits, test pits, dug drainage ponds, and cypress domes, which can all have circular to semi-circular map profiles and have closed topographic depression elevations. The median diameter of the documented sinkholes was 25.9 meters (85 feet). Diameters ranged from 0.6 meters to 179 meters (2 feet to 587 feet). The median depth was 4.6 meters (15 feet). Depths ranged from 0.3 meters to 18.3 meters (1 foot to 60 feet).

**Observations for future project mapping**

An important goal of the pilot project was to identify a set of karst feature field observations by which sinkholes could be differentiated from similar shaped features non-karst features. Those observations would then be used in the statewide mapping phase of the project. A two-year timeline to map the whole state and a limited project budget meant using investigative geophysical surveys or drilling could not be employed to further confirm if a depression was truly a sinkhole below the subsurface. Therefore, the karst feature field observations are important in order to map existing and recently formed sinkholes with reasonable confidence.

**Karst feature field observations**

Without technical on-site subsurface investigation, only professional judgment can be used to make a reasonable determination based upon known information and site observations.

**Observations**

- Depression has complete topographic closure (i.e. once a liquid or sediment crosses the topographic threshold it cannot flow out)
- Signs of surficial deformation past & present
  - Vertical to sub-vertical soil cracks concentric to depression’s perceived center point which may create a complete to partial ring around the depression
  - Sagged(ing) ground in relation to the near-vicinity ground surface topography
  - Soil creep or slumped(ing) soil
  - Arcing trunks of trees and shrubs attempting to re-straighten to vertical orientation within the depression perpendicular to depression’s perceived center point, indicating soil creep
  - Trees, shrubs, or other vertical features that are leaning or sagging into the depression
  - Exposed rock or semi-indurated sediments which otherwise would not be exposed at near-vicinity ground surface topography
  - Water marks on foliage indicating the depression is actively internally draining
  - Water flow marks on ground which orient sediments or foliage litter towards the lowest elevation(s) within the depression
  - Vegetation showing signs of stress or dying within the depression

**Model explanation, Weights of Evidence modeling technique**

Use of the Weights of Evidence (WoE) modeling technique involves the combination of diverse spatial data that are used to describe and analyze interactions and generate predictive models (for a detailed discussion of this statistical modeling technique see Bonham-Carter, 1994 and Raines et al., 2000). WoE is a data-driven process that relies on mathematical relationships between known occurrences as model training sites to create maps from weighted continuous input data layers. These input data layers, known as evidential themes, are then combined to yield an output data layer (or result of the model), known as a response theme (Raines, 1999). WoE was adapted to mineral potential mapping in a GIS and is based on the application of Bayes’ Rule of Prob-
ability, with an assumption of conditional independence, which occurs when an evidential theme does not affect the probability of another evidential theme (Raines et al., 2000). Although Bayesian theory has been applied to ground-water related issues in recent years (e.g., Soulsby et al., 2003; Meyer and Nicholson, 2003; and Feyen et al., 2004), the specific application of WofE to the potential for sinkhole formation is not known.

When applied in this project, WofE was used to generate sinkhole favorability response themes (expressed in probability maps). These response themes were generated in the Environmental Systems Research Institute (ESRI) ArcGIS version 10.2 environment. WofE was executed using the Spatial Data Modeler Tools (ArcSDM toolbox) which is public domain and available through the ESRI arcscripts pages. The fundamental approach and basic nomenclature of WofE is described in the following sections.

**Study Area**
The initial step in implementing a WofE model is the identification and delineation of a study area extent (i.e., pilot county boundaries). This is a critical step since the area identified is used in the calculation of weights and probabilities throughout the modeling process.

**Training Sites Theme and Prior Probability**
Training sites are locations of known features, also known as occurrences in the literature. In mining applications for example, existing mines are known as occurrences. In an aquifer vulnerability assessment, wells with water quality indicative of high recharge are potential known occurrences. In this study, existing or known, true karst features are considered occurrences. Training points are used in WofE to calculate the following parameters: prior probability, weights for each evidential theme, and posterior probability of the response theme.

Training points are converted to represent a unit area of the study area, such as a grid cell within a GIS application. For the sinkhole favorability model, each cell size represents one square kilometer. The prior probability is calculated by dividing the training point unit area (total number of training points multiplied by 1 km) by the total study area and represents the probability that a training point will occupy any given unit area within that study area, independent of any evidential theme data. In less complex terms, the prior probability is based on prior knowledge of the problem without the benefit of supporting evidence. In the sinkhole study example, prior probability could be described as the proportion of known sinkholes within the study area.

**Evidential Themes**
An evidential theme is defined as a set of continuous spatial data that is associated with the location and distribution of known occurrences, i.e., training points. In GIS terms, an evidential theme is analogous to a data layer or coverage. Evidential themes in the mining example might include the location of hydrothermal ore deposits or proximity to faults. In the sinkhole project, proximity to closed topographic depressions and overburden thickness are examples of evidential themes.

Weights calculated in WofE establish spatial associations between training points and the evidential themes. Depending on the data comprising an evidential theme, in order to deal with random processes, it may be necessary to re-classify the data into categories prior to analysis. This is completed by grouping large sets of data into fewer, more manageable categories that are meaningful. For example, if an evidential theme consisted of a data layer of confining unit thickness divided into one-foot thickness intervals, it might be necessary to classify the data into 3 meters (10ft.) or 6.1 meters (20ft.) intervals to generalize the dataset and make it more manageable and can maximize the spatial association between the map pattern and the point pattern.

Weights are calculated for each evidential theme based on the presence or absence of training points with respect to the study area. A positive weight is calculated for areas that have more points than would be expected by chance; the weight is associated with occurrence of evidence. Conversely, a negative weight would be calculated for areas that have fewer points than expected; the weight is not associated with occurrence of evidence (or non-evidence). A weight of zero indicates that there is no association between training points and the evidential theme, or that the evidential theme is not a discriminating layer.

While performing the initial sinkhole pilot study several data sets were evaluated but not used because they were not discriminating and therefore added nothing to the model. This reaffirms the idea of using a data-driven model versus an expert knowledge model in that two of the layers that were deemed logical as predictors of favorable areas for sinkhole formation did not, in reality, work. These were themed layers depicting the distance to surface streams or surface water bodies since karstic areas are internally drained. Swallets and streams may appear in sinkhole prone areas but they are often dry streams and only flow during heavy rainfall events. The logic is that sinkholes are strongly associated with areas that do not have streams or surface water features.
turns out that some of the water features are sinkholes that breach the water table and are classified as lakes. It may be more accurate to classify water filled sinks differently or look at density of water bodies based on area instead of the presence or absence of either feature. It is also worth noting that the data layers, in their current state, were insufficient as predictor maps and therefore were excluded from this analysis.

Weights can be calculated using three distinct methods: categorical, cumulative ascending, or cumulative descending. The categorical method is used to calculate weights for evidential themes where the theme’s values are not ordered (e.g., a geologic map). The cumulative ascending method is used to calculate cumulative weights in a proximity analysis. In this case areas nearest a training point have a strong association while those farther away have a weak association. In this method, areas represented by smaller values of an evidential theme have a stronger association with training points, and those represented by larger values of an evidential theme have a weaker association with training points. Area and number of points are determined cumulatively from the first class to the last class. This method is applicable for themes where the points are mainly associated with the lower values of the evidential theme (e.g., overburden thickness). The cumulative descending method is used to calculate the cumulative weights from the last class to the first class in the opposite way of cumulative ascending. This method is applicable for themes where the points are mainly associated with the higher values of the evidential theme (e.g., soil hydraulic conductivity).

Generalization of evidential themes follows calculation of weights in the WofE modeling process. Themes are generalized in an effort to establish which areas of the evidential layers share a greater association with locations of training points. During the calculation of weights for each evidential theme, a contrast value is calculated, which is the difference between the positive and negative weights (positive weight – negative weight) described above. Contrast is a measure of a theme’s significance in predicting the location of training points and helps to determine the threshold or thresholds that maximize the spatial association between the evidential theme map pattern and the training point theme pattern (Bonham-Carter, 1994).

Confidence of the evidential theme is also calculated for each class, and equals the contrast divided by its standard deviation (Studentized contrast) for a given evidential theme. Confidence provides a useful measure of significance of the contrast due to the uncertainties of the weights and areas of possible missing data (Raines, 1999). Also, a contrast value that is significant, based on its confidence, suggests that an evidential theme is a useful predictor of training points. Evidential themes that do not meet the minimum confidence level of significance are not included in the models.

Following the calculation of weights, contrast is used as a threshold to generalize or break evidential themes into categories. These breaks delineate which areas of the model for each evidential layer within the study area have more association with the training points. The simplest and most common method of categorizing an ordered evidential theme is to select the maximum contrast as a threshold to determine where to place a break in the evidential data theme thereby creating two categories: one with strong(er) association with the training point theme and one with weak(er) association with the training point theme. In a few cases, more complex statistical contrast patterns are inherent in the data and may justify the creation of multiple classes in the evidential theme data.

Response Theme
Following the generalization of evidential themes, WofE output results are generated and are known as response themes. A response theme is an output data layer showing the probability (posterior probability) that a unit area contains a training point based on the evidence (evidential theme) provided. Areas of higher posterior probability indicate that an area is more likely to contain a training point, whereas areas of lower probability indicate that an area is less likely to contain a training point. As it relates to the sinkhole mapping project, a response theme can be understood as a favorability map that is displayed in classes of relative favorability based on selected karst features used as training points.

A response table is generated during calculation of each response theme and that table contains a list of evidential themes and their respective weights, contrast and confidence (of the evidential theme generalized break). In general, a positive weight (W1) for an evidential theme indicates areas where training points are likely to occur, while a negative weight (W2) for an evidential theme indicates areas where training points are not likely to occur. Contrast is the difference between the highest and lowest weights and is a measure of how well an evidential theme predicts training points. Contrast is also used to rank the evidential themes. Higher contrast values indicate those evidential themes that best predict training point locations and which are more important in the model. Conversely, a negative weight that is stronger than a positive weight indicates that an evidential theme is a better predictor of where training points are not like-
depressions (Figure 6), a layer depicting the difference between the water-table surface and the top of limestone (Figure 7) and soil hydraulic conductivity that utilizes the weighted average of the soil column thickness (Figure 8).

Each of the model evidential layers were calculated against the study area training points. A calculated weights table was used to pick the break between areas that are associated with training sites and areas not asso-

**Pilot study area results**

A preliminary favorability map of the Weights of Evidence Model was generated using four evidential themes that showed the strongest association with the training point theme and therefore were considered the strongest for predicting sinkhole areas. Those layers were overburden (Figure 5), proximity to closed topographic

Figure 5. Thickness of overburden on top of limestone surface: Layer showing the thickness of overburden on top of limestone units susceptible to dissolution. This layer that showed the strongest association with the training sites. In areas where the overburden was 32.3 meters (106ft.) or less in thickness (in red) are considered more closely associated with sinkhole formation. Areas with overburden thicknesses greater than 32.3 meters (106ft.) are not associated with the training sites.
Figure 6. Proximity to closed topographic depressions: Layer showing areas that are proximal to closed topographic depressions. Closed topographic depressions were taken from the USGS 1:24,000 topographic maps and filtered by their circularity index. The layer that showed the strongest association with active karst areas has a circularity index of 0.9 (or 90 percent round) when compared to the area of a circle with the same perimeter. The resulting polygon layer was buffered and then intersected with the training sites in order to show areas that are and are not associated with sinkholes. Red areas are more associated and are generally less than 1,390 meters (4,560.4 ft.) away. Areas with values more than 1,390 meters (4,560.4 ft.) are not associated with training sites.
Figure 7. Difference between groundwater level and the top of limestone: Top of limestone data points are used to create a layer depicting the surface of limestone that is susceptible to dissolution. The layer was subtracted from a groundwater level surface and then intersected with training sites to show areas that are and are not associated with sinkholes. Red areas are more associated with the training sites and have groundwater levels that are generally 0 - 1.5 meters (0 - 5ft.) from the top the limestone. Areas with values more than 0 - 1.5 meters (0-5ft.) are not associated with training sites.
Figure 8. Soil hydraulic conductivity (weighted average): Soils data is intersected with training sites to show areas that are and are not associated with the training point dataset. Red areas are more associated and have hydraulic conductivity rates of 207 millimeters per hour (8.15 inches per hour) and greater. Areas with hydraulic conductivity values between 131.1 and 206.8 millimeters per hour (5.16 and 8.14 inches per hour) are moderately associated with training sites and areas with conductivity values less than 130.8 millimeters per hour (5.15 inches per hour) are not associated with training sites.
Overburden thickness was calculated by taking the top of limestone surface and subtracting it from land surface. Values in the pilot study area ranged from 95.1 meters (312 ft.) thick in the extreme northeastern portion of the region to 0 meters (0ft.) which occurs mostly in the lower lying areas along the major area rivers. Intersecting the training sites with this evidential layer revealed that training sites occurred in areas with 32.3 meters (106ft) or less of overburden (Figure 5).

Closed topographic depressions are taken from the United States Geological Survey (USGS) 1:24,000 topographic maps and are the hachured closed isolines on the map. The depression features were filtered based on an index of circularity or circular index (Denizman, 2003). Since sinkholes tend to be highly circular, filtering by circularity index allows for the removal of closed topographic depressions that are highly linear (e.g., a drainage ditch). The circularity index of a feature is the ratio of the area of a perfect circle with the same perimeter as the closed depression.

In some instances multiple layers can be combined into a single layer to select for complex interactions between layers. For example, the difference between the top of limestone layer and the top of the potentiometric surface are two layers that have been combined into a single evidential theme. The combined layer references the difference between water table surface and top of limestone. The complex layer helps reveal the areas in the pilot area where the top of soluble rock is near the potentiometric surface. Presumably, this is a zone where the hydraulic pumping of the aquifer is most pronounced, thereby actively flushing sediments from cavities within the underlying soluble limestone rock layers (Figure 7).

The rate at which water moves through the soil can be an important factor in locating areas favorable to sinkhole formation. Soil hydraulic conductivity is the ability of the soil to transmit water. Soil hydraulic conductivity values were calculated for each soil horizon based on its thickness and applied to the entire soil column. Values ranged from 7.6 millimeters per hour (0.30 inches per hour) to 887.7 millimeters per hour (34.95 inches per hour).

The four evidential themes were combined in the WofE model to build the response theme, shown in Figure 9. The model revealed a strong contrast depicting areas with favorable sinkhole formation. An independent set of data points, called the Subsidence Incident Report (SIRs) database was brought in as a way of analyzing the results of the model (Figure 10). In the pilot study area there were 261 total sites reported. Of those, 163 or 62 percent fell in the highest favorability category. Another 81 sites or 31 percent were in the highly possible areas. Conversely only 1 of the 261 sites reported fell in an area determined to be unlikely. Overall the model is a better predictor of where the geology is not favorable for sinkhole formation than where the geology is favorable.
Figure 9. Results from preliminary pilot study with training sites: Weights of evidence output map from combining the four evidential themes: overburden, proximity to closed depressions, difference between water table aquifer and top of rock and soil hydraulic conductivity.
Figure 10. Results from preliminary pilot study with Subsidence Incident Reports: The Subsidence Incident Reports data was used to analyze the results of the modeled WoE response theme. 93 percent of the reports fell into the highest and highly probable areas where the geology is favorable to sinkhole formation.
References
A SEMI-AUTOMATED TOOL FOR REDUCING THE CREATION OF FALSE CLOSED DEPRESSIONS FROM A FILLED LIDAR-DERIVED DIGITAL ELEVATION MODEL

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Abstract
Closed depressions on the land surface can be identified by ‘filling’ a digital elevation model (DEM) and subtracting the filled model from the original DEM. However, automated methods suffer from artificial ‘dams’ where surface streams cross under bridges and through culverts. Removal of these false depressions from an elevation model is difficult due to the lack of bridge and culvert inventories; thus, another method is needed to breach these artificial dams. Here, we present a semi-automated workflow and toolbox to remove falsely detected closed depressions created by artificial dams in a DEM. The approach finds the intersections between transportation routes (e.g., roads) and streams, and then lowers the elevation surface across the roads to stream level allowing flow to be routed under the road. Once the surface is corrected to match the approximate location of the National Hydrologic Dataset stream lines, the procedure is repeated with sequentially smaller flow accumulation thresholds in order to generate stream lines with less contributing area within the watershed. Through multiple iterations, artificial depressions that may arise due to ephemeral flow paths can also be removed. Preliminary results reveal that this new technique provides significant improvements for flow routing across a DEM and minimizes artifacts within the elevation surface. Slight changes in the stream flow lines generally improve the quality of flow routes; however some artificial dams may persist. Problematic areas include extensive road ditches, particularly along divided highways, and where surface flow crosses beneath road intersections. Limitations do exist and the results partially depend on the quality of data being input. Of 166 manually identified culverts from a previous study by Doctor and Young in 2013, 125 are within 25 m of culverts identified by this tool. After three iterations, 1,735 culverts were identified and cataloged. The result is a reconditioned elevation dataset, which retains the karst topography for further analysis, and a culvert catalog.

Introduction
The identification of closed depressions within a landscape is important for karst studies; however the automated creation of closed depression catalogs are hampered in urban locations. Automated depression delineation and cataloging methods have been discussed elsewhere (e.g., Lindsay and Creed, 2006; Zandbergen, 2010; Doctor and Young, 2013). Of these methods, the filling method (subtracting an initial digital elevation model (DEM) from the DEM with depressions filled to spill points) tends to be preferred due to the ease and speed by which a catalog can be generated. However, in populated areas simulated water flow tends to pond behind roads, railways, and other man-made surface features resulting in falsely detected closed depressions. These problems are especially acute when using high-resolution topographic datasets such as Light Detection and Ranging (lidar), but are not unique to these DEMs. There are a number of methods to deal with these false depressions, but care has to be taken when working within karst terrain as these depressions are likely to be real and of interest to further research (Zandbergen, 2010; Lindsay and Creed, 2006).

Current means to remove topographic barriers use either manual or digital methods. Manual methods require the digitization of lines (here collectively referred to as ‘culverts’) representing underpasses beneath roads, driveways, railways, or other obstructions to actual stream flow using either aerial or field observations. As to be expected, these methods are very time consuming. Following the digital creation of a culvert inventory, this dataset is then used as input into a variety of techniques to ‘cut’ or ‘burn’ the culverts into the elevation data. This
’burning’ method normally uses the entire stream network to hydrologically condition the surface (e.g., Mitrasova et al., 1999; Maidement, 2002; Tarbonton, 2012). However, these methods can be difficult to implement, or can create steep canyons in the hydrologically corrected elevation data potentially causing errors in further analysis, particularly when streams represented as vector lines do not accurately match the elevation surface. Current methods are too aggressive for karst terrain resulting in true depressions being scrubbed from the elevation dataset. For example, the optimized cut and fill tool (Jackson, 2013) would result in the removal of true depressions. Additionally, other tools such as those developed by Poppenga et al. (2010) require thresholds (i.e. area and depth) be set which could result in smaller, artificial depressions being missed.

The goal of this study was to develop an ArcGIS toolbox to identify potential culvert locations which could then be enforced into the elevation data, thereby removing false closed depressions within karst terrain. This is done by finding the intersection of transportation routes and stream lines which are then buffered and used to lower the elevation across the man-made topography. This method minimizes the extent of DEM modification in order to retain the karst depressions. A semi-automated approach is taken with user-provided data and thresholds.

**Study area and previous work**

The study area is the Boyce 7.5-minute quadrangle predominantly covering Clarke County with smaller portions of Warren and Fredrick Counties in Virginia (Figure 1). The region spans roughly 150 km². Located within an extensive karst region of the Great Valley physiographic province of the Appalachian mountain range, the Boyce quadrangle covers part of the Shenandoah River drainage basin.

Details on the geology of the quadrangle can be found in Edmundson and Nunan (1973). Karstification in the study area has resulted in a mature dissected karst surface of moderate to low relief, with 90 m total elevation range and a mean elevation of 180 m above sea level. Sinkholes and other karstic depressions generally occur as a result of cover-collapse or suffosion processes within the residuum overlying the carbonate bedrock;

![Figure 1](image-url). (A) Digital Elevation Model of the Boyce 7.5-minute quadrangle with a draped hillshade. Intersections between NHD Streams, railways and roads. (B) An initial fill difference raster map illustrates ponding behind artificial dams created by railways and roads. The fill difference raster is created by subtracting the initial DEM from the filled DEM. (C) The second and third iterations of intersections identified by the Cutter tool are shown. Note the increase in intersections found between both of these iterations, which directly corresponds to the different flow accumulation thresholds (400,000 m² and 10,000 m²) used.
the thickness of residuum varies between zero and ten meters.

Doctor and Young (2013) presented an evaluation of an automated workflow for identifying closed depressions in the Boyce quadrangle using a simple fill-difference method. This method uses the difference between a ‘filled’ DEM raster and the original raster, and was compared to manually-delineated closed depressions. They concluded that the primary hindrance to a fully-automated process for identifying closed depressions was the presence of artificial ‘dams’ in the elevation surface where streams pass beneath transportation routes. For that study, stream underpasses, usually in the form of culverts, were identified from aerial imagery and field work, and were manually added as an input layer used to recondition the original DEM allowing streams to flow through elevation obstructions such as those created by transportation routes. Although this approach was effective, it was tedious and was not successful in identifying all possible culverts within the quadrangle thereby impacting the closed depression catalog count and morphometrics. Thus, a new automated method to identify stream underpasses was deemed necessary.

***Methods***

The method presented here for reducing the creation of false closed depressions was developed using ArcGIS tools. For this work, ArcGIS 10.2.2 was used, but the models should work with any 10.X version of ArcGIS. Additionally, a license for the Spatial Analyst Tools is necessary.

Three vector datasets were acquired covering streams, railways and road networks along with a high-resolution lidar raster data set. Stream data were acquired from the United States Geological Survey’s (USGS) National Hydrologic Dataset (NHD). Railways and roads were acquired from the USGS National Map. All vector datasets were digitized at the 1:24,000 scale, which is much coarser resolution than the lidar data. This resulted in the need to manually correct the railway shapefile (Figure 2). The resulting corrected vector data was used as input into the tools. The lidar dataset was acquired between 1 March and 9 March 2011. Acquisition took place during leaf-off conditions and after snow cover melted. The vertical RMSE was 9.0 cm while the point spacing was 1.0 m.

The Hydrocutter toolbox contains two tools: Hydro and Cutter. Together, these tools can be run iteratively in a semi-automated way employing user thresholds. Hydro simply implements a stream definition method as determined using the Fill, Flow Direction, and Flow Accumulation tools provided by ESRI within the Spatial Analyst toolbox. The result of the Hydro tool is a vector dataset which can be used as an input into Cutter. Cutter topographically enforces culverts where streams pass beneath topographic highs, generally along transportation routes, thereby providing a flow path across the artificial obstruction in the DEM and reducing false closed depressions.

The Processing Extent, Snap Raster, and Cell Size are all set to the DEM provided by the user within the tool environments automatically so that they do not need to be set by the end-user.

The workflow for the Cutter tool is outlined by Figure 3 and described in detail here. Transportation vector datasets are intersected with an initial NHD stream vector dataset. This results in two point datasets (one for railways and one for roads, respectively) which are then merged into a single point dataset representing all intersections with streams (i.e. culverts). These culverts are then buffered by a user-defined value which should span the widest railway or road. For this study, a 25 m buffer diameter around each point was used. The circular buffer polygons are then used to clip the stream vector line to provide stream segments where the streams cross the transportation routes. A Zonal Fill tool is then used to find the minimum elevation along the clipped vector segments (ESRI, 2012). This results in a raster layer which

**Figure 2.** Data quality is important to consider when running Hydrocutter toolbox. Note the issues with the railway presented here. The initial railway data (dashed red line) is offset from the edited railway (dark brown solid line).
is used to enforce this minimum elevation value into the DEM thereby hydrologically conditioning it.

This conditioned DEM is used as input into the Hydro tool. The only additional information that must be provided when running the Hydro tool is a threshold value for the flow accumulation. For this study area, a flow accumulation value of 400,000 m$^2$ was found to reasonably approximate the lengths of the NHD stream vector lines and was therefore used for the initial iteration of this tool.

The Hydrocutter toolbox was run on the lidar data covering the Boyce quadrangle. The initial iteration of the Cutter tool used the NHD streams and edited roads and railway data as input. The resulting DEM was then used as input into the Hydro tool and a flow accumulation threshold of 400,000 m$^2$ was used to approximate the surface flow accumulation of the NHD streams.

A second hydrologic conditioning iteration was carried out using newly defined 400,000 m$^2$ flow accumulation streams and transportation routes, and then the Hydro tool was repeated using a flow accumulation of 10,000 m$^2$. After the DEM had been reconditioned to account for intersections between the stream lines of 10,000 m$^2$ flow accumulation, stream lines at 2,000 m$^2$ flow accumulation were generated for visual comparison and validation.

Given the different approaches, the closed depressions catalogs identified after running the Hydrocutter toolbox described here and those of Doctor and Young (2013) are not directly comparable; however, here we heuristically compare these datasets.

**Results**

Using the Hydrocutter toolbox approximately 14 times as many culverts and stream intersections with transportation routes were found as those identified manually within the Boyce quadrangle. Of these, 75% were coincident with culvert lines from Doctor and Young (2013) within a 25 m buffer zone.

The 2,000 m$^2$ flow accumulation stream lines often represent flow routes across the DEM surface that do not have any obvious geomorphic expression of surface runoff, such as channels, swales, gullies, etc. Thus, stream lines having flow accumulation values less than 10,000 m$^2$ are not used here to define additional culverts. As a result, the 10,000 m$^2$ flow accumulation stream lines identified the majority of topographically evident stream channels as well as creating an inventory of potential culverts (Figure 4).

Doctor and Young (2013) manually identified 166 culverts of which 106 were verified by aerial imagery or field checking. Of the manually identified culverts 125 are within 25 m of the Hydrocutter culverts, 80 of these were verified. The first pass of the Hydrocutter toolbox found 260 intersections between the NHD stream vector lines and the transportation vector lines; these intersections were indicative of possible stream culverts. These culverts were burned into the original DEM, and used to generate a new stream vector line dataset that was more representative of the actual lidar-derived elevation model. The next iteration used lidar generated stream lines with a flow accumulation of 400,000 m$^2$ thus mimicking the original NHD vector stream lines and identified 272 culverts. The final iteration, using a flow accumulation of 10,000 m, identified 1,735 culverts.

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**Figure 3. Work flow of the Cutter tool within the Hydrocutter toolbox.**
Figure 4. (A) Red points indicate the initial intersections found between NHD stream data with roads and railway data. These are compared to (B) green and yellow points which are the intersections found between the 400,000 m² and 10,000 m² stream lines respectively and transportation routes. As indicated by panels A and B, the stream lines with the lower flow accumulation extend further upstream and therefore intersect a greater number of roads. (C) A detail of these differences is illustrated between initial intersections and subsequent iterations of the Hydrocutter tool resulting in differences of stream dataset quality highlighted particularly in the left half of the panel.
Conclusions

The Hydrocutter toolbox provides two products useful in karst studies. The first is a hydrologically conditioned DEM by removing obstructions to flow routing that cause false closed depressions resulting from the impoundment of water behind man-made features such as roads. The second is an inventory of potential culverts which is not only useful to karst studies, but the broader geologic, hydrologic, and environmental management communities.

Nevertheless, there are several known issues and pieces of cautionary advice that come with using this tool. First, the user-defined buffer distance around intersections needs to be large enough to allow impoundments to be breached. This can be problematic when road widths are variable and span a greater width than the buffer distance used to define the ‘cutting’ of culverts. Second, in the Boyce quadrangle a flow accumulation of 10,000 m$^2$ was determined as reasonable to represent surface water flow routes when compared to the topographic expression of the surface hydrology; however, this is not a constant value, meaning that a user should determine a reasonable flow accumulation value empirically for their study area by examining the correspondence between the elevation model and the stream lines generated. Third, errors due to poor vector data quality compared to the lidar data can be propagated through the analysis. Therefore, high accuracy of the vector data used as the initial inputs to the process is important. If good vector datasets are unavailable, manual editing to 1:24,000 scale vector data to match the lidar elevation model might be necessary, as was done here. Fourth, some re-routing of stream lines can occur between iterations of the Hydrocutter tools. This is localized to areas within or near the buffer zones of intersections (Figure 4).

The process outlined here generally improves the overall representation of flow across the DEM. It is only problematic in areas where the initial road or NHD stream data is poor. Thus, as with many tools, the quality of the data being input into the tool is inherently representative of the data quality coming out of the tool.

Although further quantitative comparison is necessary, the Hydrocutter toolbox is better at breaching man-made impoundments while preserving the natural closed depression landscape within karst terrain (Figure 5). This is vital to creating closed depression catalogs generated from lidar datasets rather than statewide inventories which adequately represent the closed depression population while minimizing false detections (Wall & Bohenstein, 2014).

Figure 5. Depressions identified by Hydrocutter are compared to manually identified depressions. Note that some of the manually identified closed depressions are grouped into one closed depression using the results of Hydrocutter suggesting a coalescence of closed depressions which are not easily identified by manual interpretation.

A possible future refinement of the toolbox would be to improve the manner in which the location of a stream pathway is delineated across an impoundment. Using the current Cutter tool, the pathway follows the pre-existing stream line that crosses an obstruction. If the stream vector is not in the correct location (i.e. where a culvert or underpass is), then errors may result which could be propagated through the analysis. Clipped streams segments may not fully connect the actual stream channels in the lidar surface. Figure 6 illustrates a possible solution which would employ a least-cost path approach within the buffer zone of intersection to optimize the likelihood of connecting the lowest point on either side of the obstruction (Poppenga et al., 2010).

Future work will focus on implementing a hybrid method between the Cutter tool and least cost path technique. This will allow for more accurate connections between low points within the buffer zone. Ideally, this will be more representative of stream flow routes and culvert locations. It will have an added benefit of further reducing false closed depressions.
Acknowledgments and Disclaimer

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Figure 6. Future development for this toolbox is illustrated by (A) The original elevation surface shown with the NHD flow line and road intersection, (B) grid cells matching the NHD were recalculated with the minimum elevation value within the buffer zone to create hydrologic connectivity. The red areas indicate false depressions that remain, (C) The least cost path through the buffer zone is used to identify the cells to calculate as the flow path through the topographic high.
HISTORY AND FUTURE OF THE MINNESOTA KARST FEATURE DATABASE

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Abstract
Since the 1990s the University of Minnesota and the Minnesota Department of Natural Resources have maintained a karst features database that is used to conduct research on karst processes and inventory karst features. Originally designed as a tabular database only, the karst features database developed into a spatial database in 2002 with tabular data stored in Microsoft Access and a spatial component managed in ESRI ArcView. In 2012 the database was converted to a single, relational database platform, PostgreSQL, with both tabular and spatial components edited in ESRI ArcMap. Custom editing forms are written in Visual Basic and are accessed in ArcMap sessions by ESRI add-ins. The current database infrastructure allows for remote editing. Read-only versions of the data are available in GIS/spatial format for public use via web services. Future development plans include links to water chemistry data, water level measurements, and other ancillary data; along with the addition of vectors to represent dye traces and polygons for larger karst features.

Introduction
Karst is recognized as a term describing both distinct landscapes—karst terrains—and distinctive hydrology related to the movement of water in soluble bedrock—karst processes. The construction of a karst features database that adequately documents both karst terrains and karst processes for researchers, regulators, and planners is a formidable task. How do uses and potential abuses impact database design and content? What should be in such a database? How does data get in, or out? While the Minnesota Karst Features Database (KFD) has been primarily research oriented, these broader questions have guided past and current database development and will continue to guide development going forward. This paper documents the history and future of the KFD, with the goal of providing the reader a better understanding of how it came to be and where it is going.

History and Methods
The Minnesota Speleological Survey created the database in the early 1970s as a sinkhole inventory. Sinkhole locations were collected on 4-by-6 inch index cards with unique identifiers, and plotted on 1:24,000-scale USGS 7.5-minute topographic maps (Alexander, 2015). About one hundred sinkholes were mapped in this manner, and this process continued into the early 1980s. Many sinkholes in Minnesota, especially those several meters or less in diameter, are ephemeral features that appear in fields and are filled, if possible, to minimize disruption of agricultural practices.

As personal computers and spreadsheet software became available in the 1980s the evolution towards fully functional geographic information systems (GIS) man-
agement of karst features data began. The first targeted sinkhole inventory for a specific Minnesota county was begun during that timeframe (Dalgleish and Alexander, 1984). The Winona County sinkhole inventory involved a systematic survey of landowners across the county. Basic location information was recorded for each sinkhole such as who found it, how and when it was found, along with some estimate, if known, of when the sinkhole first appeared. Sinkhole physical attributes were recorded, including width and depth and morphology (steep-walled or shallow-walled). Sinkhole contents were also recorded.

A total of 535 sinkholes in Winona County were mapped in this manner, each with their own unique identifier; many filled sinkholes were also reported. Data collected in 3-ring binders were later entered into fixed-format text for keypunch services and then loaded into single text files with Fortran-based retrieval and reporting capabilities (Figure 1). Sinkhole locations were digitized and stored on a main-frame computer at the University of Minnesota; allowing sinkhole distributions to be plotted and a limited number of sinkhole attributes to be displayed in map form. Formatted text records were eventually transferred to personal computer spreadsheet software.

By the mid-1980s the field-based sinkhole inventory combined with geologic mapping led to two regional observations (the first self-apparent but not documented in map form): 1) sinkholes occur where the landscape is underlain by soluble carbonate rock; 2) sinkholes occur where the bedrock is covered with less than 50 feet of sediment. Both trends were displayed as an unpublished, first-generation regional map of Minnesota Karst Lands in 1992 (Alexander, 2015; Figure 2).

In 1995 the distribution of sinkholes in Winona County Minnesota was revisited (Magdalene, 1995). Locations were digitized, and data were managed in a Microsoft Excel spreadsheet. Six hundred fourteen sinkholes were mapped, including 34 new and 39 previously unreported sinkholes. Sinkhole attributes from the earlier table structure were reviewed and refined. Combining the KFD and geologic mapping as part of the County Geologic Atlas Program Magdalene showed that sinkholes are clustered. In addition, higher densities of sinkhole occurrence were linked to a specific bedrock stratigraphic position—the contact between the Oneota Dolomite and Shakopee Formation within the Prairie du Chien Group.

Figure 1. Early sinkhole data keypunch form, Karst Features Database, 1980s (provided courtesy of T.E. Wahl).
By that time, GIS software for personal computers, PC ARC/INFO (ESRI, 1987), was fully in use in Minnesota for resource mapping. A county geologic atlas project had just begun in Fillmore County, where sinkholes and springs occur in greater concentration than anywhere else in the state. Karst features mapping, which had focused largely on sinkholes, now expanded to include springs. Springshed mapping, based on dye trace results, had already been underway for a number of years, and the KFD became the primary database for managing dye input and output locations and compiling groundwater flow routes to identify springshed boundaries.

Many of the sinkholes in Fillmore County are large enough to be visible on 1:24,000 USGS 7.5 minute topographic maps. Points were added to the KFD by digitizing closed depressions on the maps. This process captured features with a minimum size (width) of approximately 25 meters. Points were also added from the 1951 Fillmore County Soil Survey. Combined, these two sources added approximately 4,000 sinkholes to the KFD. Sinkhole distributions and depth-to-bedrock data, searchable in a GIS environment, were used to create a sinkhole probability map of the county (Alexander et al., 1995). Attribute tables were developed for springs and newly acquired and historic spring data were added to the KFD.

The period of 1998 to 2003 saw advances in spatial data technology, including the incorporation of global positioning system (GPS) equipment in standard fieldwork and increased accessibility of geospatial data, such as current and historic aerial photos. The development of ArcView for personal computers, along with its scripting language Avenue facilitated the development of custom user interface forms for data entry and editing and supported the automation of geoprocessing for more complex spatial analysis. During this period, sinkhole locations and depth-to-bedrock data were used to create a sinkhole probability map for Goodhue County (Alexander et al., 2003); digital elevation models of bedrock stratigraphic units were used to assign stratigraphic positions for sinkholes and springs in Wabasha County (Tipping et al., 2001); the KFD expanded beyond south-
eastern Minnesota to document sinkholes and springs in the sandstone of Pine County Minnesota (Shade, 2002); karst terrain was mapped in distinct units in Mower County based on surface and subsurface drainage characteristics, bedrock geology, depth to bedrock and land surface topography (Green et al., 2002a; 2002b). In all instances, the KFD was used to inventory new features and sinkholes that are now filled using historic aerial photos and soil surveys.

The greatest expansion of the KFD occurred during the period of 2003 to 2005 when funding became available to conduct regional karst investigations and concurrent database development, resulting in a fully functional karst features database (Gao and Alexander 2003; Gao et al., 2002; 2005a; 2005b; 2005c; Gao, 2008). The project formalized the database structure, code tables, and metadata that are currently in use. A combination of ArcView and Microsoft Access platforms were used, with location information stored in ArcView shape-files and attribute information stored in tables within an Access relational database. Custom user forms were developed for entering and editing data, along with report writing capability.

The project proposed structures for database features not yet implemented, including dye trace vectors and polygons to delineate karst features over large areas, such as sinkhole clusters. The project also proposed conceptual models for future database use including: spatial analysis; data mining; geostatistical analysis and descriptive analysis; and hydrogeologic analysis such as springshed delineation and springshed water budgets.

Several regional analyses were conducted by Gao (2002) to demonstrate the KFD as a research tool. Nearest neighbor analyses were used to show that sinkholes change from clustered to random to regular by scale, direction, and geologic unit (Gao et al., 2005a). Decision tree and cartographic tools were developed to create sinkhole probability maps for five southeastern Minnesota Counties (Gao and Alexander, 2008). A new map was created that included transition karst defined by depth to carbonate bedrock (Figure 3, Gao et al., 2008). This map was based on more detailed geologic mapping than was available for the first Karst Land Map (Figure 2).

In 2012, the database was converted to a single, relational database platform—PostgreSQL—with tabular and spatial components edited in ESRI ArcMap. Custom editing forms were written in Visual Basic and they are accessed in ArcMap sessions as ESRI add-ins. The

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**Figure 3.** Minnesota Karst Lands map, 2006 (Alexander, 2015; Gao and Alexander, 2008)
current database infrastructure allows for remote editing. Read-only versions of the data are available in GIS/spatial format for public use via web services.

**Concurrent Karst Research and Future of the KFD**

Throughout the past forty years, karst research in Minnesota has included dye tracing, cave exploration, speleothem dating, spring temperature monitoring, and geohazards investigations. Remediation investigations addressed tanker spill sites, fuel refineries, spring water quality, and structural (geotechnical) integrity. Regional investigations have taken place to evaluate geologic controls on groundwater flow and karst development. In all cases the KFD has played an important role in characterizing current and past hydrologic conditions in Minnesota karst terrain.

As the database expands and more users become acquainted with its use data standards, access, and main-

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**Figure 4.** Sinkhole distribution and bedrock geology, Minnesota and Iowa. (Gao et al., 2005a)
Maintenance issues become more critical. Combining data across state boundaries is possible without merging databases, but doing so requires clear metadata describing field names and definitions (Figure 4).

The focus of the database thus far has been on locations of sinkholes and springs. LiDAR data availability in Minnesota has greatly expanded the inventory of karst features at the land surface, particularly in wooded areas where inventories had been difficult and limited (Figure 5). LiDAR has also been used for landscape analysis, including identification of losing streams. LiDAR also provides remarkable elevation control—approximately 0.2 meter vertical resolution—critical for investigating relationships between karst terrain and hydrologic systems.

Future development plans include more focus on subsurface flow conditions. As proposed by Gao (2002), line features depicting dye trace vectors could be added, as well as polygon features showing springshed areas. Conduit information has also been proposed (Gao, 2002). How would this be recorded spatially? Outcrop occurrence is one possibility where conduit location, elevation, and stratigraphic position would be recorded as a karst feature. Hydraulically active fractures and conduits in boreholes could also be recorded by location, elevation, and stratigraphic position. Descriptive attributes of conduits could also be added, such as dimensions, or flux carrying capacity. Matrix and fracture hydraulic conductivity could also be recorded.

Figure 5. LiDAR hillshade data used to identify karst features, along with overlay of Spring Valley Caverns, Minnesota cave survey. Heavy black lines are air-filled cave passages; blue lines are underground streams; red “x”s are sinkholes, blue dots are springs; green dots are stream sinks; black “+”s are cave entrances and other surface features. (Alexander, 2015).
Dataloggers provide critical information for understanding the temporal variability of karst groundwater flow. Such high-resolution data should be incorporated as one-to-many data relationships associated with various points within karst flow systems. Measurements can include temperature, conductivity, and flow, as well as any other parameters for which probes and transducers are developed. Having unique identifiers for each karst feature allows points to be associated with other datasets, including water chemistry and isotopic data being stored elsewhere.

Ideally, the KFD describes and documents both karst terrain and karst processes. As described in this paper, the visibility of sinkholes has traditionally been the focus of karst feature databases. These points, however, do not adequately describe karst “plumbing” that is often the focus of karst research and remediation investigations.

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LEGACY DATA IN THE MINNESOTA SPRING INVENTORY

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Abstract
Past spring inventories have covered certain parts of Minnesota reasonably well; notably, the springs of the Minneapolis-St. Paul metropolitan area and the southeastern Minnesota karst. But hitherto, there has not been a systematic effort to create a uniform statewide inventory. The first step, before hunting down new springs, was to compile existing data and the most fruitful source of hydrological legacy data for the Minnesota spring inventory was the Department of Natural Resources (DNR) Fisheries files. Once entered into a GIS-capable database, these spring locations can help “seed the ground” so that when crews finally do take to the field to map more springs, they will have known examples to work from. Good baseline and time-series data should also help evaluate the impact of climate change and land use changes on Minnesota’s springs over time.

Introduction
Past spring inventories have covered certain parts of Minnesota reasonably well; notably, the springs of the Minneapolis-St. Paul metropolitan area (Brick, 1997) and those of the southeastern Minnesota karst (Gao et al., 2005). There has not been a systematic effort to create a uniform inventory for the rest of the state—a much larger, glaciated area. In 2014, state funding was provided for starting such a database. The first step was to compile existing data, which turned up in unexpected places, as explained below.

While there have been numerous other spring inventories around the country over the years, the neighboring state of Wisconsin’s has been the most relevant for comparison. The Wisconsin Conservation Department (WCD) from 1956 to 1962 mapped more than 10,000 springs in that state, the core of their present survey (Macholl, 2007). Conservation officers, familiar with their own areas, plotted the springs and recorded data such as flow rate and water temperature. Some of the points are not well defined, including features like the proverbial spring-fed lake. Indeed, the word “spring” was not even defined, nor distinguished from a seep. The Wisconsin Geological and Natural History Survey maintains an active research program involving these springs today, building on this earlier foundation (Swanson, 2013).

Setting aside for the moment differences from Wisconsin in climate and geology, and judging strictly by proportionate area, Minnesota should have about 15,000 springs, all else being equal. Even more than that, if you consider that only two-thirds of Wisconsin was covered by the WCD survey.

Minnesota’s Karst Features Database
The southeastern corner of Minnesota already has an existing spring inventory as part of its Karst Features Database (KFDB) which includes 2,648 springs (as of March 15, 2015). As described by Alexander and Tipping (2002):

“Since the early 1980s, the Minnesota Geological Survey and Department of Geology and Geophysics at the University of Minnesota have been mapping karst features and publishing various versions of their results in the form of 1:100,000 scale County Geologic Atlases. In the mid-1990s, the Minnesota Department of Natural Resources was assigned responsibility for the hydrogeology portions of the County Atlases and is now responsible for the karst mapping…. A karst feature database of southeastern Minnesota has been developed that allows sinkhole and other karst feature distributions to be displayed and analyzed across existing county boundaries in a GIS environment. The central DBMS is a relational GIS-based system interacting with three modules: spatial operation, spatial analysis, and hydrogeological modules. Data tables are stored in a Microsoft ACCESS 2000 DBMS and linked to corresponding ArcView shape files…. The karst inventory points were features such as sinkholes, springs, and stream sinks extracted from the karst feature database of southeastern Minnesota. Both inventory points and karst feature database are updated on regular basis. This research was supported with funding from the Minnesota Department of Health.”

The relational structure of the KFDB involved a total of 15 tables: a top-level karst feature index table, 12 mid-level tables to encompass the 12 entities and two
bottom-level tables for addresses and remarks (Gao et al., 2005).

**Unexpected Trove**

The KFDB notwithstanding, Minnesota’s equivalent of the WCD spring survey turned out to be elsewhere in the veritable trove of spring legacy data in the DNR Fisheries files. Springs are important for providing proper habitat for trout and other fishes. By the 1940s stream surveys were conducted for fishable streams ranging from major trout streams, like the Root River, to diminutive, unnamed urban creeks and rural ditches. Among these features there will be found data on springs, including location, estimated flow rate, and temperature, similar to the WCD survey. Duplicates of these forms are archived at the DNR’s Central Office in St. Paul, MN where they are filed by county, one stream per manila folder. Major rivers straddling multiple counties, such as the Minnesota and Mississippi rivers, have their own folders (The folder for the Minnesota River valley listed 500 springs where few had been known before). Streams are further identified by their Kittle Code, which identifies the watershed and order of tributaries. The folder also contains a stream management plan, “shocking notes” (the basis of electrofishing population assessments), creel censuses, hand-colored maps, onion-paper correspondence, yellowed newspaper clippings, and so forth. These folders are stored in more than three dozen tightly stuffed drawers of a huge mechanical KARDEX Lektriever (Figure 1). While the latest DNR stream surveys are being made available electronically the vast bulk of spring data can only be manually accessed from these hardcopies. Exact numbers are not yet tallied but the KARDEX “fishing expedition” likely netted several thousand features.

The Stream Survey is divided into many sections, evaluating the fitness of the stream as fish habitat and recording what species were found there. Section 29 covers “Tributaries and Springs.” Spring locations are given in terms of miles from the river mouth. GPS coordinates have become more common in the recent stream surveys. For comparison, the stated accuracy of the original WCD survey is one quarter section (Macholl, 2007).

However, different DNR fisheries field offices had different traditions of how to fill out the stream surveys. A striking juxtaposition is provided by neighboring Cook and Lake Counties on the North Shore of Lake Superior. Cook County has an abundance of recorded springs and Lake County, very few. Yet this turned out to be merely a reporting difference, not a real one.

Moreover, the folders will sometimes contain hand sketched maps with spring locations not mentioned in Section 29, so the entire folder for a given stream must be examined (Figure 2). Given the reported decline in spring flow with time (Surber, 1924; Moyle 1947) and given the decades over which these files have been amassed, it is possible that the springs were visible at one time but not another. Or perhaps the record reflects climate change or land use changes over the years.

There are drawbacks to the stream surveys from the perspective of a geologist. Spring classification is rudimentary in the extreme. Some of the more detail-oriented surveyors adopted a crude, three-fold scheme, dividing them into bank, bed, and cave springs. Apart from general remarks in the report itself, the geologic context of the springs is entirely lacking. The formation name, lithology, and so forth are not indicated.

The single most valuable find among the DNR stream surveys was a comprehensive 1922 map of the springs of the North Shore drafted on linen, 1.5 meters long, by Thaddeus Surber (1871-1949). Surber wrote an accompanying report for the North Shore (Surber, 1922) in which he points out some hydrologic paradoxes that will be the subject of a future paper by the present author. Surber is best known for his work as an aquatic biologist in southeastern Minnesota, where during his Root River survey of 1918 and 1920 he “traveled afoot along its many branches upwards of a thousand miles” (Surber, 1941). Mel Haugstad (1930-2013), a dedicated DNR...
fisheries manager, hiked the tributaries again adding further details.

The Lanesboro Fish Hatchery, established at Lanesboro, MN in the 1920s, is the repository of Haugstad’s legacy data. In a huge project directed by the Minnesota Pollution Control Agency (MPCA) the paper quadrangles with Haugstad’s detailed annotations are being scanned to make them more widely available (Broberg and Ignatius, 2015).

In addition to DNR Fisheries another DNR program, the Minnesota Biological Survey, has a database of seepage indicator plants—some of them rare—and lists of “rich” (i.e., groundwater-fed) fens, which harbor mud springs. Many of these are located along the “fenland arc” sweeping up the Minnesota River valley and along the edge of Glacial Lake Agassiz towards the Canadian border.

Another prolific source of legacy spring data was past publications of the Minnesota Geological Survey (MGS), especially the original county geologic reports by Winchell, Upham, and others from 1872 onwards. Here, the most surprising results included the number of cities in drier western Minnesota that were using springs as a municipal water source into the early twentieth century. Many of the standard county histories assembled in the reading room of the Minnesota History Center in St. Paul, have a geology chapter that is often just a reprint of this original MGS report.

The U.S. Geological Survey (USGS), especially its Water-Supply Papers, was consulted, and the Geographic Names Information System (GNIS) maintained by the

Figure 2. A blueprint showing spring locations in Carlton County, MN, as an example of legacy data. From Surber (1925), image processed by Holly Johnson (DNR).
USGS lists 10 named springs for Minnesota quadrangles and many more place names containing the word “spring.” Neighboring Wisconsin has 166 named springs listed in GNIS, perhaps because the mappers there chose to identify more of them by name. Once again, we find an illusory geological “fault line” along political boundaries. These sorts of boundaries bedevil attempts to create multi-state karst inventories.

Unfortunately, no simple query in GNIS can extract the much larger number of features simply labeled as springs (without a proper name) on USGS quadrangles.

The National Water Information System (NWIS), also maintained by the USGS, is a large repository of hydrological legacy data from many sources, but has limited and sporadic coverage for 43 springs in

Figure 3. Many “new” legacy spring locations are beginning to populate the map of Minnesota, whereas the KFDB is heavily focused on southeastern Minnesota. Jeff Green and Holly Johnson assisted with map preparation.
Minnesota; chiefly a cluster in the upper Minnesota River valley and a cluster of brine springs on the Grand Portage Indian Reservation, apparently in support of various USGS investigations. The U.S. Forest Service, especially in the Boundary Waters Canoe Area, has also compiled spring locations.

Combining these sources, the big white space on the map outside of southeastern Minnesota is becoming populated with legacy springs (Figure 3).

Conclusions
The most fruitful source of hydrological legacy data for the Minnesota spring inventory was the DNR Fisheries files. Before hunting for unmapped springs, it’s important to utilize such data. Once entered into a GIS-capable database, these locations can help “seed the ground” so that when crews finally do take to the field to map more springs they will have known examples to work from. Good baseline and time-series data should also help evaluate the impact of climate change and land use changes on Minnesota’s springs over time.

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DEVELOPMENT OF CAVITY PROBABILITY MAP FOR ABU DHABI MUNICIPALITY USING GIS AND DECISION TREE MODELING

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Abstract

Cavity collapse and settlement due to the presence of shallow solution cavities cause significant geotechnical and other engineering problems in certain areas within the Abu Dhabi City Municipality (ADM). A cavity probability map helps to identify regions that are more susceptible to the formation of cavities by identifying and analyzing influential factors contributing to its formation. Information relating to cavities was cataloged and reviewed based on available data from the Geotechnical Information Management System (GIMS), which is a consolidated geotechnical database developed by the ADM. Geological and geotechnical subsurface conditions are obtained from previous site investigation campaigns performed in the ADM region. All geotechnical, geological, and cavity related datasets are stored in a GIS geodatabase system. Based on detailed literature review, primary factors influencing formations of cavities are identified: presence of soluble bedrock, depth to Gachsaran Formation, cavity density, cavity thickness and distance to nearest neighbor. A decision-tree model based on cavity distribution was developed for cavity hazard assessment. The primary controls on cavity development are lithostratigraphic position or bedrock geology and depth to the soluble Gachsaran Formation. Most cavities tend to form in highly concentrated zones. Implementation of the decision-tree model in ArcGIS resulted in a cavity probability map. This cavity probability map is mainly based on existing borehole data. Areas not fully mapped by boreholes must be re-evaluated for cavity risk when new borehole data is available. Low Probability, Low to Moderate Probability, Moderate to High Probability, High Probability, and Very High Probability areas were delineated in the probability map.

Introduction

The Abu Dhabi Municipality (ADM) area has undergone rapid infrastructure development and urbanization in the last two decades (UPC, 2007). Almost the entire urbanized Abu Dhabi City including many of the coastal islands is reclaimed land covered by backfill material. The backfill is found mostly in places in an uncontrolled way over pre-existing, coastal barrier and supratidal sabkha sediments (Price et. al, 2012). During the process of infrastructure development and extension of Abu Dhabi
City, significant issues relating to the presence of subsurface problems including cavities and collapse features have been encountered (Tose and Taleb, 2000). Cavity collapse has presented a significant geohazard across parts of the Municipality (Mouchel, 2012). The Gachsaran Formation, which is composed of interlayered mudstone and gypsum, underlies all of the ADM area and is known to be vulnerable to cavity formation in the area. The mudstone and gypsum beds within the upper part of the Gachsaran Formation are prone to dissolution; numerous sinkholes have been reported, particularly in the zone between Abu Dhabi International Airport and Mafraq (Farrant et al., 2012; Mouchel, 2012).

In recent years, Geographic Information System (GIS) are used for manipulation and management of spatial data. There have been studies that apply GIS as a tool to identify or highlight regions that are more prone to cavity formation (Gao et. al, 2007; Yilmaz, 2007; Dai et. al, 2008; Cooper, 2007; Amin and Bankher, 1996; Hu et. al, 2001). The main objective of this study is to access relative probabilities of cavity occurrences in the ADM using GIS tools.

**Geological and Geographic Background**

The study area in ADM is approximately 11,000 km². It includes the mainland urban area of Abu Dhabi in addition to the coastal islands. The coastal area is relatively flat. Topographic elevation rises to approximately 35 m above sea level to the east and southeast across an arculate ‘escarpment’ trending from Mafraq in the south to Al Shahama in the north (Price et al., 2012).

The near surface geology of coastal Abu Dhabi Islands consists of Quaternary marine, aeolian, sabkha, and fluvioglacial deposits overlying variably cemented Pleistocene sands (Macklin et. al, 2012). Most solution cavities occur further inland in regions such as Shakbout City, Zayed City, and regions surrounding the Abu Dhabi International Airport as shown in Figure 1. Inland geology of the ADM consists of Aeolian sand, active sabkha sequences, dune-bedded sandstone, marine developed carbonate mudstone and sandstone, and evaporite deposits (Tose and Taleb, 2000). The ADM is underlain by the Gachsaran Formation that is part of the Neogene system (Alsharhan and Narin, 1997). The Gachsaran Formation is a thick evaporitic basinal succession that was deposited in a shallow marine/brackish setting with input from a nearby land source indicated by plant material. It is well known from offshore oil wells, but is only poorly exposed onshore in the Abu Dhabi Area where it is recorded in numerous temporary excavations and boreholes that have penetrated up to 100 m of interbedded mudstone and gypsum (Farrant et al., 2012a). Small exposures occur around Mafraq, Shakbout City, Shahama, Al Bahya, and along the foot of the Dam Formation escarpment around the Al Dhafra Air Base at Al Maqta (Farrant et al., 2012a, b).

Evaluation of the lithological sections indicated that ground excavations had periodically intercepted open voids in the mudstone and gypsum, and the loss of fluid circulation was commonly reported on drilling logs. Borehole data indicated that most of these cavities occur close to the top of rock, often at the interface between the overlying superficial deposits or sandstone and the underlying mudstone and gypsum. The data also showed that the cavities are most prevalent where the Gachsaran is closest to the surface. This formation of cavities is believed to be formed by groundwater movement along the interface of the mudstone and gypsum layers forming cavities that are more vulnerable to collapse in the vicinity of the top of bedrock.

The source for location and information of cavities for the study is mainly from a borehole database maintained by the Municipality of Abu Dhabi. The database consists of 21,257 geotechnical borings (Geotechnica, 2014). This borehole database is called Geotechnical Information Management System (GIMS). The GIMS for Abu Dhabi City supports a consolidated geotechnical database in accordance with internationally accepted standards. A preliminary geodatabase was developed to
manage spatial data acquired during the data collection process of this study and 1201 cavities were identified and extracted from the GIMS database.

GIS Geodatabase

In the last decade, GIS and database management systems have been widely developed to manage and analyze spatial data relating to geologic, geotechnical and karstic features (Cooper et al. 2007; Lei et al. 2001, Gao et al. 2005). Spatial data manipulation in GIS environments is a key function of any GIS application (Demers, 1997). There are numerous advantages to manage spatial information and GIS data layers in a geodatabase environment as it allows for coordinated relationships between feature classes, which enable the creation of domains thereby reducing errors during data entry (Ormsby and Burke, 2004). A geodatabase enables storage in a single file or folder and is more efficient for storage of large datasets (FLNRO, 2013). The geodatabase supports a model of topologically integrated feature classes, similar to the coverage model. It also extends the coverage model with support for complex networks, relationships among feature classes, and other object-oriented features (MacDonald et al., 2001).

For this study an ESRI geodatabase called Geohazard Information Management System (GHIMS) was developed to store, manage, and analyze data relating to karstic features, such as cavities, surface subsidence, and presence of soluble bedrock formations, in addition to other information contributing to local geohazards. All data storage and management were performed in ArcCatalog and all data manipulations were performed in ArcMap. The GHIMS geodatabase is a tool developed to analyze regional geohazards within the ADM. The geodatabase contains a specific set of feature datasets, feature classes, raster catalogs, and raster classes together with feature attributes, subtypes, and domains; suitable for a variety of geologic, hydrogeologic, and risk assessment maps. In addition to basic geology (lithology, cavity location, etc.), the geodatabase includes damaged buildings and roads survey data, susceptibility of cavity to collapse, and geohazard risk assessment. This paper documents all layers relevant to the karstic geohazards in the region, solution cavities under the surface (Tose and Taleb, 2000). Table 1 shows the major components of the GHIMS geodatabase.

Discussing all datasets stored within the GHIMS geodatabase is not within the scope of this paper. Only layers that store information relevant to the solution cavities, which serve as input data is discussed. The KARST_CAVITY (KC) feature dataset consists of six feature classes as shown in Figure 2. There are four point feature classes and two polygon feature classes. Cavity collapsibility (KC_CVT_CLIPSB) is a point feature class that shows the distribution of stable or unstable cavities. The stability of cavities depends on a series of stability charts produced from running simulations on a finite difference model using a software called FLAC 3D. This analysis is outside the scope of this paper. Halite_Bhs (KC_HALITE_BH) is a point feature class that provides

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Table 1. Major components of the GHIMS geodatabase are listed here. The GHIMS geodatabase is suitable for storing and managing a variety of geologic, hydrogeologic, and risk assessment related information. "*" indicates data layers that are relevant to this paper.
the locations of boreholes that have halite or rock-salt listed in the geology description of the boring logs. The halite or salt layer is an evaporite crust that is susceptible to dissolution. Old_Risk_Map (KC_OLD_RSK_MP) is a polygon feature class that represents the existing cavity risk map developed based on previous studies (Tose and Taleb, 2000). This feature class has been used only as a reference and is not used in the development of the cavity probability map. Salt_Layer (KC_SLT_LR) is a polygon feature class that provides the possible spatial extent of sub-surface halite zones. It is derived from the existing cavity risk map and from querying geologic descriptions provided in the GIMS borehole database. Void_Depths (KC_VD_DPTH) is a point feature class, which stores information relating to the presence of cavities or voids, and the depth to these cavities or voids based on data from borehole log descriptions from the GIMS database. Water_Loss (KC_WTR_LOSS) is a point feature class, which stores information relating to the event of water or drilling fluid loss at the time of drilling as noted from borehole log descriptions from the GIMS database. This layer could indicate probable locations of subsurface voids or cavities. Since it is not a confirmatory source for presence of cavities, this layer is also used for reference only.

**Parameters Contributing to Cavity Formation**

Karstic cavities are geologic features that result from water erosion in soluble rocks over time due to seasonal groundwater variation and/or groundwater flow and the associated seepage forces. The developed void system results in randomly shaped cavities that vary widely in size, geometry, and location within the soluble rock. In Abu Dhabi area, cavities were detected as sizable caves encountered during construction of infrastructure and during drilling from the loss of fluid circulation or string drop as documented in boring logs. The formation and collapse of the karstic cavities may be triggered by changes in the groundwater regime, changes in surface drainage, and construction work or urban development. In the Abu Dhabi area, irrigation inland and construction related dewatering within the urban area is likely to be one of the key triggers for sinkhole development via enhanced dissolution and flushing out of existing sediment filled cavities (Farrant et al., 2012).

A total of 1201 cavities are identified by querying the GIMS borehole database using SQL. The Shakhbout City areas contained 67% (i.e., 806 out of the 1,201 inventoried cavities). Other areas where significant number of cavities occurred include the southeastern Zayed City, the Abu Dhabi International Airport and the Al Falah areas. A small number of cavities were sparsely distributed in other areas. However, some boreholes indicate the presence or multiple cavities at different depths. In such cases the cavity closest to the surface is used for the cavity risk assessment. Eliminating multiple cavities in the same boreholes, the total dropped to 729 cavities nearest to the surface. Bedrock solubility, depth to Gachsaran Formation, cavity density, cavity size, and point pattern analysis were used as contributing factors in the formation of cavities.

**Depth to Gachsaran Formation**

The Gachsaran Formation, which is composed of interlayered mudstone and gypsum, underlies all of the ADM and is known to be vulnerable to cavity formation in the area. The mudstone and gypsum beds within the upper part of the Gachsaran Formation are prone to dissolution; numerous sinkholes have been reported, particularly in the zone between Abu Dhabi International Airport and Mafraq (Farrant et al., 2012; Mouchel, 2012). Evaluation of the lithologic sections indicates that ground excavations have periodically intercepted open voids in the mudstone and gypsum, and the loss of fluid circulation...
is commonly reported on drilling logs. Borehole data indicate that most of these cavities occur close to the top of bedrock, often at the interface between the overlying superficial deposits or sandstone and the underlying mudstone and gypsum. The data also shows that the cavities are most prevalent where the Gachsaran is closest to the surface. This formation of cavities is believed to be formed by groundwater movement along the interface of the mudstone and gypsum layers forming cavities that are more vulnerable to collapse in the vicinity of the top of bedrock. In other areas such as Abu Dhabi Island and Al Falah, cavities have been encountered within the stratigraphically higher sand and sandstone layers, as well as at the interface with the Gachsaran.

Figure 3 shows a histogram of the distribution of cavities in relation to the depth to Gachsaran formation at the cavity location. It is evident that the closer to the surface of the Gachsaran Formation the more likely the formation of cavities. Figure 4 shows the extent and depth to the Gachsaran Formation.

**Cavity Density**

Cavity density provides the number of cavities present per square kilometer. The cavity density is calculated using the Point Density tool under the Spatial Analyst toolbar in ArcMap application. The Point Density tool calculates the density of point features around each output raster cell. Conceptually, a neighborhood is defined around each raster cell center, and the number of points that fall within the neighborhood are added together and divided by the area of the neighborhood (Silverman, 1986). Figure 5 shows the cavity density output calculated in ArcMap.

**Cavity Size**

In the field, cavities tend to propagate in vertical and lateral directions, but since the source of cavity data is discreet points, cavities are assumed as two dimensional features. Cavity size is estimated using the thickness of voids based on boring log data. Cavity size varies from 0.1 m in thickness to 3 m in thickness. Cavities with thickness greater than 3 m were also observed although these were few in number compared to the total dataset. The largest cavity encountered is around 17.5 m thick. Majority of the cavities are 0.1 to 1 m thick. Cavities

![Figure 3](image1.png)

**Figure 3.** Histogram showing the distribution of cavities in relation to the depth to Gachsaran formation.

![Figure 4](image2.png)

**Figure 4.** The areal extent and vertical depth of the Gachsaran Formation below ground surface level.

![Figure 5](image3.png)

**Figure 5.** Cavity density raster output created using location of cavities as input source in ArcMap environment using the point density tool.
smaller than 1 m were found to be generally stable, as supported by the numerical analyses results. Figure 6 shows the histogram for cavity distribution with respect to cavity size and Figure 7 shows the areal distribution of cavities based on cavity size.

**Point Pattern Analysis**

Pattern analysis is the study of the spatial arrangement of point features in two-dimensional space. A pattern analysis usually demonstrates if a distribution pattern is random, dispersed, or clustered (Gao et al., 2005). In addition, a distribution pattern containing clusters of high or low values can also be identified by pattern analysis. Distances to the first through the 9th nearest neighbors were conducted for cavities in different lithological materials and geographical clusters. Figure 8 demonstrates a histogram of the distance to the nearest cavity for all cavities. The median distance to the first through the 9th nearest cavity is linearly increasing within the Gachsaran Formation as shown in Figure 9.

The overall Distance to Nearest Neighbor (DNN) distribution of all cavities does not follow Poisson, Normal,

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**Figure 6.** Cavity size variation represented in histogram format. Majority of the cavities fall between 0.1 m to 1 m in thickness.

**Figure 7.** The spatial distribution of cavities based on cavity size.

**Figure 8.** Histogram and cavity distribution with respect to distance to nearest cavity. The distribution of cavities with distance to nearest cavity greater than 160 m follows a normal distribution.

**Figure 9.** The median distance to the first through the 9th nearest cavity in the Gachsaran formation.
or Log-Normal distributions. However, the distribution of the DNN for all cavities more or less follows normal distribution once DNN is greater than 160m.

**Decision Tree Model and Implementation**

One of the important advantages of geoinformatics techniques is that it can be used to extrapolate the occurrence of local events over a wider territory using statistical methods and predict the possibility of occurrence of these local events over an expanded territory. Geoinformatics technology or GIS applications can be used to develop multi-parametric models that can make predictions based on a set of training examples. Several studies have developed multi-parameter models based on multi-scenario considerations to make predictions on the occurrence of sinkholes, cavities and other geohazards (Koutepov et al. 2008; Gao et al. 2007; Yilmaz, 2007; Cooper, 2007; Tolmachev, 2003; Ragozin and Yolkin, 2004, Kaufmann, 2008).

The purpose of multi-parametric model in assessing cavity collapse hazards is to divide the study area into sub-areas of different hazard or probability levels. To this end, spatial data mining aids in discovering spatial patterns among various contributing parameters (Shekhar and Chawla, 2002). A study in Great Britain uses a detailed karst database and assigns severity of dissolution hazards by assessing local bedrock and superficial geology and sub dividing regions into high, moderate, and low risk zones based on a ranking or scoring system (Cooper, 2007). A similar scoring system was developed in Missouri by assigning scores to sub classes of multiple parameters such as depth to water table, bedrock characteristics, proximity of nearest sinkhole, and distance to nearest structure from existing sinkholes (Kaufmann, 2008).

A more rigorous multi-parameter model is the frequency ratio model. Parameter maps that are used in the collapse susceptibility analyses are divided into four groups such as: geological and hydrological, topographical, land use, and vegetation cover. Each of these parameters is further subdivided into sub-classes and cavity collapse hazard is calculated as a function of the frequency of cavities occurring each of the subclasses (Yilmaz, 2007).

Another common multi-parametric modelling approach is the probabilistic method (Tolmachev, 2003; Ragozin and Yolkin, 2004). In the probabilistic approach, sinkhole or cavity collapse risk is expressed in terms of the probability \( P_s \) of formation of sinkholes in a specified period (for example, during the service life of a building) on the studied territory, which may cause impermissible deformation of structures, or in terms of the probability \( P \) that there will be no such sinkholes (reliability), i.e., \( P = 1 - P_s \).

Decision tree models are one of the most widely used techniques for inductive inference (Mitchell, 1997; Winston, 1992). A decision tree model uses a top-down approach and consists of multiple nodes (Gao et al., 2007; Hu et al., 2001). Each node indicates a test condition followed by the next node all the way to the last node (Tan et al., 2005). In this study the decision tree model is implemented to develop a cavity probability map given the study area and extent. The decision tree method is more suitable for integrated and regional scale assessments of complicated phenomena such as occurrence of cavities (Hu et al., 2001). Based on the contributing parameters listed in this study a decision tree model was developed as shown in Figure 10. The primary controls on cavity development were lithostratigraphic position or bedrock geology and depth to the soluble Gachsaran Formation. The majority of the cavity population tends to form in highly concentrated zones. Neighborhood effect plays a very important role in cavity distribution and formation.

Figure 11 represents the various spatial data manipulations performed in ArcMap to create the input layers for the final cavity probability calculations. The existing bedrock geology layer was reclassified into soluble and insoluble bedrock units based on their susceptibility to dissolution. The depth to Gachsaran Formation raster layer was queried from the GHIMS geodatabase and reclassified into to two units: pixels representing values of depth to Gachsaran Formation less than 30 m and pixels representing values of depth to Gachsaran Formation greater than or equal to 30 m.

Similarly cavity density raster layer and cavity thickness layers were reclassified in two value rasters as shown in Figure 11. To create the input layer for distance to nearest cavity the mean and standard deviation of DNN were used to define boundaries (Gao and Alexander, 2003). Using the Buffer tool in ArcMap environment raster layers indicating boundaries within 210 m, 400 m and 600 m were created and were combined using the Union tool in ArcMap. Using Model Builder tool in ArcMap, the decision tree model was implemented using the input layers shown in Figure 11. A pictorial representation of the model built to calculate the cavity probability map is shown in Figure 12.

**Results**

Implementation of the decision tree in ArcGIS resulted in a cavity probability map. Figure 13 shows the cavity probability map developed for the ADM area. The cavity
Figure 10. Decision tree model created to assign cavity risk probability for the ADM region. The decision tree includes characteristics of bedrock geology, depth to the Gachsaran Formation, cavity density, cavity size, and distances to the nearest cavities in the ADM area.

Figure 11. Cartographic flow chart representing the implementation of the decision tree model in ArcMap environment. This flowchart represents the process to create the input layers for the final cavity probability calculation.

The probability map divides the study area into regions of low probability, low to moderate probability, moderate to high probability, high probability, and very high probability. The descriptions of these probability areas are as follows.

**LOW PROBABILITY**
Areas underlain by the soluble Gachsaran Formation and the depth to the Gachsaran Formation is equal to or greater than 30m are shown on the map as having low probability for cavity development.

**LOW TO MODERATE PROBABILITY**
Areas underlain by the soluble Gachsaran Formation and the depth to the Gachsaran Formation is less than 30m are shown on the map as having low to moderate probability for cavity development. The cavity density is less than one cavity per square kilometer. The expected future cavity development is generally low in these areas, but is moderate where small cavity clusters have developed.

**MODERATE TO HIGH PROBABILITY**
Areas in which cavities are a routine part of the subsurface and the minimum cavity density is 1 cavity per square kilometer. Higher probability cavity clusters are
Figure 12. Pictorial representation of the ArcMap Model Builder file used to calculate the final cavity probability map.

Figure 13. The cavity probability map developed in ArcMap environment based on decision tree modeling technique.
usually contained with the moderate to high probability. The minimum distance to the nearest cavity is 400 m for smaller cavities (less than 3m in thickness) and 600m for larger cavities (greater than and equal to 3m).

**HIGH PROBABILITY**
Areas in which cavities are a common part of the subsurface and the minimum cavity density is 1 cavity per square kilometer. The minimum distance to the nearest cavity is 210 m for smaller cavities (less than 3m in thickness) and 400m for larger cavities (greater than and equal to 3m). New cavities are expected to form in these areas.

**VERY HIGH PROBABILITY**
Areas in which cavities are dominant features of the subsurface and the minimum cavity density is 1 cavity per square kilometer. The minimum distance to the nearest cavity is 210 m and at least a large cavity (greater than and equal to 3m) occurs within these areas. Four of these clusters containing extremely large cavities (greater than and equal to 10m) would be very susceptible for future cavity development.

**Discussion and Conclusions**
The cavity probability map, when compared with earlier, elementary versions of zone level cavity risk assessment studies, produces a more structured and objective approach towards analyzing patterns in the spatial distribution of cavities (Tose and Taleb, 2000). However, other influential parameters controlling formation of cavities such as groundwater chemistry and fluctuation, land use and topography, as well as anthropogenic changes to landscape and groundwater were not considered in the study due to the lack of data availability. This cavity probability map is mainly based on existing borehole data. Areas not fully mapped by boreholes need to be re-evaluated for cavity risk once new borehole data are available. Also, in this study cavities are assumed as discontinuous 2D features, while in reality cavities tend to develop and propagate in vertical and lateral directions.

**References**
Geotechnical Risk Map, Presidential Affairs 44 Plots
Tennant EW. 2007. A sample geodatabase structure for managing archaeological data and resources with ArcGIS. Technical Briefs in Historical Archaeology 2: 12-23
EVALUATION OF CAVEITY DISTRIBUTION USING POINT-PATTERN ANALYSIS

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Abstract  
The presence of solution cavities of different sizes poses major engineering problems in some areas of Abu Dhabi City Municipality (ADM) underlain by soluble rocks such as gypsum, calcarenite, or mudstone. This is especially critical if they are located at a relatively shallow level and are likely to cause settlement or sudden soil collapses. The Gachsaran Formation, which is composed of interlayered mudstone and gypsum, underlies all of the ADM and is known to be vulnerable to cavity formation in the area. Information associated with cavities was cataloged and reviewed based on available data from an existing geotechnical borehole database maintained by the ADM. Cavity data obtained from borehole information were analyzed to examine cavity distributions based on the following factors: lithology, geographic clusters, cavity density, cavity size, depth to cavity, and depth to bedrock. All cavities were grouped into geographic clusters and lithological clusters for point-pattern analysis. Most cavities (87 percent) occur in mudstone or gypsum, or at an interface between these two rock types, which compose part of the Gachsaran Formation. Geographically the majority of cavities occurred in the Shakhbout City area hence pattern analysis including average nearest neighbor analysis, Moran’s I for measuring spatial autocorrelation, and G-statistics for measuring high/low clustering were conducted in this area using spatial statistics tools in ArcGIS. Average nearest neighbor analysis and Moran’s- I show that cavities are strongly clustered in this area with a high confidence level (>99 percent). General G-statistics identified a high clustering (hot spot) of cavities with relatively high values of depth to cavity, depth to bedrock, and number of cavities per borehole. No highly clustered large cavities were detected by the General G-statistics. Additionally, distances to the first through the ninth nearest neighbors were determined for cavities in different lithological materials and geographical clusters. Outcome of these spatial correlations and statistical analysis can be used to conduct risk assessment and the probability of occurrences of cavities in the future.

Introduction  
Presence of solution cavities of different sizes poses major engineering problems in some areas of Abu Dhabi City Municipality (ADM) underlain by soluble rocks such as gypsum, calcarenite, or mudstone. This is especially critical if they are located at a relatively shal-
Evaluation of the lithologic sections indicated that excavations periodically intercepted open voids in the mudstone and gypsum, and the loss of fluid circulation was commonly reported on drilling logs. Borehole data indicated that most of these cavities occur close to the top of the bedrock often at the interface between the overlying superficial deposits or sandstone and the underlying mudstone and gypsum. This formation of cavities is believed to be formed by groundwater movement along the interface of the mudstone and gypsum layers forming cavities that are more vulnerable to collapse in the vicinity of the top of rock (Farrant et al., 2012a).

Geohazard risk maps are currently available only for the Shakhbout City and Zayed City areas within the ADM. Most notably, Tose and Taleb (2000) developed a ground condition “risk” classification map for the former Shakhbout City and Zayed City areas. Although ostensibly designed to identify generally adverse subsurface conditions, Tose and Taleb’s classification scheme correlated risk with shallow, less than 20 m or 66 ft below ground surface [bgs], cavity distributions (heights) and so-called “broken subsurface strata” extents, as inferred from extensive geotechnical boring and geophysical survey data.

A similar, relatively simplified, cavity-based geotechnical risk classification map was developed by local practitioners for a discontinuous 44-plot area located within Shakhbout City (Spektra Jeotek, 2011; 2012). On this map risk distribution was based solely on cavity (void) density and depth below ground surface. An overall low-risk classification was ascribed to individual plots in which voids were determined to be located more than 16 m or 52 ft below the ground surface. In contrast, high risk was ascribed to plots in which “ground flaws” (including voids and water-loss instances) were largely confined to depths between 3 m (10 ft) and 10 m (33 ft); no voids were located shallower than 3 m (10 ft). Hazard within any class (low, medium, or high) was in turn determined by inferred cavity densities. Specifically, very low-risk conditions were assigned to individual plots with no more than three deep cavities. Moderately low-risk conditions were similarly assigned to plots with less than 10 (more than 3) deep cavities. Slightly higher but still generally low-risk conditions were assigned to plots containing abundant (more than 10) but deep cavities.

These existing classification schemes did not consider other significant cavity stability factors; such as cavity cover thickness, overburden lithology and mechanical characteristics, hydrogeologic conditions, and groundwater geochemistry due to lack of data availability.

**Study Area**

Abu Dhabi is located in the stable cratonic region of the Arabian Plate. The study area covers an area of 11,000 square kilometers (4,250 square miles). It includes the mainland urban area of Abu Dhabi in addition to the coastal islands. Based on data availability the extent of study area was chosen as shown in Figure 1. The coastal area is relatively flat. Topographic elevation rises to approximately 35 m (115 ft) above sea level to the east and southeast across an arcuate ‘escarpment’ trending from Mafraq in the south to Al Shahama in the north (Price et al., 2012). Almost the entire urbanized Abu Dhabi City including many of the coastal islands is reclaimed land covered by backfill material. The backfill is found mostly in places in an uncontrolled way over pre-existing, coastal barrier and supratidal sabkha sediments.

The sedimentary sequence underlying the region consists of a relatively flat-lying assemblage of Paleozoic through Cenozoic carbonates and evaporites with interbedded clastic horizons to a thickness of approximately 8,000 m or 26,250 ft (Al-Jallal and Alsharhan, 2005). Above this are extensive Holocene aeolian deposits forming the sand dunes of the Rub’ al Khali, as well as localized sabkha sequences. A sabkha is defined as a flat area prone to periodic inundation and evaporate depositions, dominated by carbonates or sulphates (Al-Farraj, 2005). They are commonly formed in arid shallow-shelf environments, and are formed in response to two environmental conditions: deflation of sediment surfaces and sediment accumulation in a lagoon, or by a combination of both processes (Evans, 1970). Most of the solution cavities occur in the Gachsaran Formation part of the Neogene system (Alsharhan and Narin, 1997). The Gachsaran Formation is a thick evaporitic basinal succession that was deposited in a shallow marine/brackish setting with input from a nearby land source indicated by plant mat-
It is well known from offshore oil wells, but is only poorly exposed onshore in the Abu Dhabi Area where it is recorded in numerous temporary excavations and boreholes that have penetrated up to 100 m (328 ft) of interbedded mudstone and gypsum (Farrant et al., 2012a). The Gachsaran Formation is covered by the Abu Dhabi Formation along the coast, and by younger Miocene and Quaternary sediments inland. Small exposures occur around Mafraq, Shakbout City, Shahama, Al Bahya, and along the foot of the Dam Formation escarpment around the Al Dhafra Air Base at Al Maqatrah (Farrant et al., 2012a, b). Figure 2 shows the extent and depth to Gachsaran Formation within the study area.

Abu Dhabi Cavity Characteristics and Distribution

The ADM maintains a borehole database consisting of around 21,000 geotechnical borings. This borehole database is called Geotechnical Information Management System (GIMS). The GIMS for Abu Dhabi City supports a consolidated geotechnical database in accordance with internationally accepted standards. For this study, the GIMS borehole dataset was queried for string drops (also recorded as ‘free fall of drilling rod’ in the field logs) or loss of water, which are indicators of voids or cavities within a given boring. Since these are only indicators of the presence of subsurface cavities and voids for the purpose of this study it is assumed that these indicators are in fact subsurface cavities and voids adopting a conservative approach. A detailed geophysical and ground exploration investigation should be performed for confirmatory and verification purposes. A preliminary geodatabase was developed to manage spatial data acquired during the data collection process of this study.

A total of 1201 cavities are identified by querying “string drop”, “free fall” or “loss of water” the GIMS borehole data. However, some boreholes may encounter multiple string drop, free fall, or loss of water features. The top most cavity for each borehole is used for the cavity hazard assessment for this phase of the GGHIP. Therefore, a total of 729 cavities nearest to the surface for each borehole were selected for analysis. Overburden thick-
Fractures of the soluble rocks through suffosion or piping processes. Therefore, the depth to the Gachsaran Formation would be an important criterion for cavity hazard assessment. Other reasons for occurrence of voids in insoluble materials could be due to drilling activities due to weak material collapsing.

Figure 3. Spatial distribution of cavities in the Abu Dhabi Municipality.

Figure 4. Cavity distribution in different lithological materials. Chart showing the occurrence of cavities in different types of lithologies prevalent in the ADM.

Table 1. Cavity distribution in different lithological materials.

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<tr>
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<tr>
<td>Mudstone</td>
<td>209</td>
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<tr>
<td>Siltstone</td>
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<td>Sandstone</td>
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<tr>
<td>Calcarenite</td>
<td>1</td>
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<td>Soil</td>
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</tbody>
</table>
*| Gypsum - Mudstone Interfaces | 162 |
| Other Interfaces       | 20                 |

* Caution occurring at interfaces were not assigned to one particular lithology. Instead, they were classified into Gypsum-Mudstone Interfaces and Other Interfaces. Gypsum-Mudstone Interface can be either Gypsum occurring over Mudstone or Mudstone occurring over Gypsum, with cavities in between. Other Interfaces includes combinations of all lithological materials except the Gypsum-Mudstone Combination with cavities in between.

The occurrence of cavities in different types of lithological materials in the ADM area is shown in Table 1. Figure 4 shows a chart representation of Table 1. Most cavities (87%) occur in Mudstone, Gypsum, or at an interface between these two rock types, which compose part of the Gachsaran Formation (Ga). The Gachsaran Formation is a thick evaporitic basin succession consisting of carbonates and evaporites, with marls and thin limestone (Bahroudi and Koyi, 2004). It does not form natural outcrops at surface (Farrant, A.R., et al., 2012). However, the dissolution of carbonate and evaporites within this formation causes subsurface voids formed in the ADM area. Even though some voids occurred in non-soluble rocks such as siltstone and sandstone, they were most likely associated with the dissolution of Gachsaran Formation underneath. Since the Gachsaran Formation is so extensive in the ADM area soil and sediment above the Gachsaran Formation can migrate down into voids and fractures of the soluble rocks through suffosion or piping processes. Therefore, the depth to the Gachsaran Formation would be an important criterion for cavity hazard assessment. Other reasons for occurrence of voids in insoluble materials could be due to drilling activities due to weak material collapsing.
Mudstone has the second highest distribution of cavities among the other lithological material in the ADM. Even though gypsum is known to be more soluble than mudstone, mudstone layers generally have low compressive strength compared to gypsum layers. The mudstone in ADM is characterized as highly weathered with intact compressive strength as low as 100 kPa which is less than 2% of the lowest intact compressive strength of the gypsum core samples tested. Three factors can be attributed to weathering and cavity formation in the mudstone: repeated cycles of wetting-drying; the highly weathered nature of the encountered mudstone, given the fact that it is made of fine-grained sedimentary rock of lightly cemented clay and silt, will enhance fines washout from rain infiltration and groundwater flow; and lastly dissolution-crystallization of relatively soluble minerals of gypsum interbedded within the mudstone (Canton et al., 2001).

**Point Pattern Analysis**

Many attempts have been made in the past to study patterns among point data in various natural systems. Clark and Evans (1954) and Thompson (1956) developed a nearest-neighbor analysis (NNA) method which has been used in many research areas. Another study (Drake and Ford, 1972) analyzed the patterns among two generations of sinkholes in Mendip, England by comparing the mean distances of the first to the twelfth nearest neighbors between the two generations of sinkholes.

A comprehensive investigation of cavity distribution is critical to conduct hazard assessment in the ADM area. Point pattern analysis is the first step to examine if the cavities are clustered or randomly distributed. Depth to bedrock, depth to cavities, cavity density, cavity size or thickness, and distributions of cavities in different geographic and lithological clusters help to characterize locations where cavities would likely occur. Pattern analysis is the study of the spatial arrangement of point features in two-dimensional space (Gao, 2002). ArcMap provides tools to analyze point pattern distribution that can be used to determine clustering or level of dispersion among the different data points based on the size of the study area. The Average Nearest Neighbor tool measures the distance between each feature centroid and its nearest neighbor’s centroid location. It then calculates the average of all these nearest neighbor distances. If the average distance is less than the average for a hypothetical random distribution, the distribution of the features being analyzed is considered clustered (Ebdon, 1985). The Spatial Autocorrelation (Global Moran’s 1) tool measures spatial autocorrelation based on both feature locations and feature values simultaneously. Given a set of features and an associated attribute, it evaluates whether the pattern expressed is clustered, dispersed, or random (Getis and Ord 1992; Griffith, 1987). The High/Low Clustering (Getis-Ord General G) tool measures the concentration of high or low values for a given study area. The High/Low Clustering tool is most appropriate when there is a fairly even distribution of values and unexpected spatial spikes of high values need to be identified (Mitchell, 2005).

**Results**

A pattern analysis usually demonstrates if a distribution pattern is random, dispersed, or clustered. In addition, a distribution pattern containing clusters of high or low values can also be identified by pattern analysis. This section discusses the results of the pattern analysis performed on the cavity dataset.

**Pattern Analysis in the Shakbout City Area**

Since the majority of cavities occurred in the Shakbout Area pattern analysis, including average nearest neighbor analysis; Moran’s I for measuring spatial autocorrelation; and G-statistics for measuring high/low clustering, were conducted in this area using spatial statistics tools in ArcGIS. Figures 5 through 11 illustrate results of point pattern analysis of cavities in the Shakbout Area. Average nearest neighbor analysis (Figure 5) and Moran’s I (Figure 6) show that cavities are strongly clustered in this area with high confidence level (>99%). General G statistics identified high clustering (hot spot) of cavities with relatively high values of depth to cavity, depth to
bedrock, and number of cavities per borehole (Figures 8 and 9). No highly clustered large cavities were detected by the General G statistics (Figure 7).

**Pattern Analysis for Factors Influencing Formation of Cavities**

Depth to Gachsaran Formation, depth to bedrock, depth to cavity, and cavity size distributions were conducted in three geographic clusters including the Shakbout City and southeastern Capital District, the Abu Dhabi International Airport, and the Al Falah areas. Most cavities occurred in areas surrounding the Shakbout City area, including the southeastern Capital District area, and these areas represent typical geological settings for the cavity hazard assessment. Therefore, results of depth to bedrock, depth to cavity, and cavity size distributions for cavities within the Shakbout City and the southeastern Capital District area are discussed in this paper. Cavity

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**Figure 6.** Shakbout Area – Moran’s I with cavity size indicates a clustering of cavities with similar size.

**Figure 7.** Shakbout Area – General G with cavity size indicates large cavities are not clustered.

**Figure 8.** Shakbout Area – General G with depth to cavity indicates cavities occurring at similar depths are clustered.

**Figure 9.** Shakbout Area – General G with depth to bedrock indicates high clustering of cavities at certain depths to bedrock.
size is a two-dimensional attribute represented by the thickness of each cavity or the distance between the top and bottom elevations of each cavity. Depth to Gachsaran Formation, depth to bedrock, and depth to cavity all follow normal distributions (Figure 10). Cavity size distribution (Figure 11) is more random similar to the Poisson distribution, which is consistent to results of the General G statistics (Figure 7).

**Nearest Neighbor Analysis**

Distances to the first through the ninth nearest neighbors were conducted for cavities in different lithological materials and geographical clusters. Figure 12 demonstrates a histogram of the distance to the nearest cavity within the Gachsaran Formation. The median distance to the first through the nineth nearest cavity is linearly increasing within the Gachsaran Formation (Figure 12).

For nearest neighbor analysis of the entire ADM area, some cavities may have a nearest neighbor that lies outside of the district boundaries or areas without detailed borehole data. This phenomenon is called edge effect. To avoid edge effects, cavities were evaluated for proximity to district boundaries or areas without enough borehole data. Some isolated cavities are very far away from the main populations. These areas have not been fully investigated and some cavities might exist, but may not be mapped or recorded in the database. Three kinds of cavities were removed for NNA: cavities that have nearest neighbors outside of the clustered area whose distance

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**Figure 10.**
Cavity distribution in relation to depth to Gachsaran Formation, depth to bedrock and depth to cavity follow a normal distribution.

**Figure 11.**
Cavity distribution in relation to cavity size follows poisson distribution indicating a random spatial distribution.
to the nearest neighbor (DNN) patterns are significantly different from those in the clustered area, cavities whose DNN are greater than the distance to the boundary of the project area, and some isolated cavities whose neighborhood has not been fully investigated for cavities by boreholes. The overall DNN distribution of all cavities does not follow Poisson, Normal, or Log-Normal distributions. However, the distribution of the DNN for all cavities more or less follows normal distribution once DNN is greater than 160m.

A decision tree model based on cavity characteristics and the distributions of cavities was developed for hazard assessment in the ADM area (Figure 13). The decision tree includes characteristics of bedrock geology, depth to the Gachsaran Formation, cavity density, cavity size, and distances to the nearest cavities in the ADM area. The primary controls on cavity development are lithostratigraphic position or bedrock geology and depth to the soluble Gachsaran Formation.

**Conclusions**

It is evident that soluble bedrock is definitely prone to cavity formation in comparison to insoluble rock, as majority of the cavities occurring in insoluble rock such as mudstone can be attributed to its weak compressive strength properties (Canton et. al, 2001). Out of the various soluble bedrock formations within the ADM the Gachsaran Formation has the highest likelihood for cavity formation. Cavities are formed either due to the dis-

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**Figure 12.** (a) Distance to the Nearest Cavity and (b) Median Distance to the Nth Nearest Cavity within the Gachsaran Formation.

**Figure 13.** The decision tree model that was developed for hazard assessment related to cavities in the ADM area.
solution of mudstone or gypsum layers in the formation, or at weak and weathered zones present at the interface between these two lithological materials. It is also evident that more cavities are likely to be formed in regions with shallow bedrock than in regions with relatively deeper bedrock.

Based on the cavity distribution in relation to depth to Gachsaran Formation (Figure 10) it is more likely that cavities are formed in locations where the Gachsaran formation occurs at a depth of less than 30 m (100 ft) below ground surface. Similarly, based on the histogram for distribution of cavities in relation to cavity size (Figure 11), it is statistically more likely that a cavity prone region develops smaller sized cavities (less than 3 m or 10 ft thick) than larger sized cavities (greater than or equal to 3 m or 10 ft in thickness). The majority of the cavity population tends to form in highly concentrated zones indicating that neighborhood effect plays a very important role in cavity distribution and formation.

The decision tree model quantifies depth to Gachsaran Formation, depth to cavity, cavity density and distances to the nearest cavity in the Abu Dhabi Municipality. This decision model, when compared with earlier, elementary versions of zone level cavity risk assessment studies, produces a more structured and objective approach towards analyzing patterns in the spatial distribution of cavities and supplements the existing cavity distribution maps when comparing the depth and resolution of evaluation. However, other influential parameters controlling formation of cavities, such as groundwater chemistry and fluctuation; land use and topography; and anthropogenic changes to landscape and groundwater, were not considered in the study due to the lack of sufficient data. While this decision tree model defines certain quantitative requisites for determining regions that are more susceptible to forming cavities, this decision process can only predict future occurrence of cavities with low accuracy as information relating to all cavities used in this study are solely collected from boring logs. This contributes to a lot of noise in the accuracy of cavity distribution. Also, in this study cavities are assumed as discontinuous 2D features, while in reality cavities tend to develop and propagate in vertical and lateral directions. Therefore, this decision tree model needs to be constantly updated and verified as newer site investigation studies are performed and made available.

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A METHOD OF MAPPING SINKHOLE SUSCEPTIBILITY USING A GEOGRAPHIC INFORMATION SYSTEM: A CASE STUDY FOR INTERSTATES IN THE KARST COUNTIES OF VIRGINIA

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Abstract
Karst is a landscape underlain chiefly by limestone that has been chemically dissolved by acidic groundwater, producing subsurface voids that may lead to sinkholes at the surface if the overlying soils can no longer support their own weight and collapse. The western counties of Virginia have a high concentration of karst areas due to widespread occurrence of carbonate rock exposures, and their geomorphic development within the Appalachian mountains. As a result, the Commonwealth of Virginia Hazard Mitigation Plan recommends that the Virginia Department of Transportation (VDOT) develop a method to determine the roadways and regions most susceptible to experiencing sinkholes, in an effort to reduce the possibility of reported sinkhole damage to property. While many noninvasive methods exist to detect subsurface voids, such as electric resistivity imaging, microgravity, ground penetrating radar, and seismic surveys, these methods are time consuming and costly.

This study proposes the use of a geographic information system (GIS) to create a susceptibility map of regions in the karst counties of Virginia, and in particular along interstate highways, that are most susceptible to future sinkhole development. Five factors that have previously been shown to play a role in the acceleration of sinkhole formation in Virginia include: bedrock type, proximity to fault lines, drainage class, slope of incised river banks, and minimum soil depth to bedrock. The analysis compares 1:24,000 scale maps of existing sinkholes developed by Virginia Department of Mines Minerals and Energy (DMME) with a series of maps representing differing combinations of each of the five factors to determine which weighted combination is most appropriate to use for a final representative sinkhole susceptibility map. The layers representing each factor are created using publicly available tabular and spatial data taken from the United States Department of Agriculture (USDA) Soil Survey Geographic (SSURGO) Database, the United States Geological Survey (USGS) National Map, the USGS Mineral Resources Online Data, and the National Weather Service. The methodology used to gather information specifically from the SSURGO database is highlighted within this paper. Data from the SSURGO database is used to create the bedrock type, drainage class, and minimum soil depth to bedrock layers. A substantial benefit to this methodology is that the new technique can be adjusted to accommodate for sinkhole susceptibility in other karst regions, by simply adjusting the input layers to consider the specific geology of a particular region.

Introduction
Karst terrain forms as acidic groundwater interacts with soluble bedrock, during which subsurface draining causes unique solutional patterns to carve into the rocks, forming cavities. The resulting voids introduce the potential to trigger land subsidence in the event that the topsoil filters into the voids, forming sinkholes (Hubbard, 2001).

The western counties of the state of Virginia contain abundant karst areas, because of the widespread occurrence of carbonate rock exposures, and their geomorphic development within the Appalachian mountains, ultimately locating the karst areas in long valleys containing extensive folds and fractures of limestone and dolomite bedrock (Belo, 2003). This folded and faulted geologic setting results in a regional topography defined by differential weathering of rock units, and provides a natural setting for karst terrain and sinkhole formation as carbonate strata are exposed at or near the surface.

Sinkholes pose engineering complications and the risk of damaging property and endangering lives if developed in a highly populated or well-traveled area. This paper focuses on the natural factors of sinkhole formation, and their combination within a geographic information system (GIS) in order to create maps of sinkhole susceptibility. While impossible to fully eliminate
natural karst hazards, losses and damages can be alleviated through effective implementation of investigative techniques where areas of greater sinkhole susceptibility may be identified (Dai et al., 2008; Ivey Burden, 2013). Due to the public availability of spatial and tabular datasets provided by agencies such as the United States Geological Survey (USGS), the use of GIS techniques has become significantly useful to state and local governments in the field of natural hazards (Whitman et al., 1999). This investigation proposes a new method of using a GIS and data from the Soil Survey Geographic Database (SSURGO) to create layers to predict where those sinkholes might form in an effort to avoid such dangers, specifically along Virginia highways. The results provide an inexpensive and quick method of better locating proposed roadway passages to aid in avoiding impact to karst areas (Moore et al., 2008) and determining which roadways may require immediate safety evaluations, ultimately minimizing environmental threats to life and property and maximizing land use (Muckel, 2004). Additionally, factors input into the methodology developed within this study could be adjusted to consider the geology of other karst regions with similar data availability.

**Background**

There have been few studies aiming to accurately estimate sinkhole risk due to the lack of detailed datasets, spatial analysis, and historical records on the subject. Karst maps were the sole method of assessing subsidence potential. However, the Virginia Hazard Mitigation Plan claims “a high percentage of karst geology in a jurisdiction does not necessarily [imply] that the whole locality is at risk for land subsidence” (Virginia Department of Emergency Management, 2003). Without a well-established set of guidelines that predict probabilities, a true risk determination cannot be formed.

In a study on sinkhole distribution in Virginia, Hubbard (2001) determined sinkhole locations by stereoscopic viewing and panchromatic aerial photography, field-checking any questionable sinkholes. Hubbard determined that detected sinkholes mainly occur in regions where carbonate rocks are present, where structural folds and faults exist, and where carbonate bedrock is adjacent to deeply incised rivers and tributaries (Hubbard, 2001). Additionally, it was noted that not all sinkholes can be detected by aerial stereophoto pairs, since aircraft tilt makes certain shallow sinkholes entirely unrecognizable while making other low slope regions appear as sinkholes when in fact they were none. Hubbard (2001) estimated that it would take 250 years to map every single sinkhole in Virginia’s Valley and Ridge province using solely aerial photography and field-checks. However, recent acquisitions of Light and Detection and Ranging (LIDAR) data in Virginia allow for the creation of highly accurate (sub-meter) elevation models that can be used for rapid, precise detection of sinkholes (Doctor and Young, 2013).

The growing body of sinkhole datasets has driven scientists to look further into the subject to develop trends that could be implemented into a GIS to create maps of sinkhole susceptibility across broad areas. Hyland (2005) verified that there is a correlation of sinkhole proximity with existing fault lines.

Water flow in karst areas is not manifested on the surface in karst regions as much as in non-karst regions because it percolates into the subsurface caves and conduits. Hence a region with rapid surface drainage, or a greater hydraulic gradient and lack of surface water, might imply subsurface drainage pathways that could potentially lead to the possibility of sinkhole formation in that particular area. Smaller surface streams often do not exist or endure as voids accommodate most of the water into the network of conduits and fissures, leaving only the stronger, more heavily flowing rivers to remain above ground. Proximity to these deeply incised rivers remaining above the surface is most likely indicative of a sinkhole susceptible region, because of the steepened hydraulic gradient and the resulting increase in groundwater flow for those areas (Muckel, 2004).

Green et al. (2002) decided that sinkhole risk studies should focus on shallower regions of bedrock, concluding that the timescale for which sinkholes may develop can be hours to months for shallow depth to bedrock, where it may be decades to centuries with a thicker depth to bedrock.

This study aims to create a sinkhole susceptibility analysis map by combining 5 factors – bedrock type, proximity to fault lines, drainage class, proximity to incised river banks, and depth of the overlying soil – into a single representative map spanning the western counties of Virginia. However, this paper specifically will explain in detail the methodology used to create the bedrock type, drainage class, and depth of overlying soil layers, which all call upon data specifically from the SSURGO database.

**Methodology**

**Study Area**

The region of interest in this study involves twenty-seven counties in Virginia that contain karst terrain, west of the Blue Ridge (Hubbard, 2001; Figure 1). The region
includes approximately 29,853 square kilometers and ranges from $-83^\circ40'32''$ to $-77^\circ19'42''$ latitude and from $39^\circ27'57''$ to $36^\circ35'37''$ longitude.

**Data Acquisition and Preparation**

For the final representative sinkhole susceptibility map, data was taken from four sources. Bedrock type, depth to bedrock, drainage classes, and county and map unit boundaries were obtained from the SSURGO database (websoilsurvey.nrcs.usda.gov), digital elevation models and the Virginia state boundary were from the USGS National Map (nationalmap.gov), fault lines were downloaded from the USGS Mineral Resources Online Spatial Data (mrdata.usgs.gov), and rivers were obtained from the National Weather Service website (weather.gov). The analysis created a compiled ranked map determining regions of potential sinkhole formation based on five unique layers created in ArcMap 10.1. Each layer contained a map of regions assigned a ‘Sinkhole Value’ ranging from 1-15 (1 is low susceptibility and 15 is high susceptibility), representing the level of potential hazard based on the corresponding risk factor. The distinct maps were ultimately combined using weights representing the corresponding factor’s influence on predicting sinkhole regions, and then the most appropriate combination was statistically determined using a residual sum of squared errors test, for use in the final representative susceptibility map.

The main contribution of this paper, however, is the methodology created to extract relevant sinkhole layer data from the SSURGO database. This database has been created by the National Cooperative Soil Survey over many years and spatially references surface soil data at scales ranging from 1:12,000 to 1:63,360. The database offers a detailed description of the surveyed soils; however, since it is a collection of soil descriptions from various soil scientists, it is a notoriously difficult database to work with as it contains very little uniformity among entries, varying soil descriptions, and some duplicate entries. In the creation of the bedrock type, drainage class, and depth to bedrock layers for the final representative sinkhole susceptibility map produced by this study, Python codes were created to efficiently sort and gather relevant information from the SSURGO tables for the creation of the aforementioned layers.

**Bedrock Type**

Bedrock type is a contributing factor to sinkhole formation since sinkholes have proven to form in regions of relatively pure carbonate rocks. The bedrock type layer was derived from SSURGO tabular data located in the Component and the Component Parent Material (COPM) tables for each of the 27 counties of interest. Desired attributes from the individual tables were combined into a single table based on common fields through a Python script, converted to pseudo-code for simplicity:

```python
    # Loop through each county’s SSURGO database
    # Select the parent material field in the database for each map unit and add to a new table
    # Add a new column to table called “Sinkhole Value”
```
A Python script was written creating a table that combined each map unit with its corresponding drainage class.

Only 7 different drainage classes were listed, so the regions with poor drainage had Sinkhole Values defined as the odd numbers ranging from 1 to 5, and regions that were excessively drained had Sinkhole Values defined as the odd numbers ranging from 9 to 15, since this most likely meant water was being absorbed into subsurface karstic drainage systems. Zero was assigned to the excess regions with no drainage class assignment. Sinkhole Value 7 had no assignment in this layer. (Figure 2C).

**Drainage Class**

Drainage plays a role in predicting sinkhole risk occurrence because it provides information on how rapidly water will drain through the soil type. Drainage class (drainagecl) was a field defined in the SSURGO Component table, composed of values ranging from excessively drained to very poorly drained. A Python script was written creating a table that combined each map unit with its corresponding drainage class.

The resulting output table was imported into ArcMap, where it was spatially joined with the map units. The depth ranges were converted to raster and were then reclassified into 15 equal intervals, defined for each 15cm increment below the surface. Shallow soil cover overlying the bedrock was assigned the highest Sinkhole Value of 15, and the deepest amount of soil cover was assigned the lowest Sinkhole Value of 1 (Figure 2E).

**Depth of Overlying Soil**

The timescale during which sinkholes may develop is much shorter in regions of shallow depth to bedrock. Thus a sinkhole factor layer representing depth to bedrock was created using SSURGO data found in the Mapunit Aggregated Attribute (MUAGGATT) table. The bedrock minimum depth entry for each map unit along the 27 counties was recorded into a table using a Python script, based on a common map unit key found in the Mapunit and the MUAGGATT tables.

The resulting output table was imported into ArcMap, where it was spatially joined with the map units. The depth ranges were converted to raster and were then reclassified into 15 equal intervals, defined for each 15cm increment below the surface. Shallow soil cover overlying the bedrock was assigned the highest Sinkhole Value of 15, and the deepest amount of soil cover was assigned the lowest Sinkhole Value of 1 (Figure 2E).

**Proximity to Fault Lines**

Hyland (2005) determined a correlation between sinkhole formation and proximity to existing fault lines. To create this fault risk layer in the final risk map in ArcMap, the “Multiple Ring Buffer” tool was used around the fault lines extracted from USGS Mineral Resources Online Spatial Data. Rings ranging from 0 – 3000 feet from the faults in increments of 200 feet were created and converted to raster, then reclassified into Sinkhole Values of 1 through 15. Values were highest at regions closest to the faults, decreasing outwardly as distance increased from faults (Figure 2B).

```python
# Loop through the table, and define the “Sinkhole Value” for every map unit depending on its parent material
# Define pure limestone parent material a Sinkhole Value = 4
# Define combined limestone and dolomite parent materials a Sinkhole Value = 2
# Define other calcareous parent materials a Sinkhole Value = 3
# Define all other parent materials a Sinkhole Value = 1

The parent material kind and origin were defined in the COPM table, where formation kind was described as “alluvium”, “colluvium”, or “residuum”, and formation origin ranged through a number of bedrock origins, such as limestone, sandstone, or shale. The Component table defined the corresponding representative component percentage. The script removed any duplicate or null entries as well as any entries where the parent material kind was alluvial, since alluvium soils rarely play a role in karst development. If a sinkhole does in fact exist in a region with alluvial deposits, it is more likely a result of one of the other factors in this study (proximity to faults or depth to bedrock, for example) rather than a result of the bedrock type (Hyland, 2005).

A new field called the “Unweighted Sinkhole Value” was added to the resulting table, determined by the parent material origin, assigning pure limestone origins the highest value of 4, limestones and dolomite combinations a value of 3, values containing only partial carbonate rock a value of 2, and entirely clastic and non-carbonate origins the lowest value of 1. The final table from the script was imported into Microsoft Excel, where a final weighted Sinkhole Value was calculated per map unit using the component percentage and the Unweighted Sinkhole Value:

- Component1 Percentage x Unweighted Sinkhole Value1 +
- Component2 Percentage x Unweighted Sinkhole Value2 +
  ...
  = Final Sinkhole Value for Corresponding Map Unit

While the final weighted Sinkhole Values ranged from 1-4 as non-integer values, the values were reclassified into 15 equally incremented categories to remain consistent with the next four layers. This final table was imported into a Microsoft Access Database to then be imported into ArcMap, where it could be joined spatially with the spatial map units using the “Add/Join” tool in ArcMap (Figure 2A).

**Drainage Class**

Drainage plays a role in predicting sinkhole risk occurrence because it provides information on how rapidly water will drain through the soil type. Drainage class (drainagecl) was a field defined in the SSURGO Component table, composed of values ranging from excessively drained to very poorly drained. A Python script was written creating a table that combined each map unit with its corresponding drainage class.

Only 7 different drainage classes were listed, so the regions with poor drainage had Sinkhole Values defined as the odd numbers ranging from 1 to 5, and regions that were excessively drained had Sinkhole Values defined as the odd numbers ranging from 9 to 15, since this most likely meant water was being absorbed into subsurface karstic drainage systems. Zero was assigned to the excess regions with no drainage class assignment. Sinkhole Value 7 had no assignment in this layer. (Figure 2C).

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Slope of Incised River Banks
A higher risk for sinkhole formation near incised river banks can be attributed to a higher hydraulic gradient and resulting increase in nearby groundwater flow. To determine the degree of incised rivers, Virginia rivers downloaded from USNWS geospatial data were added to the map layer, with a buffer region constructed using the “Multiple Ring Buffer” tool in ArcMap to account for river widths. Slopes along riverbanks were identified based on digital elevation model rasters imported from the USGS National Map and clipped to only exist within half-mile buffer zones around each river. The slopes of the elevation model rasters were computed using the ArcMap “Slope” tool. The total range of slopes were split into 15 equally incremented categories and slopes of regions were reclassified into one of the 15 consequent categories, assigning low slopes in the regions surrounding rivers to a Sinkhole Value of 1 and high slopes to a Sinkhole Value of 15 (Figure 2D).

Statistical Analysis (Creating Weights)
In order to analyze how individual factors influenced sinkhole occurrence in the karst counties of Virginia, twenty-eight different risk maps were created using the ArcMap Raster Calculator by combining the five individual risk layer raster images using a series of different chosen weights, as seen in the equation below:

\[(A \times \text{Bedrock Type}) + (B \times \text{Proximity to Faults}) + (C \times \text{Minimum Depth to Bedrock}) + (D \times \text{Drainage Class}) + (E \times \text{Slope of Incised River Banks}) = \text{Weighted Combination Susceptibility Map}\]

where A, B, C, D, and E are chosen weights assigned to its corresponding combination, and A+B+C+D+E = 1. Because of the infinite possibilities of weight assignments, values were assigned in increments of one tenth. Bedrock type has been shown in existing karst literature to be the most influential risk factor contributing to sinkhole formation, therefore combinations were chosen giv-
ing bedrock type the highest weight for all combinations except a control combination (combination 1), where each layer was assigned equal weight.

Upon completion of the 28 distinctly weighted risk maps, the map that most closely and statistically corresponded to the imported data of mapped sinkhole locations would be chosen. A Python script that could loop through each combination, automating the following steps, was constructed. Spatial data containing existing mapped sinkholes was added into the map. Each weighted risk map was converted from a raster image to a polygon shapefile based on its Sinkhole Level (1-15) so that it could be clipped into the boundary of the existing sinkholes. Using the “Dissolve” tool, the newly clipped risk map was condensed into fifteen total zones, based on its defined sinkhole levels. A new field was added to the attribute table of the polygon to calculate the total areas of each distinctive Sinkhole Level. In Excel, the areas of sinkholes corresponding to each Sinkhole Level, the total area of existing sinkholes, and the percentages of each individual level compared to the total sinkhole area were computed and recorded. For simplicity, the 15 Sinkhole Levels were condensed into five risk zones, defining values 1-3 as a Low Risk Zone, 4-6 as a Medium-Low Risk Zone, 7-9 as a Medium Risk Zone, 10-12 as Medium-High Risk Zone, and 13-15 as a High Risk Zone.

The ideal percentages of existing sinkhole areas per risk zone were defined so they could be statistically compared with the observed percentages using a Residual Sum of Square (RSS) error test. In an ideal situation, there would be no actual sinkholes found in the Low Risk Zone and the highest percentage of actual mapped sinkholes would be found in the High Risk Zone, with a linear relationship between those zones in between and the sum of the total percentages being 100. Hence this investigation predicted that percentages should be 0, 10, 20, 30, and 40% respectively, per increasing risk zone. Error between the actual and the ideal models exemplifies how well the experimental data fits the expected, so the goal is to minimize the RSS. RSS values for each combination were computed and ranked. Thus the combination with the lowest RSS was the value most closely matching the ideal situation and was used for the final map.

Results

Final Map with Interpretation

The final map (Figure 3), created from combination 25 which had the smallest RSS when compared with the predicted model, was created using the following equation:

\[
\text{Weighted Map} = (0.6x\text{BedrockLayer}) + (0.1x\text{ProximityToSinkholes}) + (0.2x\text{DrainageClass}) + (0x\text{SlopeOfIncisedRiverBanks}) + (0.1x\text{DepthToBedrockLayer})
\]

Figure 6 also displays a USGS Karst Terrain map beside the final sinkhole risk map. Karst terrain is defined to be terrain containing subsurface fissures, caverns, and voids resulting from chemically dissolved limestone bedrock, thus if the two maps relate, we can be confident in the conclusions reached. From a visual comparison, the resulting values make sense, since the regions with karst terrain on the USGS map align with the higher risk regions on the sinkhole risk map.

Sources of Error

The final weights used to combine the layers into the ultimate sinkhole susceptibility map were based on a statistical comparison between the constructed predicted at risk regions and existing sinkholes. However, to make clear that his sinkholes were mapped as a guideline and not a set of perfectly defined structures, Hubbard digitized his data using a scale that was 10 times less accurate than the scale the public had requested. Furthermore, the aerial photography used in Hubbard’s study cannot accurately detect all sinkholes due to aircraft tilt, which creates the illusion of a sinkhole where it may not exist and does not recognize shallower sinkholes at all. The incomplete or flawed representation of sinkhole locations serve as a source of error in choosing the appropriate weights for the combination of risk layers.

By using the USDA Soil Survey data, a degree of error was inevitable due to the fact that data tables from which a significant portion of the risk layers were derived were incomplete for specific factors in a majority of counties within the study. Additionally, SSURGO data contained values corresponding to Virginia counties but not all major Virginia cities, which likely have their own GIS database. This led to voids in the final map.

The type of data available further limited the scope of this study. The slope of incised riverbanks risk layer was determined using digital elevation models, and raster images were created using remote sensing or based on existing topographic maps. While these are helpful for analyzing the surface topography of the region, difficulties in karst analyses arise since the directions of subsurface flow and hydrologic connections between sinkholes or subsurface aquifers are not represented in them (Taylor et al., 2008). It would be much more useful to know the direction of water flow in those surrounding incised river banks, since large slopes on the surface do not necessarily denote fast paced water travel through subsurface fissures and waterways.
Finally, risk layers compiled in creating the final map were created and assigned Sinkhole Values based on hypotheses derived from background knowledge and engineering judgment. It is always possible that a sinkhole value was assigned incorrectly.

Conclusion
This analysis used a geographic information system and readily available data from the SSURGO database to create a map that represents regions most at risk of sinkhole formation in the karst counties of Virginia. While these results serve as a general guideline for mapping karst regions in Virginia, it is important to understand that a risk map created based on generalities cannot be substituted for a site-specific analysis. Different karst landforms relate to one another, but the combinations and behaviors of the relationship between local, hydrological, and climactic conditions are numerous (Ford and Williams, 1989). While this map provides a general understanding of karst terrain in Virginia, the final product and the methods taken to reach it are specific to the region of interest. In order to apply this technology to new regions, a thorough background of the geology for that region is necessary and factor layers for sinkhole formation be adjusted accordingly. However, once factors for a new region have been defined, this methodology can be applied to produce similar results.

References

Figure 3. Comparison of USGS karst terrain (Davies et al., 1984) with final risk map.


FINDINGS SPRINGS IN THE FILE CABINET

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Abstract
The Minnesota Pollution Control Agency (MPCA), in partnership with other agencies, is currently undertaking comprehensive sub-basin assessments statewide over a ten-year period. Southeast Minnesota has over 17,500 kilometers of perennial and intermittent streams, making the task of comprehensive sub-basin assessment challenging; the task is further complicated by karst geology. In the summer of 2014, a pilot project began between the MPCA and Minnesota Department of Natural Resources (DNR) to digitally preserve paper documents which capture qualitative and quantitative data about the hydrology, water chemistry, geomorphology, biology, land use and karst features of southeast Minnesota streams. The paper documents in file cabinets were not in an accessible or easy-to-use format; as such, they were in a 'data silo.' The task was to preserve the documents so as to make the data usable by converting the documents into a digital format (Adobe PDF, GeoTIFF, ESRI Feature Class). To date, more than 4,000 documents (of an estimated more than 12,000) have been converted, made text-searchable, prepared for storage in a document management system, and made more accessible through a geographic information system (GIS). This previously inaccessible data is an important piece in understanding the karst region of southeast Minnesota. Within the documents scanned thus far, over 400 springs and other karst features have been identified, which are not currently recorded in Minnesota’s Karst Feature GIS Database.

Introduction
The karst region of the “Driftless Area” in southeast Minnesota proves to be challenging for the Minnesota Pollution Control Agency (MPCA), which is tasked with assessing the streams of that area. Karst features, such as sinkholes and springs, can change the properties of a stream. For example, the presence of sinkholes can modify the contributing drainage area to a stream and a groundwater-fed spring can change temperature, chemistry and flow of a stream.

Stream assessment is conducted to determine if the stream supports its designated use, such as supporting aquatic life. Assessment is part of the Minnesota Water Quality Framework’s Watershed Restoration and Protection Strategies (WRAPS) process — the comprehensive monitoring and assessment of Minnesota’s 80 major watersheds (United States Geological Survey Watershed Boundary Dataset subbasins (HUC08)) on a ten-year cycle. This process includes identifying water quality impairments, sources of pollution, areas in need of protection and restoration, and strategies to achieve and maintain water quality standards and goals.

Stream assessment requires the review of existing information as well as the collection of new data. Of particular importance is the identification of karst features, because karst features can play a controlling role with respect to a stream’s physical, chemical and biological properties. Existing information may come from previous work by the MPCA or, likely, from other agencies. These may be other state agencies, or may be federal, local or non-profit agencies. Existing information may also come from landowners or local land users (e.g., hunters, fishers, birders, fungi and plant collectors).

The opportunity for collaboration among these agencies is great; the challenge is lack of resources (monetary or time) or organizational barriers, which may inhibit proactive communication and collaboration. Simply stated, one agency may not know what data and research another organization has or is working on. When data is not proactively shared or accessible and searchable, it is not practically usable.

Such was the case with the Minnesota Department of Natural Resources (DNR) stream documents of southeast Minnesota. The MPCA and DNR have a strong history of collaboration, but the challenge was that the paper
documents in file cabinets were not in an accessible or easy-to-use format; as such, they were in a ‘data silo.’ In the summer of 2014, MPCA staff recognized that there was an immediate business need for the DNR’s stream documents to be in a digital format. Therefore, a pilot project began between the MPCA and DNR to digitally preserve the stream documents which capture qualitative and quantitative data about the hydrology, water chemistry, geomorphology, biology, land use and karst features of southeast Minnesota streams. The MPCA recognized that this data was valuable and could be an important resource for stream assessment (e.g., identifying interrelated factors impacting a stream’s biologic community), especially because these documents contained information about the location and characteristics of karst features.

By digitizing these stream documents, the data would become:

- Protected from physical damage or loss.
- Easily accessible, searchable, shareable.
- A collective resource enhancing one another, thus producing a more spatially and chronologically complete story of a particular stream.

**Process**

The task was to preserve the data contained in the documents so as to make the data usable. This was accomplished by converting the documents into a digital format (Adobe PDF, GeoTIFF, ESRI Feature Class). Proceeding by watershed, MPCA staff removed folders from the file cabinets one at a time. Each document within a folder was carefully prepared by removing staples and repairing any damage prior to scanning each to a computer in Adobe PDF format (these are herein referred to as the digital documents). In addition to the scanned copy saved on a computer, a backup copy was saved to an external hard drive. Documents were then reassembled and returned to their respective folders in the file cabinets. Finally, folders were marked to indicate that the scanning of all of the documents in each folder had been completed. A key part of this process was the document-naming scheme. Each digital document was named in a way to effectively “tag” it with keywords such that it could be seamlessly loaded into the MPCA’s OnBase document management system for storage and retrieval. The digital document-naming scheme also allowed for the digital documents to be accessed through ESRI ArcGIS. Once scanned, Adobe Acrobat Professional software was used to make the digital documents text searchable and to create a searchable index. The indexed documents were organized and placed into folders by document name (e.g., stream assessments, survey reports, survey summaries, etc.).

To incorporate the digital documents into a GIS format, many steps were taken to ensure all documents were linked with their related stream feature class. After the scanned documents were indexed with Adobe Acrobat Pro, a log file was created of all the file paths and document names. The log file was imported into a Microsoft Excel Workbook where file paths were organized by theme, each theme a Microsoft Excel Worksheet (e.g., assessments, survey reports, summaries, etc.). Worksheets were then converted to an ESRI geodatabase table. In ESRI ArcMap, each theme table was related by kittle number (unique stream numbering system developed by the MN DNR) to a stream feature. As a result, users are able to use the Identify Tool to view PDF documents associated with a stream reach, simply by clicking on the feature in the ArcMap software.

In addition to the stream documents (which were generally typed or hand-written text documents including a few illustrations or maps), there were 147 large (greater than 21.59 cm x 27.95 cm) maps. These maps were generally USGS 7.5 minute series topographic map quadrangles, with hand-written annotation. On these maps were identified springs, seeps, electrofishing stations and other notes, such as water temperature and qualitative land cover descriptions, such as “good pheasant habitat.” These maps were sent to a private contractor who was able to scan these larger size documents. The maps were returned to the MPCA as digital images (in Tagged Image File Format [TIFF]). Within the ArcMap environment, the digital images were georectified, or geographically referenced. Once georectified, the hand annotation could be digitized; in other words, points of interest such as springs in the digital image were made into a GIS feature class.

**Discussion**

Thus far, over 4,000 paper documents have been made digital, text searchable and indexed. In practical terms, a MPCA staff member could search for a certain word within a digital document or within the whole collection of digital documents, enabling a stream assessment team...
to easily search for karst features mentioned within a stream’s digital documents. Furthermore, in OnBase the digital documents serve as a back up to the paper document in the DNR’s file cabinet. Within OnBase, documents can be found using keywords such as the DNR Stream Kittle Number, year (approximate date of creation of the original paper document) or document type (stream survey, stocking record, stream assessment, etc.). The storage of these digital documents in OnBase should be a more efficient way (rather than researching paper documents while physically sitting in the DNR office) to manage and share this data within the agency as well as outside the agency. Furthermore, the text searchable digital documents allow for copying and pasting text from the digital document to another type of document, such as a word processing document.

In ArcMap, the relationship between the stream feature class and the tables (tables that include the file paths to the digital documents), allow MPCA staff to use ESRI ArcMap’s Identify tool to select a stream of interest and be provided with a list of digital documents associated with that stream. Using ESRI ArcMap’s Hyperlink functionality, MPCA staff can click on a document listed within the Identify Tool window and open the digital document in Adobe Acrobat Reader — effectively digitally opening up the DNR folder, formerly only possible by visiting the physical office. The combination of accessing the digital documents from ArcMap and the text searchability of the digital documents allow MPCA staff to not only find springs mentioned within the digital document, but also to digitize that information as a spring feature (that is geographically referenced) in ArcMap, based on the information in the text. For example, a digital document’s text that reads “spring, 0.5 cfs, 200 meters upstream from Bridge 2” may be sufficient to orient the reader to that digital document; however, when that spring location is captured as a geographically-referenced feature in a GIS feature class, it can be examined in context with other features on a sub-basin-wide scale. As these springs and other karst features are digitized through this process, they can be appended to the respective feature class.

From the 147 large format maps, digitizing was conducted to create a GIS feature class of springs from the hand-annotated locations on the maps. Next, the feature class was compared to the existing spring features within the Karst Feature Database created by the Minnesota Geological Survey and University of Minnesota. The comparison resulted in the determination that this process had identified 400 springs and other karst features not already in the Karst Feature Database. These 400 new features were shared with DNR, Minnesota Geological Survey and University of Minnesota staff in hopes of field verification and eventual inclusion within the database. The great benefit of this pilot project is providing Minnesota Pollution Control Agency staff access to robust existing information for their stream assessment process; this information is particularly important for the complex karst region of southeast Minnesota. This was accomplished by liberating data about southeast Minnesota’s streams and karst features from the ‘data silo’ that is a file cabinet of paper documents. The resulting digital documents, digital images, and GIS feature classes has made the data more manageable, accessible, and shareable.
NEW METHODOLOGIES AND APPROACHES FOR MAPPING FORESTED KARST LANDSCAPES, VANCOUVER ISLAND, BRITISH COLUMBIA, CANADA

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Abstract  
Mapping is an essential tool for land management and is typically used to assess the nature and characteristics of a land surface, along with its resource features and values. Mapping of karst landscapes is of particular importance for the temperate rainforests of Vancouver Island on the west coast of British Columbia (BC), where both forestry and natural resource development activities occur. A set of BC Government standards for mapping karst have been developed at varying scales (reconnaissance, planning-level and detailed), and incorporate procedures to assess the potential and vulnerability of karst, applying qualitative analysis of various surface and subsurface karst attributes. Previously, karst maps have been compiled by GIS mapping software and generated as static graphic images. However, it is now possible to make maps more accessible and interactive by uploading them as overlays within Google Earth (as KML files) or other imagery platforms. It is also possible for these maps to be taken into the field using tablets, phones, or iPads, thus allowing for on-site data collection and resource evaluation. A key focus of this research is the development of a `Karst Map of Vancouver Island’ that outlines the known extent of the karst and the likely contributing non-karst catchments. The mapping will also be used to identify regions of varying karst potential/vulnerability and notable karst areas/features. Trials have been completed to see how detailed imagery of karst features and areas collected by UAV technology can be used for karst evaluation. Karst areas have also been mapped using LiDAR, which has the potential for detecting surface karst features beneath the forested canopy and for assessing the overall sensitivity of forested karst areas.
EVALUATION OF VETERINARY PHARMACEUTICAL AND IODINE FOR USE AS A GROUNDWATER TRACER IN HYDROLOGIC INVESTIGATION OF CONTAMINATION RELATED TO DAIRY CATTLE OPERATIONS

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Abstract
Standard groundwater tracers such as Rhodamine WT, Fluorescein, Eosin and Tinopal CBX effectively provide a snapshot of hydrological conditions over a brief period of time and in a tightly controlled setting. However, in complex environmental situations with multiple potential sources, groundwater hydrologists are often seeking groundwater tracers that have extended longevity in the natural environment and the ability to directly pinpoint source locations. After reviewing operations of the nearby dairy it was determined that emerging contaminants, specifically two bovine veterinary pharmaceuticals (antibiotics), cephapirin sodium (CEPNa) and cephapirin benzathine (CEPB), and a sanitation agent, elemental Iodine (I) may have potential as extended longevity groundwater tracers if analytical methodology could be established. Initially, sample analysis indicated that cephapirin is undetectable in unconcentrated samples of lagoon wastewater at parts per billion (ppb) concentration; pre-concentrated samples which utilized solid phase extraction allowed for better detection at part per trillion level. Concentrated samples from one of the two lagoon cells sampled (cell #3), detected cephapirin at 13.14 ppt level, while cell #1 failed to detect any cephapirin present. Controlled laboratory testing later indicated that in a wastewater environment cephapirin degrades to approximately 20% of initial concentrations within 4 days, with complete degradation within 6 days. Degradation patterns in surface water and groundwater samples were less dramatic and at slower rates. Degradation curves of the surface and groundwater samples indicate that concentrations of cephapirin are still detectable for approximately 25 days. Unconcentrated Iodine samples collected in lagoon cells ranged from 50.896 ppb and 1,704.55 ppb with variations determined to be a result of the primary inflow of the lagoon. Cephapirin’s use as a long term groundwater tracer does not seem to be an immediate option. Further research may reveal that its degradation products are potentially useful as a tracer. In some instances, such as catastrophic discharges of large volumes of milk when samples can be collected and analyzed quickly, the use of cephapirin as an environmental tracer may prove possible. The validity of pharmaceutical iodine as a groundwater tracer appears to be much greater than that of cephapirin. Iodine was detected in all of the environmental samples including the highly organic and anaerobic environment of the dairy wastewater lagoon. This study concludes that iodine is capable of surviving the hostile wastewater environment. If sufficient data is collected to determine natural background levels, iodine may prove useful in determining hydrological connections between iodine laden dairy effluent and the underlying groundwater.

Introduction
Investigation of the hydrologic environment in the vicinity of a fitness center in Southeast Missouri began in the summer of 2010. This investigation included a series of water traces utilizing the standard fluorescent dyes such as fluorescein and Rhodamine WT to evaluate potential sources of bacteria entering the center’s well. The sources included surface water infiltration surrounding the wellhead, onsite septic system failures and a nearby dairy operation (Pierce, 2012).

After reviewing operations of the nearby dairy it was determined that additional groundwater tracers may be present in the form of veterinary pharmaceuticals and sanitizing agents. A two-fold investigation was then initiated. The first phase of the investigation was to determine if these pharmaceuticals/sanitizing agents were detectable in a laboratory environment. Once the analysis methodology was proven to be successful,
If a hydrologic connection between the CAFFC well and the nearby dairy operation exist then testing should reveal cephapirin, used in the dairy operation, to be present in CAFFC water samples. Additionally, iodine levels should be elevated in water samples when compared to natural background iodine concentrations.

To determine if Cephapirin was readily detectable Dr. Honglan Shi at Missouri University of Science and Technology was contracted to develop liquid chromatography-tandem mass spectrometry (LC-MS/MS) methods to analyze water samples for the bovine antibiotic. In addition Dr. Shi developed protocol for inductively coupled plasma mass spectrometry (ICP-MS) to analyze for total iodine concentrations. Upon completion, the methodological process was put into service to analyze for these constituents in the field samples associated with the CAFFC well investigation.

Cephapirin and Iodine Analysis Development

The results of the cephapirin analysis methodology development were mixed. Dr. Shi was able to successfully develop a reliable, fast and simple method for detecting the base Cephapirin (CEP) molecule in water samples to the part per trillion (ppt) level. However difficulties were also observed in the analysis of field samples, stability studies showed that the Cephapirin degrades to form a degradant (DACEP) in all sample matrices (water and lagoon effluent). In a series of further test conducted on spiked samples the degradation was shown to be highest in the lagoon samples with complete degradation of CEP to DACEP within six days when left at room temperature. Figure 1 highlights these degradation rates. Analysis of samples in both groundwater and surface water matrices also show similar degradation patterns, but at a slower rate. Testing revealed that the currently analyzable degradants was also unstable and quickly degraded to undetectable levels.

The ICP-MS analysis methodology for total iodine in water samples was previously developed by Dr. Shi and the MS&T laboratory in 2009 and would be used for the investigation (Shi and Adams, 2009).

Field Sampling and Investigation

As part of the initial dye tracing, a well survey was conducted by MGS to locate and inventory any groundwater wells servicing residence, farms, or
collected from each cell of dairy’s two-cell wastewater lagoons. A representative population of inventoried wells was selected for sampling and analysis to establish natural background levels of iodine. As well as a small nearby spring and surface water samples from Williams Creek and the small tributary below the dairy operation. A total of nineteen water samples were initially collected for analysis. Two additional samples were collected at a later date to allow for the SPE and pre-concentration analysis. Figure 2 is a map showing all sample locations.

Samples were collected using procedures to meet MDNR standards and the methodology needs for Dr. Shi’s analysis. Water samples were collected in pre-cleaned, 4-ounce, Teflon- capped, wide-mouth amber bottles. For well water collection at homes, the faucet aerator was first removed (if present) and the water allowed to flow for approximately 5 minutes. Sample bottles were then filled. Surface water sample bottles were submersed and filled directly in the sampling bottle. For wastewater

Following the survey, wells in the vicinity of the CAFFC site were sampled and analyzed for the presence of cephapirin and iodine. Wastewater samples were

![Figure 1. Cephapirin degradation patterns by matrix.](image1)

![Figure 2. Sample location map.](image2)
samples a reusable sample dipper was utilized and subsequently cleaned and decontaminated between uses to avoid cross-contamination. Field blanks, trip blanks, and duplicate samples were collected throughout the sampling event as a quality assurance measure. The collected samples were sealed and placed in an iced cooler then transferred to the University of Missouri Science and Technology laboratory using standard MDNR chain of custody protocol.

Results
Analysis of collected samples initially indicated no detection of cephapirin in the wastewater lagoons at part per billion (ppb) levels. A second round of sampling was conducted on the wastewater lagoons. At this time larger volumes of sample were collected so that a solid phase extraction (SPE) could be used to pre-concentrate the cephapirin and allow for better detection sensitivity to the PPT level. In this analysis large volumes of lagoon waste were reduced to approximately 10% of its original amount while still retaining the concentrations of cephapirin. In this second analysis a two liter sample taken from the taken from cell #3 of the dairy lagoon treatment system, detected 13.14 ppt of Cephapirin. Analysis of a sample from cell #1 failed to detect any cephapirin.

Iodine levels collected from the study area ranged from 4.939 ppb to 1704.55 ppb and are shown in Table 1. Iodine was detected in all eight drinking water wells with concentrations ranging from 4.939 to 24.712 ppb with a mean concentration of 12.532 ppb and a standard deviation of 6.672 ppb. Six surface water samples were collected and analyzed with concentrations ranging from 6.671 to 21.247 ppb. The mean concentration of iodine in surface water samples was 14.732 ppb with a standard deviation of 5.489 ppb. A single sample from a shallow spring was collected and determined to have 5.855 ppb iodine present.

Samples collected and analyzed from lagoon cell #1 and lagoon cell #3 contained 50.896 ppb and 1,704.55 ppb iodine, respectively. A considerable difference in iodine concentrations is obvious in the lagoon samples. Further investigation revealed that the lagoons are not successive and that the source area and influents of waste were different for each. The primary waste source for lagoon cell #1 is runoff from the animal feed lot and loafing areas. Lagoon cell #3 received some barnyard runoff similar to cell #1. However, the influent of wash water from the milking parlor was also observed discharging into the cell. Since iodine is used to sanitize teats prior to milking its presence in the daily cleaning effluent is not uncommon and likely explains the large difference in iodine concentration between cell #1 and cell #3.

Conclusion
Due to the instability of cephapirin its use as a long term groundwater tracer does not seem to be an immediate option. Future research may at some point identify a specific cephapirin degradant product that has the stability and environmental longevity required for groundwater tracing, but at this time those degradants have yet to be identified. In some instances such as direct runoff, catastrophic lagoon failures or discharges of large volumes of milk, where samples can be collected and analyzed in a rather quick time frame, the use of cephapirin or its degradants as an environmental tracer may prove possible.

Table 1. Sample locations and iodine concentrations.

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Sample Name</th>
<th>Concentration PPB</th>
</tr>
</thead>
<tbody>
<tr>
<td>FB001</td>
<td>Field Blank</td>
<td>&lt;0.2</td>
</tr>
<tr>
<td>TB001</td>
<td>Trip Blank</td>
<td>&lt;0.2</td>
</tr>
<tr>
<td>WW001</td>
<td>Class Act Family Fitness Center</td>
<td>16.178</td>
</tr>
<tr>
<td>WW002</td>
<td>Koehler Engineering Well</td>
<td>11.472</td>
</tr>
<tr>
<td>WW003</td>
<td>University Farm Well</td>
<td>11.93</td>
</tr>
<tr>
<td>WW004</td>
<td>Volkerding Well</td>
<td>4.939</td>
</tr>
<tr>
<td>WW005</td>
<td>New Midway Dairy Well</td>
<td>24.712</td>
</tr>
<tr>
<td>WW006</td>
<td>Old Midway Dairy Well</td>
<td>6.255</td>
</tr>
<tr>
<td>WW007</td>
<td>Dave Brown Well</td>
<td>19.31</td>
</tr>
<tr>
<td>WW008</td>
<td>Wayne Eakins Well</td>
<td>5.457</td>
</tr>
<tr>
<td>SS001</td>
<td>Upper Williams Creek</td>
<td>6.671</td>
</tr>
<tr>
<td>SS002</td>
<td>Lower Williams Creek</td>
<td>9.216</td>
</tr>
<tr>
<td>SS003</td>
<td>Tributary at Shale Lane</td>
<td>21.247</td>
</tr>
<tr>
<td>SS004</td>
<td>Tributary just above confluence with Williams Creek</td>
<td>20.23</td>
</tr>
<tr>
<td>SS005</td>
<td>Tributary just below outfall of dairy lagoon #3</td>
<td>13.117</td>
</tr>
<tr>
<td>SS006</td>
<td>Tributary above dairy lagoon outfall (Ponco Lane)</td>
<td>17.913</td>
</tr>
<tr>
<td>SP001</td>
<td>Spring on University Farm</td>
<td>5.855</td>
</tr>
<tr>
<td>LG001</td>
<td>Lagoon Cell #1</td>
<td>50.896</td>
</tr>
<tr>
<td>LG002</td>
<td>Lagoon Cell #3</td>
<td>1704.55</td>
</tr>
</tbody>
</table>
The validity of iodine as an extended-longevity groundwater tracer appears to be much higher than the cephapirin. Iodine was detected in all of the environmental samples including the highly organic and anaerobic environment of the dairy wastewater lagoon. Iodine levels within dairy lagoons #1 and #3 were found to be 206% to 6898% higher than background samples. This study concludes that iodine is capable of surviving a hostile wastewater environment. If sufficient data is collected to determine natural background levels, the use of iodine may prove useful in determining hydrological connections between iodine laden dairy effluent and the underlying groundwater.

References
KARST INFLUENCE IN THE CREATION OF A PFC MEGAPLUME

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Abstract
Perfluorochemicals (PFCs) are fully-fluorinated organic chemicals used to produce a wide range of industrial and commercial products. Their extreme persistence and mobility in the environment and nearly ubiquitous presence in humans and wildlife has raised serious concerns regarding their environmental and human health effects. In the 1940s to 1970s, PFC-bearing wastes were disposed of in three unlined landfills in Washington County, Minnesota. The resulting co-mingled PFC plumes created a “megaplume” that contaminated over 250 km² of groundwater in four major drinking water aquifers; affecting eight municipal water supply systems and thousands of private wells. Site investigations revealed that karst features, particularly in the Prairie du Chien Group (OPDC), and groundwater-surface water interactions played a critical role in contaminant migration.

Introduction
Perfluorochemicals (PFCs) are a class of fully-fluorinated organic chemicals that have been used to produce a wide range of industrial and commercial products. Their extreme persistence and mobility in the environment and nearly ubiquitous presence in humans and wildlife has raised serious concerns regarding their environmental and health effects (Giesy and Kannan, 2002; Olsen, et al., 2003; ATSDR, 2009). As a result, in 2006, the EPA announced it would seek the phasing out of production and most uses of two PFCs, perfluoro-octane sulfonate (PFOS) and perfluorooctanoic acid (PFOA).

3M Corporation (3M) researched and produced PFCs in Cottage Grove, MN since the 1940s. From the 1940s to 1970s, PFC-bearing wastes were disposed of at the 3M Cottage Grove facility and in three unlined landfills in Washington County, MN (ATSDR, 2008). All of these sites were investigated for industrial solvent contamination in the 1980s and had active groundwater extraction systems operating. In 2003, 3M notified the Minnesota Pollution Control Agency (MPCA) that PFCs had been detected in the drinking and production water at the Cottage Grove plant. The MPCA and Minnesota Department of Health (MDH) began testing Washington County public and private wells for PFCs in 2003. This work eventually delineated three major co-mingled PFC plumes that contaminated over 250 km² of groundwater in four major drinking-water aquifers; affecting eight municipal-water supply systems and thousands of private wells (Yingling, et al., 2014).

Initial groundwater flow modeling predicted more limited, discreet PFC plumes (Barr Engineering, 2005). It was also assumed that a north-south trending groundwater divide that bisects the county would largely limit PFC contamination to the west of the divide. However, further investigation revealed groundwater-surface water interactions and the influence of karst features in a key bedrock aquifer created unanticipated pathways for contaminant migration.

Hydrogeology of Southern Washington County
Southern Washington County is located on the eastern edge of the Twin Cities Basin and is underlain by the lower Paleozoic sedimentary sequence typical of the Hollandale Embayment, covered by 3 – 30 m of unconsolidated glacial drift and alluvium (Figure 1). Reactivation of Proterozoic faults sometime after the Middle Ordovician resulted in large-scale, northeast-southwest trending block faults with vertical displacement of up to 45 m and associated subperpendicular and subparallel jointing (Figure 2; Mossler and Bloomgren, 1990; Mossler and Tipping, 2004).

Four major aquifers provide the majority of drinking water for community and private wells in the area: St. Peter Sandstone (OSTP), Prairie du Chien Group (OPDC), Jordan Sandstone (CJDN), and Tunnel City Group (CTCG; formerly the Franconia). Although separated by lower permeability layers or “leaky” aquitards (as in the case of the St. Lawrence [CSTL], between the CJDN and CTCG), bedrock faults and vertical joints have compromised the integrity of these layers, particularly in the southern- and eastern-most portions of the county. This, in part, accounts for hydraulic communication between the aquifers.
This portion of the OPDC is sometimes referred to as the “high transmissivity zone” (HTZ), as much of the groundwater flow within the OPDC occurs at this horizon. Runkel et al. (2007) reported horizontal hydraulic conductivities up to 0.3 km/day in some domestic wells in the city of Lake Elmo that intersect solution-enlarged bedding-plane fractures in the OPDC. Wheeler (1993) reported dye trace flow velocities up to 10 km/day for the OPDC in southeastern Minnesota. Tipping et al. (2006) reported borehole observations elsewhere in Minnesota of vertical flow upward from the CJDN and downward flow from the upper Shakopee Formation toward the HTZ, with outward flow into the aquifer at that horizon. Upward flow from the CJDN also was documented, under pumping conditions, in monitoring wells at one of the PFC disposal sites in Washington County (Weston Solutions, Inc., 2007).

Tipping, et al. (2006) reported increased aperture size and density of solution-enlarged fractures where the Shakopee Formation is buried less than 60 meters, relative to those in more deeply buried areas. It is also expected that solution-enlarged features (bedding plane partings, systematic and non-systematic fractures, etc.) would be larger and more abundant near the erosional surfaces of the now buried bedrock valleys (Runkel et al., 2003).

PFC Plume Delineation

Nearly 2,000 public, private, and monitoring wells in southern Washington County have been sampled since 2003, providing detailed spatial and temporal delineation of the PFC plumes. The most widespread of the PFCs detected is perfluorobutanoic acid, PFBA (Figure 2). PFBA is one of the most mobile PFCs due to its extremely low soil/water adsorption coefficient value (Kd<0.01). The extent of the PFBA plume far exceeds early modeling predictions of the plumes (Barr Engineering, 2005), including transport across the groundwater divide. This was also observed in the distribution of other PFCs (ATSDR, 2008).

While most of the PFCs mapped generally appear to follow fairly typical concentration trends - highest at the source areas and decreasing with distance downgradient - PFOS in the Oakdale-Lake Elmo area does not (Figure 3). Low concentrations of PFOS are detected in the Washington County Landfill monitoring wells and the PFOS plume dissipates to below detectable levels (<0.05 µg/L) a short distance downgradient. However, further
south and southwest (i.e. downgradient) of the site, “hot spots” of elevated PFOS concentrations (>2 µg/L) in private wells were identified. Also, the plume southwest of these “hot spots” has a distinctive “fingering” pattern (Figure 3). This suggested that other factors besides simple contaminant transport by regional groundwater flow were at work.

**Groundwater-Surface Water Interactions**

Review of the groundwater sampling data revealed that all PFOS detections in Lake Elmo private wells were located south and west of Raleigh Creek (Figure 3). This intermittent stream emerges from a series of wetlands near the 3M-Oakdale Disposal Site, where high levels of PFOS (>45,000 µg/L) have been detected in the water-table aquifer. Sampling of Raleigh Creek in 2006 detected concentrations of PFOS, PFOA, and PFBA up to 8 µg/L near the disposal site and up to 3 µg/L near the “hot spot” in Lake Elmo (ATSDR, 2008). The surface water concentrations correlated closely to PFC levels detected in the nearby private wells completed in the OSTP and OPDC up to 55 m below ground surface.

Between 1988 and 1995, groundwater extracted from the Washington County Landfill as part of the industrial solvent remediation was discharged to a nearby storm sewer. This sewer ultimately discharges to Raleigh Creek, approximately 2 km north of Eagle Point Lake (Figure 4). MDH estimates more than 450 kg PFBA and 35 kg PFOA were discharged to the creek during this time (ATSDR, 2008).

Eagle Point Lake and Lake Elmo are classified as “flow through” lakes, where groundwater recharge and discharge occur in different parts of the lakes (Washington County, 2014). Groundwater discharges to Eagle Point Lake from the north and recharges back to the groundwater near the south and west shoreline. Groundwater also discharges to Lake Elmo from the north and recharges back to the groundwater near the east and southeast shoreline.

Historically, during flood state, Eagle Point Lake overflowed through a series of wetlands to Lake Elmo, on the east side of the groundwater divide. To control flooding of Lake Elmo, a pipeline was constructed from Eagle Point Lake to Horseshoe Lake.

PFCs were detected in the surface water of Eagle Point Lake and Lake Elmo. As with Raleigh Creek, the lake water PFC concentrations correlated closely to groundwater samples from nearby OPDC wells (Figure 3, Table 1). PFCs have also been detected in OPDC wells downgradient (east-southeast) of Horseshoe Lake at concentrations similar to those detected in Lake Elmo (samples GW-4a and SW-4).

The PFC data illustrate the intimate groundwater-surface water interactions which help explain, in part, how PFCs spread further than modeling predicted. For example, PFCs discharge from the groundwater at the Oakdale Disposal Site into Raleigh Creek. Then, as the creek flows toward the city of Lake Elmo, it transitions from a gaining to a losing stream (just east of I-694), acting as a linear source of PFCs recharging to the groundwater and co-mingling with the PFC groundwater plume sourced from the Washington County Landfill (Figure 4). At Eagle Point Lake, PFCs from Raleigh Creek are added to those discharging to the lake from the groundwater. From Eagle Point Lake, some PFCs recharge to the ground-water, resulting in the area of contaminated wells south and southwest of the lake. However, some
of the Eagle Point Lake water (and PFCs) recharge to the groundwater on the east side of the lake and then discharge into Lake Elmo while some is piped directly to Horseshoe Lake. In Lake Elmo, PFCs from Eagle Point Lake combine with PFBA discharging with the groundwater into the north end of the lake. The PFCs then recharge into the groundwater on the east side of the lake. Similarly, PFCs discharge from Horseshoe Lake to the groundwater to the east and southeast.

### The Role of Karst in PFC Migration

Groundwater-surface water interactions alone cannot account for the anomalous distribution of PFCs entirely. For example, PFCs were unexpectedly detected in private wells east of the Washington County Landfill, across the assumed groundwater divide, even though groundwater flow at the landfill is to the south-southwest (Figure 4). Similarly, movement of PFCs through the groundwater between Eagle Point Lake and Lake Elmo is unexpected, given the location of the groundwater divide between these two lakes.

The places where the PFC plume crosses the groundwater divide appear to coincide with buried bedrock valleys, along the walls of which the HTZ within the OPDC subcrops below the Quaternary deposits. As noted above,

### Table 1. Co-located Groundwater (GW) and Surface Water (SW) Samples (in µg/L). Sample locations shown on Figure 3

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>PFOS</th>
<th>PFOA</th>
<th>PFBA</th>
</tr>
</thead>
<tbody>
<tr>
<td>SW-1</td>
<td>5.2</td>
<td>1.9</td>
<td>1.4</td>
</tr>
<tr>
<td>GW-1</td>
<td>3.3</td>
<td>3.0</td>
<td>3.4</td>
</tr>
<tr>
<td>SW-2</td>
<td>0.3</td>
<td>0.3</td>
<td>0.3</td>
</tr>
<tr>
<td>GW-2</td>
<td>0.2</td>
<td>0.4</td>
<td>1.5</td>
</tr>
<tr>
<td>SW-3</td>
<td>0.6</td>
<td>0.4</td>
<td>0.3</td>
</tr>
<tr>
<td>GW-3</td>
<td>0.6</td>
<td>0.5</td>
<td>1.1</td>
</tr>
<tr>
<td>SW-4</td>
<td>0.26</td>
<td>0.08</td>
<td>0.4</td>
</tr>
<tr>
<td>GW-4</td>
<td>0.07</td>
<td>0.06</td>
<td>0.4</td>
</tr>
<tr>
<td>GW-4a</td>
<td>0.058</td>
<td>0.067</td>
<td>0.29</td>
</tr>
</tbody>
</table>

Figure 3. Anomalous distribution of PFOS in the Oakdale-Lake Elmo area. Data shown is a compilation of PFOS results for all aquifers. Where lakes are shaded, color indicates concentrations detected in surface water. Areas without shading indicate no PFC samples have been collected. Numbered, co-located surface and groundwater results are presented in Table 1.
where the Shakopee Formation is buried less than 60 m, the density and aperture of solution-enlarged features increase creating high flow velocity groundwater conduits. Mapping of where the HTZ subcrops below the Quaternary deposits (Olsen, 2008) yields nearly perfect correlation to the areas where the PFC plume crosses the groundwater divide (Figure 5).

As observed in aquifer tests and flow metering in area wells, the high hydraulic conductivities measured in the HTZ create significant potentiometric head differences that can draw water from both overlying and underlying aquifers where they are hydraulically connected by a borehole or fractures. The same phenomenon appears to occur along the bedrock valleys, which provide large scale hydraulic connections between the aquifers above and below the Oneota Formation. This allows for rapid migration of PFCs into the OPDC via the HTZ. In places, the head differences are apparently great enough to locally displace the groundwater divide from its assumed location, allowing the PFCs to appear to cross the divide as they migrate through the HTZ and are then transported east-southeast with the regional groundwater flow. Another explanation may be the strong anisotropy in karst, which can allow for flow directions that differ from the gradient (Tipping, personal communication). This may also account for the apparent “upgradient” migration of PFCs (relative to regional groundwater flow) along some branches of the bedrock valley, as seen immediately east of the Washington County Landfill (Figure 5). Although regional groundwater flow in this area is to the east-southeast, PFCs are detected in wells 600 m north of the bedrock valley.

The southwest-trending finger-like lobes of the PFC plumes (Figures 3, 4, and 5) may also be an indication of

Figure 4. Anomalous PFOA distribution illustrates complex PFC transport pathways. Data shown is a compilation of PFOA results for all aquifers. Where lakes are shaded, color indicates concentrations detected in surface water. Areas without shading indicate no PFC samples have been collected.
karst influence on contaminant transport. The “fingers” are oriented roughly 240 degrees. This is consistent with lineament orientations reported by Mossler and Tipping (2004). It seems reasonable to suspect that higher flow rates in solution-enlarged, subparallel joint sets in this area provide preferential pathways for PFC migration.

Summary

The PFC investigations in southern Washington County illustrate the potential for highly complex contaminant migration pathways in karsted aquifers, particularly for highly persistent chemicals. While groundwater modeling provided initial direction for groundwater sampling, a thorough understanding of the underlying bedrock features, groundwater-surface water interactions, and the distribution and orientation of karst features, was critical to accurately delineating the PFC plumes. This had significant implications for public health, as many of the drinking water wells located in the PFC “hot spots” - that were not predicted by modeling - exceeded Minnesota drinking water standards and required treatment to remove the PFCs.

Disclaimer

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References


Figure 5. Correlation of areas where high transmissivity zone (HTZ) subcrops and PFCs cross groundwater divide. The white arrows indicate areas where HTZ subcrop (shown in yellow hatching) correlates with PFC migration across the groundwater divide. Note that PFCs are detected at significant distance “upgradient” of the buried valley. Data is a compilation of PFOA results for all aquifers.


ATSDR. 2008. Public Health Assessment:


Abstract
Karst groundwater contamination presents great challenges for efficient monitoring because of rapid, discrete transport and the diversity of contaminants. Here a low cost approach is described and applied to Hidden River Cave, Kentucky, where a long history of contamination has been experienced. Local knowledge was acquired through informal interviews and coupled with observations of contaminant residues, faunal distributions and fluorescence spectra in the cave. The resulting patterns were interpreted using Google Earth and Street View to identify specific contaminant sources in the affected sub-catchment of the cave. Despite success in matching contaminant sources with the contamination history and pattern, the informal nature of the investigation renders it unacceptable as the basis for any intervention. But such low cost studies will be needed for the majority of contamination occurrences where financial resources are very limited. A radical revision of our adversarial approach to environmental management will be required for such a change to occur.

Introduction
Alternative monitoring technologies
Water contamination is a major karst hazard that presents great challenges for efficient monitoring (Gutiérrez et al., 2014). Contamination can arise from a wide array of chemicals that may react, be released, sequestered, or transported through a variety of processes. This complexity must be resolved to comprehend, prevent and remediate the contamination. Yet a comprehensive understanding of water contamination requires a logistically impractical characterization of the composition, timing, and spatial distribution of contaminants. Agencies frequently focus resources on the detailed composition of contamination, rendering locational and temporal characteristics as lower priority. This analytical bias reflects legislation and a strong technical emphasis, in comparison with considerations of sampling. Consider the widely used binary contingency framework to categorize spatial (point and diffuse source) and temporal (acute and chronic release) characteristics. This is convenient, but provides little substantial insight in dealing with real world contamination. The result seems to be an expensive emphasis on handling and analytical protocol rather than a consideration of appropriate spatial and temporal sampling strategies. Here, an informal screening approach is taken to environmental monitoring, using a variety of qualitative and quantitative tools for a very low cost environmental appraisal.

Karst aquifers have presented peculiar difficulties for monitoring because rapid longitudinal transport makes detection and mapping of transient contamination challenging using monitoring wells. Yet complementary transverse dispersion of a much lower order may yield results consistent with porous medium models, confounding ready interpretation. As a result, practical groundwater models are yet to be widely accepted in karst investigations, or else porous medium models are used to over-ride a karst conceptual model (Worthington & Smart, 2012).

Fortunately, the drainage conduits in karst aquifers allow efficient groundwater tracing to be applied to resolve contaminant trajectories and catchment areas. Furthermore, conduits generally function as dendritic networks, at least under base flow conditions. This allows the utilization of springs or trunk conduits as monitoring sites that integrate the runoff from the upstream drainage area, much the same as surface water monitoring sites are used to characterize a catchment area (Quinlan et al., 1991). The limitation of this approach is that a contaminant source cannot be precisely located within the catchment area other than through sequential tracing and tracking—an expensive proposition.

Identifying the source of karst contaminants is feasible, but requires substantial resources if it is to be successful in litigation. This cost and uncertainty prevents resolution of many contamination problems, even egregious events. The proposition explored here integrates a wide
spectrum of inexpensive monitoring strategies: the use of the mapped cave network and snapshot field observation and sampling, coupled with open access digital mapping (Google Earth and Street View) to tie contamination events to a surface land use. It is also assumed that cave biota (cave crayfish- *Orconectes pellucidus*) populations and kills will reflect recent acute contamination (Chapman, 1993), although the sensitivity, response and recovery dynamics are unknown. Furthermore, an emphasis is placed on narrative sources in the design of field sampling and the interpretation of results. The project was undertaken through annual college field excursions of a few hours duration from 2008-2014.

**Hidden River Cave**

Hidden River Cave lies under the town of Horse Cave, Hart Co. Kentucky, and is accessed through a collapsed sinkhole managed by the American Cave Museum. The cave consists of a dendritic network of canyons and collapsed domes (Figure 1). The main stream (East River) rises from rubble at the bottom of the entrance collapse and drains an area of ~150 km$^2$ based on dye tracing (Quinlan & Rowe, 1977). A much smaller stream (South River) drains ~8km$^2$, and has been explored up three main tributaries: Kneebuster, Lover’s Lane, and Wheet River. The latter is the dominant source of the South River, but disappears under breakdown to re-emerge as the South River. Additional small tributaries and seeps enter this section of the cave and are presumed to link to numerous surface sinkholes, although none of these are known entrances. In addition, a number of poorly drained depressions in Horse Cave have been augmented by drilled drainage wells; casings of two of these are encountered in the cave passage.

Hidden River Cave has a rich history of exploitation as a tourist cave and hydro-power site, but also as a dump site (Veni et al., 2001). Raw sewage and untreated industrial waste (dairy and metal plating effluent) were discharged into a sinkhole at the extremity of the Lover’s Lane tributary of the South River. These contaminants resulted in eutrophic, anoxic conditions, annihilating cave life and rendering the cave and the City of Horse cave foul and repellent (Quinlan & Rowe, 1977). Fortunately, this degradation was publicized and remedied by the establishment of a regional sanitary sewer system in 1983, following which the cave ecology recovered and the cave could be reopened for tours.

*Figure 1.* Line plot of Hidden River Cave overlain on Google Earth (March 2014) image. Main features in the cave and some surface features.
Since the millennium, there has been substantial industrial development in the catchment area of the South River, contaminants have been encountered in the cave and periodic annihilation of the cave crayfish has occurred. But in the absence of any formal monitoring and recording, such episodes are generally narrative rather than scientific. Lack of funds and expertise has limited the documentation and action on these contamination events. A routine cave tour by a senior college class in 2008 initiated studies of the contamination that have continued annually to the present day.

Monitoring
From 2008, each year in September for one afternoon, a small student group has explored the South River -Wheet River section of Hidden River Cave. Informal discussions with the staff provided an idea of the contemporary crayfish numbers and any kills, contamination events perceived in the cave and possible surface origins and triggers. The number of living and dead cave crayfish occupying pools through the cave have been noted, along with descriptions of any detrital material.

Analysis of water samples for putative transient contaminants is prohibitively expensive because not only is the contaminant species unknown, but there may be little indication of which samples merit analysis. The notion of screening using surrogate water quality variables is long-established in karst hydrology. For example, electrical conductivity has been widely used as an inexpensive, continuous surrogate for total dissolved solids, ionic concentrations and even chemical equilibria (Meus et al., 2014). Electrical conductivity is readily measured, but is a relatively poor surrogate for groundwater contamination. A continuous record has been obtained from the East River from fall 2013-2014, but the complementary logger in the South River has yet to be recovered, so these results are not presented here.

Dye tracing literature has long dealt with “background fluorescence” as a nuisance (Smart & Karunaratte 2002), but a wide range of environmental contaminants have expression in fluorescence. For example, fuels, raw sewage, road runoff and overland flow can all be revealed in the fluorescence spectrum, as can leaking treated water (Hartell et al., 2007; Mudarra, Andreo, & Baker, 2010; Quiers et al., 2014.). However, none of these contaminants has a direct quantifiable fluorescence signature, and this indicates an empirical rather than analytical approach. In other words, there is no formal calibration for contaminants, although confidence can be gained using samples of known provenance or composition.

Water samples have been collected from cave streams, tributaries, pools and drips, and analyzed using a UV-VIS Spectrofluorometer to obtain synchronous scans at Δλ=20 and 90 nm with 5nm slit settings using a tap water blank and ambient fluorescent lighting for calibration. The raw scans were smoothed, normalized and standardized using the collective median for all samples at a given wavelength, and the median of a wavelength range for each sample, respectively. The resulting spectra show anomalies with respect to the sample ensemble for each year.

An inherent component of the contaminant narrative was putative contaminant sources. These sites were explored and documented, as time permitted, on the surface using a custom ArcGIS field collector application and through Google Earth and Street View.

Monitoring results
Narrative
The initial modern era contamination event was by plastic injection molding stock pellets (“nurdles”), likely originating from Dart Containers, a packaging company with a plant adjacent to the former sewage treatment plant and a warehouse in the upper Wheet River area. A bark mulch contamination event was described in the Wheet River, with a postulated origin outside the catchment area. A variety of contaminants (liquid cement, tar, oil, soap, fuel) was described at the drainage well casing. A strong odor of chlorine was commonly encountered in the Waterfall Room, where a perennial cascade enters from the cave roof. Crayfish kills were not clearly linked to these events, but were interpreted to originate from the headwaters of the Wheet River where a condiments factory has a waste treatment facility.

Residues in the cave
Plastic nurdles were encountered embedded in mud on the cave wall, mostly downstream of the Lover’s Lane tributary suggesting an origin from that tributary. Bark mulch residues were restricted to the main Wheet River. Crayfish exoskeleton fragments were found with some of the bark, having similar sediment transport properties. Soap, concrete, LNAPL and brown staining were observed at the drainage well. Chlorine odor was detected at the waterfall room. All these observations confirm the narrative descriptions. In addition, a few hundred ml DNAPL blob (Loop & White, 2001) was observed in 2008 at the end of Kneebuster Crawl. The DNAPL appeared to have drained from an erosional scar in an adjacent sediment bank. A dead crayfish lying on the DNAPL was being consumed by a live crayfish. A small trickle in the wall of the cave downstream of the entrance...
was coated in a filamentous slime, presumably bacterial in composition.

**Cave Crayfish etc.**

The crayfish population varied considerably from year to year, but part of this variability will reflect changes in the diligence and skill of observation, illumination power and stream turbidity (The Wheet River is particularly muddy and permits a one-time survey moving upstream). A professional biotic survey has been undertaken in recent years (in prep.) generating superior data indicating under-recording in our surveys, but general consistency of distribution patterns.

The general trend has been for a gradual increase in Wheet River crayfish population since 2008, although in part this may reflect improved illumination and skill. Very few were encountered in Wheet River in 2008, with the exception of five or six individuals occupying a tributary stream. In 2009, this small cluster had expanded out into the main river, largely downstream, and were observed feeding in detritus, including crayfish fragments. In contrast, there has been a gradual decline in counts in the Kneebuster tributary.

Crayfish were occasionally encountered in ad hoc locations such as small inlets and isolated pools up to 5 m above stream level and well above flood lines. A small number of surface crayfish (*Cambarus tenebrosus*) were also encountered in the Wheet River. The implications of this are not known.

Blind cave fish (*Typhlichthys subterraneus*) were observed rarely in the Wheet River since 2013. Isopods were seldom evident, although numerous in an inlet tube downstream of the cave entrance. This suggests the cave ecosystem is still sub-optimum, either due to residual contamination or ongoing events.

The patterns observed indicate that cave crayfish are responsive to presumed contamination. They are surprisingly mobile and therefore able to migrate to and from refugia, such as clean tributaries. They are not restricted to the saturated zone. In addition to avoiding contaminated areas, the crayfish are drawn to rich food sources, including their own species. Depending on the contaminant, this may perpetuate mortality.

**Fluorescence monitoring**

Replicate analyses, replicate samples and longitudinal samples in particular stream reaches showed very consistent spectra. Care had to be taken to avoid sediment contamination of samples as even small inclusions resulted in anomalous spectra. Disposable plastic 5ml pipettes proved to be the most efficient way of collecting a clean sample, avoiding inconsistency from collecting LNAPLs by dip sampling and allowed sampling from quite shallow pools. The marked fluorescence anomalies arising from floating product and sediment elution suggest quite complicated migration and sequestration of contaminants (Vesper & White, 2004; Vesper & White, 2006). However, the results are difficult to replicate and are a source of variance, extraneous to the neutral density, aqueous phase population drawn on for statistical analysis.

Normalized fluorescence spectra (e.g. Figure 2) were analyzed based on broad spectrum ranking, as well as particular peaks indicative of dyes, sewage and fuels. The broad spectrum fluorescence shows distinctive clustering for particular streams, depending on prevailing runoff conditions, so that the East River may or may not have higher levels than the South River in a given year. Figure 3 shows the ranked 350-450 nm normalized fluorescence at $\Delta\lambda=90$nm, highlighting the clustering of broad fluorescence and dissolved organic matter in particular sub-catchments (Birdwell & Engel, 2010). (The synchronous scan offset and exact waveband does not materially affect this outcome, although year-to-year order is highly variable).

More systematically, the upstream Wheet River has lower contamination than the downstream South River, although it is the dominant water source. An inconsistent variety of short wavelength emitters (typically fuels and lubricants) have been detected in seeps and drips in the breakdown section between the Wheet River and South River. Particularly marked concentrations appear in drips adjacent to the injection well casing and a large stalagmite. There is clearly contamination of this sector of the cave, although it does not exhibit consistent spectral form, indicating either a variety of contaminants, or differential transport and degradation (e.g. volatilization) of the various components through different tracer routes.

Small perennial cascades enter the cave at the Waterfall Room and adjacent Board Room, often with a distinctive chlorine odor that may indicate treated drinking water, possibly accounting for the perennial flow even after sustained droughts. Spectra from the Waterfall show very low broad fluorescence, a characteristic of mains water. Perplexingly, the waterfall room spectrum shows a distinctive 515nm peak of comparable magnitude every year from 2008-2014. This peak is characteristic of the common tracer dye uranine (sodium fluorescein). The consistent level over many years suggests that this contaminant is not a tracer dye that would be expected to exhibit a declining concentration over time. Its as-
The small slimy trickle downstream of the entrance shows a distinctive and consistent 435 nm peak, typical of optical brighteners in raw domestic sewage (Hartell et al., 2007).

**Summary and putative sources**

The patterns of crayfish numbers and contaminants are poorly correlated, or may be coincidental. For example, the bark mulch does not appear to have been sufficient to eradicate crayfish from the Wheet River, although it does clearly indicate a garden center as a putative source. The plastic beads are almost certainly from a plastics production facility. The general contamination in the breakdown area near the drainage wells suggests a construction or trucking operation. Uranine can be tentatively derived from a garage or car wash. Chlorination and low fluorescence indicates a mains water leak or disposal of consumed water by general runoff. Raw sewage indicates a poorly linked sanitary sewer. The DNAPLs observed in 2008 have not been encountered since. It is assumed that this contaminant is a legacy of the era of wholesale contamination that had been sequestered in cave sediments. The variety of fluorescent signatures associated with mains water suggests an industrial source, possibly a garage or car wash where uranine stained antifreeze is being disposed. However, the invariant concentration suggests a steady release of tracer for which there is no obvious release mechanism.
arising from sediment elution suggests long-term sequestration of contaminants in the sediments (Vesper & White, 2004). The subsequent task is to explore surface land use looking for the sources arising from narrative accounts and those indicated by the contaminant pattern in the cave.

**Surface Land-use**
Targeted field traverses have been undertaken from 2011-present, but overt site investigation tends to raise concern and even alarm in residents. Google Earth and Street View reconnaissance is much less confrontational, provides some historical perspective and allows insight into inaccessible or concealed areas, or activities not occurring at the time of field reconnaissance. A high-grade cave survey was geo-registered in Google Earth allowing for a direct comparison of surface land use, cave streams, and subsurface contamination.

**Google Earth**
High quality satellite imagery allows mapping of major industrial installations and waste management facilities. It is not always easy to categorically identify the industrial activity, and certainly Google Earth is not suited to detect acute contamination events given the infrequent imagery. Furthermore, the topographic rendering in Google Earth is not particularly accurate, and should only loosely be employed to define sinkhole catchments.

The regional sewage treatment facility (the origin of former contamination) is clearly visible. Immediately adjacent is a set of stock supply hoppers for Dart Container. Knowing that the plastic beads are typically stored in such hoppers, and that the adjacent sinkhole connects directly to the cave, it can be hypothesized that a spill at the site resulted in transport of plastic beads into the Lover’s Lane tributary of the South River. (No stream has been detected in or emerging from the Lover’s Lane tributary, but a link is inferred from past pollution history). A nearby Garden Center showed mulch bins adjacent to sinkholes near the Wheeth River headwaters.

The locale of the condiments factory can be seen first developing in 2003, with a major waste treatment installation in 2006. Some major reconstruction and erosion control is documented in 2010, and a new fermenta-
tion tank and waste lagoon appears in 2014. (Figure 4). Other land use is more tentative and required review in Street View for confirmation.

**Street View**

Even minor backstreets of Horse Cave have been faithfully recorded in Street View, although private industrial areas are not mapped. Efforts were concentrated on the contaminated area around the drainage well, assuming a fairly direct passage of surface contamination into the underlying cave, or its inferred tributaries.

The garden center postulated to have generated the mulch contamination events indicates runoff may have been directed towards a number of sinkholes in the headwaters of the Wheet River. Numerous small and larger industrial installations occur over the contaminated section of cave. A car wash was inspected showing runoff entering directly into a storm drain in the major depression above the cave. (Although topography is not necessarily a guide to storm water runoff in urban areas, there does not seem to be much storm water management in Horse Cave, other than draining ponded water into the subsurface through injection wells). There is also a small garage, a rather disordered recycling center, and a fuel transfer depot/ gas station a little further away.

Narrative descriptions and field inspection (see below) identified a primary candidate drainage well adjacent to a carpentry shop, an unlikely source of contamination. However, immediately to the west of the injection well is a pre-mix concrete depot providing a source for the diverse contaminants encountered in the cave (fuel, lubricants, cement; Figure 5). Street View reveals the loading areas being hosed down with runoff draining directly to the drainage well, despite an apparent settlement tank on the site. Street View imagery shows a large dump truck loading at the food processing waste facility, suggesting solids removal.

**Field reconnaissance**

Field investigations were directed by hypotheses raised using Google Earth and Street View, but constrained by concerns over trespass and possible alarm. For example, it was not possible to look for residues of plastic beads around the postulated sink point because it was not accessible. However, at the garden center traces of mulch clearly indicated transport into a sinkhole south east of

![Figure 5](image_url). Google Earth Image March 2014. Premix concrete plant in Horse Cave. Runoff from the site follows the purple line that terminates at an open drainage well. The orange line represented the underlying cave passage, suggesting ~25 m mis-registration.
the storage bin. Any spills from the equipment maintenance area would also enter this sinkhole, feeding directly into the Wheet River.

A number of drainage wells and sink points were located in depressions over the cave, but uncertainties remain as to which feed directly into the cave. The drainage well receiving runoff from the pre-mix concrete facility showed contaminated runoff from pressure washing on the forecourt, off the site, through swales and into the drainage well. The on-site settling tank was stagnant and not utilized.

The condiment factory had extensively re-engineered the surroundings to direct storm water into sinkholes. In 2011, a sinkhole immediately north of the waste management facility was undergoing extensive modification. A collapse feature had developed in the floor of the sinkhole. By 2014 a large retention pond (~0.2 ha) had been created at this site (Figure 5).

However an outflow pipe limited the retained depth to a maximum of ~1m, with the outflow being directed to a major sinkhole. Some erosion of the floor makes its ultimate integrity uncertain. There was no evidence of waste being dumped into the lagoon, suggesting that it may be an emergency retention pond, to contain accidental spills or excess liquids. It is not clear what the ultimate fate of the treated waste might be. Narrative accounts describe tankers hauling waste to a landfill site. The fate of surplus liquids is not known.

**Contaminants & Land Use Practices**

The suite of contaminants inferred from narratives and monitoring can be surprisingly well-tracked to normal surface land use practices and accidents that arise through day-to-day operations in an industrial park. This is surprising as the contaminant events tracked are all acute point source events—least likely to be captured by annual snapshot surveys. Only those events leaving clear residues, or surface evidence, have been identified. It is likely that the cave is actually exposed to many more events that elude detection. Such events may arise from contamination events resulting from the mobilization of contaminants sequestered in cave sediments.

Most of the acute contaminants inferred could be removed by the adoption of best management practices such as runoff capture and treatment, for instance oil & grit separation. The condiment waste source appears to have already made moves in this direction.

The chronic contamination by sewage and uranine dye have not been sourced, although in principle they should be more readily tracked. The perennial uranine source is not readily interpreted, so the release process remains a mystery. The sewage detected indicates a source that is so widespread, that it is impossible to identify a precise origin.

The success in matching contaminant sources to the immediately subjacent cave suggests that routing through the unsaturated zone is vertical through the respective sinkhole or drainage well into the cave network. The presence of coarse particulates indicates discrete openings of at least 10cm aperture. The contaminants entering the cave from the injection well are encountered outside the well casing, suggesting that the installation is dysfunctional.

There has been no follow-up on any of these contamination sources and events. This reflects the challenges of environmental protection against a background of industrial decline and stressed resources.

None of the evidence presented here would be acceptable if taken to litigation, as it lacks the formal scientific rigor required. It would be relatively straightforward to undertake local sink-cave stream traces to confirm linkages. But the evidence suggests that would just confirm the obvious. Tracking the individual contaminant sources using site-specific signatures would be prohibitively expensive, and probably ineffective for acute events. The apparent sequestration of contaminants in sediment may be a route around the latter barrier, but much more investigation would be required to work out the processes involved.

**Conclusions**

The contaminant history described and inferred here is likely representative (or even under-representative) of any developed landscape, and likely repeated in karst regions around the world. Spills, releases and disposal practices are so routine that they attract little attention.

This attitude has not been helped by the geographical partitioning of source and wastewater catchments, so that water consumption is independent of waste disposal practices. Only the most egregious contamination by large, financially robust entities warrants the necessary expenditure of efforts in monitoring and litigation. It is not clear that egregious contamination is necessarily the optimum target if environmental conservation is a priority. Persistent minor contamination may be more harmful.

Here, it has been shown that a fairly comprehensive, yet legally inadmissible investigation has identified a num-
ber of contaminant sources that could be readily ameliorated, likely for far less than the cost of litigation. The foundation of this work lies in the social sciences-utilizing informal interview techniques to establish a focus for subsequent field monitoring. The field monitoring is also open ended, focusing on mapping rather than rigorous compositional analysis. The techniques are largely qualitative and yet have led to a valuable characterization of contamination at little cost. It is perhaps time that we managed our environment more through a comprehensive understanding rather than an adversarial review scrambling for scientific superiority.

More fundamentally, this study confirms that karst environments are vulnerable to routine contamination, and the damage is likely rapid and profound. However, despite possible sequestration of contaminants in sediments, the dendritic structure of karst drainage means rapid flushing with relatively little transverse dispersion. The drainage structure of karst aquifers therefore tends to preserve refugia. Furthermore cave biota are certainly preserved in the refugia and may even actively migrate into them when threatened by contamination.

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SPATIOTEMPORAL RESPONSE OF CVOC CONTAMINATION AND REMEDIAL ACTIONS IN EOGENETIC KARST AQUIFERS

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Abstract
The northern karst region of Puerto Rico has a long and extensive history of toxic spills, chemical waste, and industrial solvent release into the subsurface. High potential for exposure in the region has prompted aggressive remediation measures, which have extended for over 40 years. Of particular concern is contamination with chlorinated volatile organic compounds (CVOCs) because of their ubiquitous presence and potential health impacts. This work evaluates historical groundwater quality data to assess the spatiotemporal distribution of CVOC contamination in the karst aquifer system of northern Puerto Rico, and its response to remedial action in two superfund sites contaminated with CVOCs. Historical data collected from different information sources with different monitoring objectives is evaluated spatially and temporally using Geographic Information System (GIS) and statistical analysis. The analysis shows a significant extent of contamination that comes from multiple sources and spreads beyond the demarked sources of pollution. CVOCs are detected in 65% of all samples and 78% of all sampled wells. Groundwater shows continued level of contamination over long periods of time, demonstrating a strong capacity of the karst groundwater system to store and slowly release contamination. Trichloroethene and Tetrachloroethene are the most frequently found, although other CVOCs (e.g., Trichloromethane, Dichloromethane, Carbon Tetrachloride) are detected as well. The spatial and temporal distributions of CVOCs seem to be highly dependent upon the monitoring scheme and objectives, indicating that the data does not adequately capture the contamination plumes. Targeted remedial action using pump and treat (air stripping) and soil vapor extraction in two superfund sites has reduced concentrations over time, but the spatial and temporal extent of the contamination reflects inability to completely capture the heterogeneous plumes.

Introduction and Background
The northern karst region (Figure 1) contains two of the most extensive and productive freshwater aquifers in Puerto Rico (Lugo et al., 2001). These aquifers promote industrial, agricultural, and population growth (PRD-NER, 2008), and contribute to the ecological integrity of the rivers, springs, wetlands, and estuaries (Padilla et al., 2011).

The northern karst aquifer system comprises three major hydrogeological units (Figure 1): the upper aquifer, which is mainly composed of the Aymamon and Aguada Limestone Members; the lower aquifer, which is formed by the Lares and Montebello Limestone Members; and a confining unit that separates the upper and lower aquifer, which is comprised by the Cibao formation. (Renken et al., 2002). The upper aquifer is unconfined and also linked to the surface throughout most of its outcrop area. The lower aquifer is confined toward the coastal zone and outcrops to the south of the shallow aquifer, where it is recharged. The outcrop extents are much more vulnerable to contamination due to direct interaction with the surface.

The karst aquifers of northern Puerto Rico have developed in carbonate rocks that have not undergone deep burial and are under active meteoric diagenesis, thus considered eogenetic aquifiers (Vacher and Mylroie, 2002). In addition to well-developed conduit networks, these aquifiers have high primary porosity and permeability and significant flow components through the rock matrix. As a result, both conduit and matrix flow contribute flow in these aquifers. These flow characteristics allow for extensive water production from the aquifiers and make the systems highly vulnerable to pollution (Göppert and Goldscheider, 2008). The highly heterogeneous distribution of flow characteristics in karst systems also imparts an enormous capacity to store and convey contaminants from sources to potential expo-
sures zones. Consequently, karst aquifers may serve as an important route for long-term contaminant exposure to humans and wildlife.

The northern karst region of Puerto Rico has had a long and extensive history of toxic spills, chemical waste, and industrial solvent release into the subsurface (Padilla et al., 2011; Hunter and Arbona, 1995; Zack et al., 1987).

Liquid-waste was also injected into the confined aquifer system prior to the 1970s (Zack et al., 1987). After 40 years of high potential exposure and remediation measures, pollution persists. Since 1983, twelve superfund sites, which comprise 55% of all superfund sites on the island and 13 Corrective Action (RCRA CA) sites have been included in the study area (Figure 1). Several types of contaminants have been detected at superfund sites in Puerto Rico, including: chlorinated volatile organic compounds (CVOCs), pesticides, heavy metals, and contaminants of emerging interest, such as pharmaceuticals and phthalates. Of particular concern are the CVOCs due to their ubiquitous presence in the environment, potential for exposure, and health impacts. CVOCs are present in 58% of superfund sites in the study area, and include: Trichloroethylene (TCE), Tetrachloroethylene (PCE), Chloroform (Trichloromethane, TCM), Carbon tetrachloride ($\text{CCl}_4$), Methylene Chloride (DCM), 1,1-Dichloroethane (1,1-DCA), 1,1-Dichloroethylene (1,1-DCE), and 1,2-Cis-dichloroethylene (cis 1,2-DCE). These pollutants are commonly used as industrial solvents, degreasers, and paint removers; and have been known to cause cancer and reproductive problems (ATSDR, 2011).

This work evaluates historical groundwater quality data to assess the spatiotemporal distribution of CVOC contamination in the karst aquifer system of northern Puerto Rico (Figure 1) and its response to remedial action in two superfund sites: The Upjohn Company (UJ) and Vega Alta Public Supply Wells (VA-PSWs). Historical data collected from different information sources was evaluated spatially and temporally using GIS and statistical analysis.

In 1982, the Upjohn Company reported a 57.92 m$^3$ accidental spill of $\text{CCl}_4$ from an underground storage tank (USEPA, 1984). Shortly after the incident, the company installed monitoring wells in the surroundings, established recovery well for $\text{CCl}_4$ extraction, and provided an alternative water supply to the population. The UJ facility was listed in the Superfund sites in 1984. In 1985, the tank farm area was covered with fiber glass-reinforced concrete to contain the contamination and prevent the infiltration into the subsurface. A total of 45.42 m$^3$ (78.4% of total reported volume) of $\text{CCl}_4$ had been removed by

![Figure 1. Hydrogeology of Puerto Rico and associated superfund, RCRA CA, and landfill sites (Irizarry, 2014; Torres, 2013).](image-url)
1985 using 19 vacuum extraction wells in the soil and a groundwater extraction well installed down gradient of the spill area (USEPA, 1988). The company ceased CCl₄ use and applied a pump and treat technology with air stripping water treatment (1988). Recent 5-year evaluations and review reports suggest that potential human exposure is under control (USEPA, 2014).

In 1983, the United States Geological Survey (USGS) reported TCE and PCE contamination (Guzmán and Quiñones, 1984) in the Puerto Rico Aqueduct and Sewer Authority (PRASA) wells of the Vega Alta region. Although the time of contaminant release is unknown, it is suspected that contamination had been present for a long time. The VA-PSWs site was listed in 1984 in the Superfund program, and an air stripper was immediately installed in one of the PRASA wells, which operated only until 1985 due to technical problems. In 1987, EPA selected the following actions: closure of private wells within the contaminated area; investigation of contamination sources; water softening pretreatment of PRASA wells; and installation and operation of 4 air strippers. The treated effluent was to be discharged to the PRASA distribution system for public use and to Honda creek. The groundwater treatment started in 1994, but due to changes in pumping conditions the plume of the contamination was no longer captured by the treatment wells and the EPA required the installation of extraction wells for treatment in a new location. Remedial actions, including pump and treat and soil vapor extraction (SVE) in the source, began operations in 2002 (almost 20 years after pollution was reported). Since December 2002, approximately 75,708 m³ of water has been treated each month (USEPA, 2014).

**Data Collection and Analysis**

Historical and spatial groundwater quality data were compiled to assess the CVOC contamination in the karst aquifers of northern Puerto Rico. The study area (Figure 1) extended from the municipalities of Toa Baja and Toa Alta on the east to the Municipality of Arecibo on the west. Water quality data included temporal CVOC concentrations spanning from 1982 through 2013 in the study area. Historical data was collected from different information sources with different scope and monitoring objectives. Water quality, site information, and site location were collected from: Steele (2011), U.S. Geological Survey data base (USGS, 2008) and reports (Guzman, et al., 1986; Guzmán and Quiñones, 1984; Guzmán and Quiñones 1985; Sepulveda, 1999; Conde and Rodriguez, 1997), EPA data base (USEPA, 2008), Caribbean Environmental Protection Agency (CEPA) local office, Puerto Rico Environmental Quality Board (PREQB) (PREQB, 2011), University of Puerto Rico at Mayaguez (UPRM), and Puerto Rico Department of Health (PRDoH). Data from EPA included water quality results from superfunds and RCRA CA sites in the study area (Figure 1). The PRDoH data included water quality for PRASA and NON-PRASA drinking water systems. PRASA data included water quality for point water sources in the distribution system, and was collected quarterly for compliance with drinking water regulations. NON-PRASA drinking water systems have been defined as those that are not supplied by PRASA and contain more than 25 people and 5 connections; most of this data came from industries. Data from the PREQB included water quality measurements from their monitoring stations (Steele, 2011), but they only had a few monitoring wells within the study area and the records were for a brief period of time. The PREQB, superfund division provided some special monitoring events of Scorpio Recycling and limited amount of data from Pesticide Warehouse I. Since 2011, the UPRM has been collecting current groundwater samples from 17 wells in the study area twice a year. The samples have been analyzed for target contaminants and the results have been entered into a database. Once collected, data is classified and categorized into data characteristics representing categories and indices used to describe historical distribution of contaminants in the area of study. Characteristic parameters include: aquifer system (lower, upper), contaminant presence (or absence) and concentration ranges, contaminant types (e.g., chlorinated hydrocarbons, phthalates), detection limits, and information source (e.g., UPRM, USGS, EPA, Reports).

Spatial analysis was performed using GIS technologies (ArcGIS 10.1; ESRI 2012). Spatial distributions of CVOCs were analyzed as average detected concentrations over a period of time for the unconfined karst aquifer system of northern Puerto Rico. Although contamination has been detected in the confined aquifer, this work focuses on the unconfined system because of its direct connection to the surface. Average concentration distributions were developed for individual species and total CVOCs, defined herein as the sum of TCE, PCE, CCl₄, TCM, and DCM average concentrations. These species are among the five major CVOCs found in the northern karst region. Isotropic interpolation of concentrations using the Inverse Distance Weighted (IDW) method produced raster spatial distributions of the contaminants. The interpolation method did not, however, take into account groundwater flow direction. Basic statistical analysis was performed to obtain contaminant detection frequencies for groundwater samples and groundwater sites in reference to the total number. Groundwater sites refer to sampling sites (wells/springs). Because many groundwater sites have multiple temporal measure-
ments, the number of groundwater samples is greater than the number of sites. Differences in their basic statistical characteristics reflect on the detection variability within each site.

**Results and Discussion**

Analysis of the data indicates widespread presence of CVOC contamination in the study area (Figure 2) during the 31-year period between 1982 and 2013 (Figure 3). Temporal distributions data (Figure 3a) shows a high percentage of detection over the period of analysis, with 90% of samples surpassing 50% detection on a yearly basis. Similar temporal trends are observed for detection in groundwater sites, except with higher detection frequencies than groundwater samples. Over the entire period of analysis, CVOCs were detected in 65% of all groundwater samples (Figure 3. Total numbers of CVOCs samples and detections per year (a) (Irizarry, 2014) and overall percent detection of CVOC by number of samples (b) and wells (c) for the period of analysis (b) and 78% of all sampled groundwater sites (Figure 3c). Lower detection frequencies of groundwater samples over groundwater sites reflect temporal detection variability at sampling sites and indicate that some sites had both detection and no-detection of contaminants. This behavior is indicative of dynamic fate and transport processes that affect the rate at which contaminants reach potential exposures points.

The most prevalent CVOCs in the study area is TCE (52.5%), followed by PCE (39.0%), 1,1-DCE (30.4%), cis1,2-DCE (30.2%), CCl$_4$ (26%), TCM (22%), 1,1-DCA (15.3%), and DCM (9%). The detection distribution among the different CVOCs is generally associated with major sources of contamination, although data analysis suggests other potential sources of CVOC contamination. Some CVOCs may also be formed by degradation (e.g., DCE, TCM, DCM) or disinfection processes (e.g., disinfection-by-products, DBPs, such as TCM). The highest CVOC concentration is for CCl$_4$, followed by TCE, 1,1-DCE, TCM, DCM, and PCE. TCE shows higher number and percentage of samples above MCL, followed by CCl$_4$, PCE, 1,1-DCE and DCM.

**CVOC Spatiotemporal Distribution**

Spatial distributions of TCE, CCl$_4$, TCM, and total CVOCs over the entire period of analysis (1982-2013) show widespread contamination of CVOCs in the study area that extend beyond known sources of contamination (Figure 2). Although generally detected over the entire area of study, high percent detection and concentrations of TCE (Figure 2a) are associated with the VA-PWS superfund site, whereas those for CCl$_4$ (Figure 2b) are associated with the UJ site. Presence of CVOCs in other areas indicates other potential sources of CVOC contamination, including the formation of degradation and disinfection by-products. The spatial concentration
distribution of TCM (Figure 2c) shows that this contaminant is found widely distributed throughout the study area. The highest TCM concentrations are found near the UJ site and near the Naval Security superfund site in the Toa Baja area. These high concentrations are most likely associated with the CCl₄ contamination, either as an additional source or as a degradation by-product of CCl₄. The widespread presence of TCM may also be attributed to other potential sources of contamination, as well as its formation as a disinfection by-product. Spatial concentration distributions of total CVOCs (Figure 2d) show widespread contamination of CVOCs over the study area.

Temporal concentration distributions of TCE in a well near the VA-PSW superfund site (Figure 4) and CCl₄ in a well near the UJ site (Figure 5) show an overall decreasing trend. The decrease is mostly associated with active remediation activities of the major superfund sites, but is also due to degradation processes. Degradation of TCE and CCl₄ is reflected in the increasing concentrations of their degradation by-products, DCE (Figure 4) and TCM (Figure 2c) near the VA PWS and UJ superfund sites, respectively. It is important to note that, even under active remediation, concentrations have not reached regulatory levels and a significant number of samples still have concentrations above the maximum contaminant level (MCL).

Spatiotemporal data of total CVOC concentrations show widespread CVOC distributions that vary in space and time (Figure 6). The variability of CVOC spatiotemporal distributions is attributed to: (1) changes in contamination input sources; (2) dynamic fate and transport processes;
(3) variation in remedial and monitoring schemes; (4) differences in spatial and temporal data resolution; and (5) changes in groundwater use. Changes in contaminant input result from addition or removal of contaminated sites. Fate and transport processes (e.g., advection, dispersion, sorption, degradation, mass transfer) affect the mobility and persistence of the contaminants. In karst systems, these processes are influenced by the mode of transport (i.e., conduit vs. diffusive flow) and by the heterogeneous of the systems. As a result, fate and transport processes are also influenced by time-dependent hydrogeological conditions. Changes in water extraction practices driven by hydrological and water quality conditions further accentuate the variability in spatiotemporal concentration distributions. The effect of monitoring scheme and data resolution (density) variations on the apparent spatiotemporal concentration distribution can be observed when comparing spatial concentrations distributions for different years. Total CVOC concentration distributions for 1990 (Figure 6a) appear to be of lower extent and concentration ranges than that for 1984 (Figure 6b), suggesting a decrease in the contamination
plume. Comparison of the 1990 distributions with 1996 and 2000 (Figure 6c and 6d), however, show a stronger and much widespread contamination at a later time. The apparent increase is not caused by higher contamination, but by an increase in the amount of spatial data available for the analysis from the period between 1993 and 2000, which showed the greatest amount of data collected (Figure 3a). Changes in monitoring well locations also captured higher concentrations near the VA-PWS superfund site that were previously unnoticed. The smaller extension of the contamination in 2009 (Figure 6e) is not only attributed to decreasing concentrations due to natural attenuation and remediation, but also to the lower amount of data resolution available. This data shows that the difficulty in defining concentration distribution patterns in karst region result from, not only the complex processes influencing the mobility and persistence of contaminant, but also the observation and management schemes of the contaminated sites.

Response to Remedial Action
The response of eogenetic karst aquifers to remedial actions is evaluated with respect to two superfund sites in northern Puerto Rico: the UJ and VAPSW superfund sites. These sites, although not representative of all sites in eogenetic karst systems, have been subjected to extensive temporal and spatial contamination, and provide valuable information regarding factors influencing the distribution and cleanup of contamination. Spatial concentration distribution of TCE (Figure 2a), CCl₄ (Figure 2b), and Total CVOCs (Figure 2d and 6) show that the contamination extended beyond demarked superfund sites. Extensive spatial contamination is attributed to rapid transport through conduit networks of heterogeneous character, and to the lack of rapid response to characterize, monitor, and remediate the sites. This can be seen when comparing CVOC concentration distributions associated with the VA-PWS and UJ superfund sites. Both sites show contamination extending kilometers beyond the delineated superfund sites, long temporal tailing distributions, and high capacity to store and slowly release contamination over long periods of time. Long tailing is associated with mass transfer limitations and heterogeneous distribution of variable-capacity transport regions and storage compartments for contaminants in the eogenetic karst system. The TCE concentration distribution (Figure 2a) related with the VA-PWS site, however, shows a more extensive contamination area than that of CCl₄ (Figure 2b) associated with the UJ superfund site. Although contamination in both of these sites was reported at about the same time (1982-1983), it is suspected that contamination in the VA-PWS site occurred long before it was reported. Monitoring and remediation efforts in this site were implemented long after the contamination occurred. Poor characterization and monitoring efforts, in conjunction with the hydrogeological complexity of the VA-PWS site, resulted in transport areas of unknown concentrations. Some of these areas were discovered at later stages of the remedial action implementation. Rapid response to contamination events in the UJ site, on the other hand, incited prompt implementation of site characterization, monitoring, and remediation activities. As a result, the areal extent of the UJ-related contamination was smaller than of the VA-PWS. The rapid response in the UJ site also resulted in a faster decrease of CVOC initial concentrations than in the VA-PWS site, at which concentrations are still at concerning levels (Figure 4 and 5). Comparative analysis of remedial response actions for the UJ and VA-PWS superfund sites indicate the importance of proper site characterization and design of monitoring scheme, particularly for complex sites, such as those in karst systems.

Conclusions
Analysis of the historical contamination data in the northern karst groundwater system shows a significant extent of contamination that comes from multiple sources and extends beyond the demarked sources of contamination. Groundwater quality data shows sustained level of contamination over long periods of time, suggesting that the complexity and the heterogeneity of the karst give the system a strong capacity to store and slowly release contamination. Spatial and temporal distributions of CVOCs show to be highly dependent upon the site characterization, monitoring scheme and objectives, indicating that the data does not adequately capture the contamination plumes. Although the selected remedial action (pump and treat combined with air stripping and soil vapor extraction) in two superfund sites has reduced concentrations over time, the spatial and temporal extent of the contamination reflect inability to completely capture the heterogeneous plumes in the complex karst system. The comparison and contrast of the initial response time in two superfunds shows the early response of remediation results in a rapid reduction of the contaminants concentration and more control in the extent of the contamination plume.

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DETERMINATION OF THE RELATIONSHIP OF NITRATE TO DISCHARGE AND FLOW SYSTEMS IN NORTH FLORIDA SPRINGS

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Abstract

The Suwannee River Water Management District has collected quarterly discharge and water quality data from 30 1st and 2nd magnitude springs in the Suwannee River Basin since 1998. These data were collected quarterly well into the late 2000s and constitute a valuable database for characterizing spring discharge behavior.

Trend and correlation analyses were used to compare the relationships of NO$_3^-$ + NO$_2^-$ (nitrate in this paper), specific conductance, and spring discharge. Trends were considered significant if alpha levels of the trend slopes were ≤ 0.05.

Data from 50% of the springs show that nitrate concentrations increase as discharge from the spring increases. Forty-five percent of the remaining springs showed no correlation between discharge and nitrate, and only 5% (2 springs with poor data) have relationships where high discharge was related to lower nitrate concentrations.

Twenty percent of the springs had positive correlations of specific conductance with discharge, 37% showed no correlation, and 43% had negative correlations between specific conductance and discharge.

Most important in terms of understanding the plumbing of the conduit systems, 40% of the springs had positive correlations between nitrate and specific conductance, 48% showed no correlation, and 12% had negative correlations.

Introduction

The Suwannee River Water Management District is located in north Florida (Figure 1). There are more known first and second magnitude springs within this district than any other comparable area of North America. These springs constitute a major economic and ecological resource to north-central Florida.

In recent years, many of these springs have experienced increases of nitrate concentrations in discharge from the springs (Hornsby and Ceryak, 1998; Katz and Hornsby, 1998; Upchurch et al., 2007; Harrington et al., 2010). These increases in nitrate concentrations are causing eutrophication in some springs, spring runs, and receiving waters (rivers and streams) (Florida Springs Task Force, 2000; Stevenson et al., 2004).

As a result of concern for increasing nitrate concentrations in spring discharge, Florida’s Department of Environmental Protection and Florida’s water management districts initiated comprehensive water-quality sampling programs that continue today. The Suwannee River Water Management District began intensive spring sampling in 1998. This paper reports on an analysis of these data from 1998 through 2007.

Spring Discharge Considerations

The springs all occur in Oligocene or Eocene limestones that constitute the upper Floridan aquifer. Dolostone is locally present, but the majority of the rock in contact with the water is limestone.

The springs vary in setting from vents or groups of vents situated at the heads of spring runs up to several kilometers in length to vents located on the margins of their receiving waters, primarily the Suwannee and Santa Fe rivers.

Figure 1. Springs of the Suwannee River Water Management District by magnitude.
Virtually all of these springs included in this investigation have been affected by increasing nitrate concentrations. For this reason, it is difficult to identify background nitrate concentrations. Upchurch (1992) reported that the mean nitrate as NO₃⁻ in rainfall was 0.97 mg/L (standard deviation = 1.01, n = 1373 samples), and in the upper Floridan aquifer groundwater in the Suwannee River Water Management District, he reported the median nitrate (NO₃⁻ as N) to be < 0.05 mg/L. Nitrate concentrations in many of the springs discussed in this paper exceed the “background” concentrations observed in regional Floridan aquifer water.

The typical time series for these springs (Figure 2; Upchurch et al. 2007) shows seasonal fluctuations in nitrate and specific conductance with time. However, the short-term variability is typically less than the long-term variability. In this paper, the relative behavior of nitrate, specific conductance, and discharge are of concern. Temporal trends, while present, do not affect the correlations discussed herein. For example, the relative behavior of nitrate and specific conductance at Manatee Springs (Figure 3) indicates a positive correlation that is significant at α = 0.05 regardless of seasonal and long-term temporal trends in discharge or other causes of seasonality.

Hypotheses and Methods
It has long been understood that water derived by diffuse-flow in karst systems has higher calcium, hardness, and pH as a result of longer residence times and rock/water contact than rapidly recharged water in conduit-flow systems (White, 1988). Typically, calcium, hardness, and other rock/water interaction indicators in diffuse-flow systems have negative correlations with spring discharge (Basset and Ruhe, 1973) as a result of dilution under high discharge conditions.

In the eogenetic karst of Florida, this pattern is often complicated by complex matrix flow in the doubly porous limestone combined with connections of flow systems with swallets and siphons and with back-flooding during high stage events in the adjacent rivers. Many of the springs in the Suwannee River drainage basin are known to be connected to swallets and in-stream siphons (Hornsby and Ceryak, 1998; Butt et al., 2007). Others have unknown connections with surface-water sources. The majority of the Suwannee River basin springs are located on the shores of the rivers (Figure 1) and, depending on headstage relationships, the springs act as estavelles.

In order to identify springs that are dominated by diffuse versus conduit flow and estavelles that are dominated by bank-storage events, correlation and trend analyses were undertaken comparing specific conductance (μS/cm), a surrogate for calcium, hardness, and other rock/water interaction indicators, with spring discharge (m³/s). As a result of concerns about the sources of nitrate in the springs, correlation and trend analyses were also used to compare nitrate (NO₃⁻ + NO₂⁻) with discharge and specific conductance.

Methods
All field analytes were obtained by trained staff, and the chemical and field analyses followed EPA and state protocols. Significance of correlations was determined by the coefficient of determination (R²) and alpha level (α ≤ 0.5).
As noted above, unless stated otherwise, the term nitrate refers to nitrate plus nitrite ($\text{NO}_3^- + \text{NO}_2^-$) concentrations.

**Hypothesis**

It is proposed that the relationships between discharge, specific conductance, and nitrate should provide insight as to the dominant flow system and nitrate sources in each spring. Figure 4 presents the hypotheses used to rationalize the origins and flow systems.

In model A (Figure 4), it is hypothesized that a negative correlation and trend between discharge and specific conductance represents the dilution of diffuse flow water by rapidly recharged, conduit-flow water. The conduit flow could be derived from swallets or siphons distal to the spring or it may represent discharge of riverine water stored in the spring conduit system and, to some degree, intergranular porosity in the eogenetic limestone. The latter source should only occur shortly after a flood in the adjacent stream with sufficient stage to cause backflow of the spring.

A positive correlation would suggest that water contained in a diffuse flow system is flushed from the system with minimal dilution. This relationship may be complicated by the widespread use of high specific conductance, Floridan aquifer water for irrigation.

Model B (Figure 4) shows the hypothesized relationships of nitrate and spring discharge. If nitrate concentrations increase as discharge increases, the nitrate is contained in a surface-water source. A negative correlation suggests that the nitrate is stored in the soil or rock column and is diluted by conduit flow.

Model C (Figure 4) shows the proposed relationships between specific conductance (an indicator of diffuse flow) and nitrate (an analyte that originates at or near the land surface by human activity). A positive correlation between the two analytes suggests that the source of the nitrate is similar to the source of the total dissolved solids represented by the specific conductance. In other words, the nitrate is stored in the soil or rock column or applied upon irrigation. A negative correlation suggests that the nitrate is derived from a surface-water source and is entering the aquifer through a swallet or siphon or by backflow of riverine water into the spring conduiting.

Many of the springs in the Suwannee River basin are estavelles that take water when river levels are high. This complicates interpretation of the conduit-dominated models. If the spring discharges relatively elevated nitrate concentrations without high river levels, one can assume that the nitrate is derived from within the springshed and transported to the spring by conduit flow. However, if the elevated nitrate, low specific conductance discharge occurs in the waning phases or shortly after an episode of high river stage during which bank storage and estavelle action in the spring occurred, then the surface-water source may not be within the springshed.

**Results**

Table 1 summarizes the results of the correlation and trend analyses. Figure 5 illustrates the relationship of specific conductance to discharge in the springs for which there was sufficient data.
**Trends of Specific Conductance with Discharge**

For the most part, there is an expected negative correlation and trend between discharge and specific conductance. This pattern is consistent with the relationship identified by Bassett and Ruhe (1973). The cluster of springs in the middle Suwannee River (Figure 5) that showed positive trends can be explained by the proximity of the springs to several large farms that irrigate with Floridan aquifer water derived from the same horizon as the spring water.

**Trends of Nitrate Concentrations with Discharge**

Figure 6 depicts the trends of nitrate versus discharge in springs in the Suwannee River basin. With two exceptions (Table 1), the springs show either a positive trend or a non-significant relationship. This positive trend suggests that the sources of nitrate are related to surface water entering the limestone aquifer through swallets, siphons, or sinkholes that do not have associated streams, or drainage of riverine water from estavelles.

Model B (Figure 4) shows expected patterns of nitrate concentrations as a function of spring discharge. If the nitrate is stored in soil or rock, the Bassett and Ruhe pattern should appear. That is, nitrate should be slowly leached from the strata during low flow, but high discharge through conduits would be expected to dilute the nitrate. Conversely, rapid recharge of elevated nitrate surface water through a conduit or release of elevated nitrate riverine water stored in an estavelle should produce a trend with a positive slope.

Most of the springs had positive slope relationships between nitrate concentrations and discharge (Figure 6).

The two exceptions with negative slopes are located in an area of heavy irrigation and, apparently, high fertilizer use.

Figure 7 illustrates the positive trend relationship in a spring (Columbia Spring; Figure 1) that is known to be dominated by riverine recharge. Much of the water discharging from this spring enters the aquifer via a siphon on the Santa Fe River just upstream from the spring (Butt et al., 2007). In this system, discharge from the spring increases as stage in the river rises, so the nitrate concentrations reflect conditions in the river.

### Table 1. Summary of correlations of discharge, nitrate, and specific conductivity.

<table>
<thead>
<tr>
<th>Correlation</th>
<th>Number of Springs</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Positive Correlation</td>
</tr>
<tr>
<td>Discharge v. nitrate</td>
<td>22</td>
</tr>
<tr>
<td>Discharge v. Sp. Conductance</td>
<td>9</td>
</tr>
</tbody>
</table>

**Figure 5.** Trends of specific conductance versus discharge in springs of the Suwannee River system.

**Figure 6.** Trends of nitrate ($\text{NO}_3^- + \text{NO}_2^-$) versus discharge in springs of the Suwannee River basin.
Trends of Specific Conductance Compared to Nitrate Concentrations

When nitrate concentrations are compared to specific conductance through time, there is an interesting pattern (Table 1) of significant correlations. Figure 8 shows the distribution nitrate and specific conductance trends in the springs. According to the hypotheses (Figure 4C), those springs with statistically significant, positive correlations (red dots on Figure 8) should reflect concurrent flushing of pore water and nitrate during high discharge events. Almost half of the springs with statistically significant correlations (Table 1), showed this positive correlation between nitrate and specific conductance.

Figure 9 shows an example of the pattern of nitrate and specific conductance from a spring with a positive correlation between the analytes, and Figure 10 illustrates the behavior of these analytes through time. This spring, Ginnie Spring (Figure 1), is located in a resort with campgrounds and is down gradient from several dairies and other sources of nitrate. In this spring, elevated nitrate concentrations that are more-or-less synchronous with high specific conductance occur during low flow.

The low concentrations in late 2004 reflect a flood event in the adjacent Santa Fe River. In this case, low specific conductivity river water appears to have diluted the concentrations of groundwater discharging from the spring.

Figure 7. Trends of nitrate (NO$_3^-$ + NO$_2^-$) versus discharge in Columbia Spring, a known resurgence of the water derived from the Santa Fe River. The spring is located in Columbia County.

Figure 8. Trends of nitrate (NO$_3^-$ + NO$_2^-$) versus specific conductance in springs of the Suwannee River basin.

Figure 9. Trends of nitrate (NO$_3^-$ + NO$_2^-$) versus specific conductance in water discharging from Ginnie Springs, a second magnitude spring on the south side of the Santa Fe River in Gilchrist County.

Figure 10. Time series of nitrate and specific conductance at Ginnie Springs. Low concentrations of the analytes results during high discharge events.
In contrast, those springs with negative correlations between nitrate and specific conductance reflect conduit flow (Figure 4C) and/or storage. Many of the springs with significant negative correlations have associated known swallets or siphons. In these springs, elevated nitrate discharge appears to reflect stormwater events within the springshed.

A few springs that exhibit negative correlations between specific conductance and nitrate do not have known swallets or siphons. One example is Poe Springs, a second magnitude spring located in a county park on the Santa Fe River in Alachua County. This spring (Figures 11 and 12) is located down gradient from farms with row crops and pasture land.

The negative correlation reflects episodes of discharge of low specific conductance, elevated nitrate water that correspond with flood events in the adjacent Santa Fe River (Figure 1) in 1998 and 2004. From this relationship with flooding, it appears that the low specific conductivity and elevated nitrate water (Figure 12) represents discharge of surface water stored in the aquifer during the flood. In other words, the data suggest that either Poe Springs is an estavelle or discharged water stored in the conduit system by a siphon located up gradient from the spring.

Summary and Conclusions
Investigation of the relationships between discharge, specific conductance, and nitrate in the springs of the Suwannee River basin suggests that the interactions of the aquifer flow system, adjacent rivers and streams, and diffuse versus conduit flow may strongly affect the patterns of nitrate release from springs. These data, which are derived from a doubly porous aquifer, clearly suggest that a simple assumption that nitrate is derived from a specific land use within the springshed may be erroneous.

Comparison of trends of time-series data allow for sorting out the myriad processes that affect spring discharge and water quality. In addition to measuring the discharge and water-quality parameters within a spring or spring run, it is advised that river discharge and stage relationships be considered and that inventories of swallets and siphons be included as part of a spring nitrate study.

References

Figure 11. Trends of nitrate (NO$_3^-$ + NO$_2^-$) versus specific conductance in water discharging from Poe Springs, a second magnitude spring on the south side of the Santa Fe River in Alachua County.

Figure 12. Time series of nitrate and specific conductance at Poe Springs. Variations in the analytes reflect diffuse flow at low discharge and surface water inflows during high flows.
THE MILLION DOLLAR QUESTION: WHICH GEOPHYSICAL METHODS LOCATE CAVES BEST OVER THE EDWARDS AQUIFER? A POTPOURRI OF CASE STUDIES FROM SAN ANTONIO AND AUSTIN, TEXAS, USA

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Abstract
This article describes resistivity imaging and natural potential (NP) data collected over six caves between the years of 2000 and 2014, which are air filled and are located in the northern part of Bexar County, San Antonio, and in the south and north of Travis County, Austin, Texas. All caves were encountered through drilling and/or excavation for construction and utility lines or power pole reconstructions. The study area falls into the part of the Recharge Zone of the Edwards Aquifer region and it represents a well-developed karstified and faulted limestone (Stein and Ozuna, 1996).

The resistivity and NP data over these 6 caves suggest that the resistivity data does not specifically determine where karstic features are located in the subsurface. However, it provides significant information on the near-surface geology and geological structure. The NP data, on the other hand, notably defines the location of cave features. Thus the merits of integrating the NP method along with the resistivity imaging over the Edwards Aquifer, in order to reduce the ambiguity in the interpretation, are evident.

Introduction
Currently, several geophysical methods exist to locate subsurface voids. These geophysical methods are resistivity (2D and 3D), NP, ground penetrating radar (GPR), gravimetry, magnetics, electromagnetics, and seismic (refraction, reflection and shear waves). The NP method is also called as self-potential.

Detecting incipient sinkholes, bedrock cavities, rock pinnacles, and other karst-related features using these geophysical methods has been proven over the years (Ahmed and Carpenter, 2003; Dobecki and Church, 2006). But each method has limitations in depth and resolution accuracy based on geological factors and void size, shape, and orientation. In addition, some methods, such as gravity, and seismic, take longer and they may be cost-inhibitive.

We have collected geophysical data over the Edwards Aquifer in the San Antonio and Austin areas for the last 15 years. We have used almost all methods mentioned above. Based on these results, we conclude that the best methods have been the combination of NP and resistivity techniques (Saribudak, M., 2010; Saribudak, 2011; Saribudak et. al., 2012a; Saribudak et. al, 2012b; Saribudak et al, 2013).

The 2D resistivity method images the subsurface by applying a constant current in the ground through two current electrodes and measuring the resulting voltage differences at two potential electrodes some distance away. An apparent resistivity value is the product of the measured resistance and a geometric correction for a given electrode array. The geometric factor incorporates the geometric arrangement of the electrodes and contributes a unit length, giving apparent resistivity values in units of ohm-meters (Ω-m). Resistivity values are highly affected by several variables; including the presence of water or moisture, the amount and distribution of pore space in the material, and temperature.

Based on our experience on the Edwards Aquifer, the expected resistivity for weathered limestone varies between 50 to 300 Ω-m, while fresh limestone is expected to produce a range of values between 350-10,000 Ω-m and more. The presence of moisture or groundwater reduces resistivity values. The presence of air-filled caves causes the highest resistivity values, but it is rare that caves are purely filled with air. A variety of sediments accumulates in caves and can be preserved more or less intact for long periods of time (Palmer, 2007). The presence of sand, gravel, and clay deposits; mineralization; faults and fractures; and perched water in caves are the rules rather than the exception. Clay-filled caves cause low resistivity values.

We acquired the resistivity data using an Advance Geosciences, Inc. (AGI) SuperStingR1 and R8 resistivity systems. We processed the data using AGI’s 2D EarthImager software.
Natural electrical currents occur everywhere in the subsurface. In seepage or cave investigations, we are concerned with the unchanging or slowly varying direct currents (D.C.) that give rise to a surface distribution of NPs due to the flow of groundwater within permeable materials. Differences of potential are most common in the millivolts range and can be detected using a pair of non-polarizing copper sulfate electrodes and a sensitive measuring device (i.e. a voltmeter or potentiometer). It should be noted that water movement should be present within or surrounding a cave in order to determine a void or cave location. Positive and negative NP values are attributed to changes in the flow conditions and the resistivity distribution of the subsurface. The source of NP anomalies can be also due to changes in topography, soils and rock conditions. It should be noted that NP measurements made on the surface are the product of electrical current due to groundwater flow and the subsurface resistivity structure. The NP anomalies do not provide information on the depth of their sources.

Two Case Studies from San Antonio Area

The location of two caves from the San Antonio area is shown with a red square in Figure 1.

Cave 1

A series of voids (cave 1) was encountered during the installation of piers into the Person Formation of Edwards Aquifer limestone (Stein and Ozuna, 1996) for a construction project. These voids had a depth of about 4 m (15 ft) and appear to be connected. Combination of lowering a tape and a video camera indicated that the cave extended as deep as 50 ft. The cave was wet and air-filled.

Following the discovery of the voids, geophysical surveys were conducted to evaluate the extent of the cave and the voids. Geophysical surveys included, resistivity, NP and ground penetrating radar methods.

Four resistivity profiles, with a profile spacing of 6 m (20 ft) were acquired across the pier locations and adjacent areas. Figure 2 displays one of the resistivity imaging profiles along with 4 borehole locations, three of which encountered the cave. The resistivity data show that the cave encompasses high resistivity (1000 Ohm-m),
medium (750 Ohm-m), and low resistivity values (200 Ohm-m).

Four resistivity profiles were combined to create a 3-D block diagram and is shown in Figure 3. A 3-D top-view of the cave area is also shown in Figure 3. The known void locations encountered by borehole drilling are shown with red circles. Three borehole locations that did not encounter the cave are shown with yellow circles. Note that the boundaries of the cave defined by the borehole data include the low and medium resistivity values as in the 2-D resistivity profile. The 3-D image of the resistivity data appear to define the geometry of the cave much better than the 2-D resistivity data.

Figure 4 shows a NP profile along the same resistivity profile shown in Figure 2. The NP data indicates a significant low anomaly where the cave is located. Correlation of the both data sets suggest that it would have been difficult to determine the precise location of the cave with only the resistivity data without either having boreholes or the NP data.

Cave 2
Cave 2 was observed along a utility trench in the north of San Antonio (Figures 1 and 5). The trench was about 4 m (15 ft) deep and 35 m (112 ft) long. The cave was air-filled and its width along the trench was about 4 m (15 ft). A measuring tape was lowered into the cave and its apparent depth was determined to be 9 m (30 ft).

Figure 6 displays the resistivity data along the utility trench. The cave’s dimensions are also superimposed on the resistivity data. The resistivity profile indicates

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**Figure 2.** Resistivity data across cave 1 along with pier locations drilled into the limestone. Black lines indicate the geometry of the cave.

**Figure 3.** A map view of 3D resistivity block diagram showing the cave geometry. Note that cave location corresponds to low resistivity values (light blue color).

**Figure 4.** NP data across cave 1 along with pier locations drilled into the limestone.
medium range resistivity values (300 to 800 Ohm-m), not high resistivity values, across the air-filled cave. The cave’s geometry defined by the resistivity data is quite correlative with the observed dimensions of the cave. The resistivity data also indicate a well-defined high resistivity anomaly between stations 49 and 55 m (160 and 180 ft), which could be interpreted as an air-filled cave by a novice interpreter based on the resistivity data only.

The NP data provided in Figure 7 shows a significant low NP anomaly across the cave. However, the NP data does not indicate any anomaly over the high resistivity anomaly that was located to the north of the cave.

**Three Case Studies from the Austin Area**

Three case studies were performed over the Edwards Aquifer in the Austin area (see Figure 1 for general location). A cave location was determined during the geophysical field work and borehole drilling in the year of

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**Figure 5.** A picture showing the cave location along the trench. The cave is located 3.5 meter below the ground.

**Figure 6.** The resistivity profile along the trench cave.

**Figure 7.** The NP data across the Trench cave.
The purpose of the study was to locate potential karstic features along a transmission line, which consisted of 25 transmission poles with 300 m (1000 ft) spacing.

**Cave 3**
A resistivity survey was conducted across the location of transmission number 15, and is shown in Figure 8. The resistivity values across the profile range between 10 and 10,000 Ohm-m. The resistivity data did not indicate any significant karstic features beneath the proposed transmission pole location. However, the NP data collected along the same profile shows a high NP anomaly where the proposed pole is located (Figure 9). This is a typical NP anomaly indicating presence of a cave.

A borehole was drilled at the proposed location, down to 25 ft depth and a 2.5-inch downhole camera was lowered into it. A cave passage at 5.2 m (17 ft) depth was encountered and it blew moist air. Another karstic feature (a minor void and a fracture) was observed at 7.2 m (24 ft) (Pete Sprouse of Zara Environmental, LLC, Pers. Comm., 2010).

In the light of the borehole data, the resistivity data did not show any specific anomaly indicating the potential presence of the cave; however, the NP data did display a unique M-shaped anomaly where the cave is located. The pole location was relocated to 20 ft to the north of the proposed location and did not have voids or caves.

**Caves 4 and 5**
The City of Austin Watershed Protection performed a hydrogeologic investigation related to the design and construction of the Martin Hill Transmission Main (TM) on the Northern Edwards Aquifer Recharge Zone. Several karst features have been identified by the City of Austin in the vicinity of the Recharge Zone. These features include a sinkhole/cave opening located behind McNeil High School; the McNeil Bat Cave, located on the east side of the high school; and 3 caves (Weldon Cave, No Rent Cave) located west of the high school and McNeil Bat Cave. To acquire such information and address these concerns multiple geophysical surveys (resistivity, NP, GPR, magnetic and conductivity) were performed across the site (Figure 10). The GPR, magnetic, and conductivity data did not provide useful subsurface information due to the presence of cultural features and the conductive soil along the geophysical profile. In this paper only the resistivity and NP results along the McNeil Road profile will be discussed.

![Figure 8. The resistivity data across a proposed transmission pole location. The black line indicates a borehole drilling location.](image1)

![Figure 9. NP data across the proposed transmission pole.](image2)
A combination of resistivity and NP data from the west side of the study area is provided in Figure 11. The resistivity data shows a high resistive layer undulating under a low resistive layer along the profile. There is no striking resistivity anomaly due to a karstic feature across the Creek. However, the NP data displays a significant anomaly, in terms horizontal coverage of 60 m (~200 ft) and a magnitude of 50 mV.

Another combination of resistivity and NP data from the east side of the study area, where the McNeil High School is located, is shown in Figure 12. The resistivity data shows a high resistive unit (red and yellow in color) in the middle of the profile and it is enclosed by two low resistivity layers below and above. The high resistivity unit appears to thicken to the east of the letter B. This observation would signal a clue to an experienced interpreter that there could be a karstic feature in this area.

The NP data, however, clearly displays a major anomaly between the stations 121 m (400 ft) and 168 m (400 and 550 ft), and is annotated with the letter B. The maximum magnitude of this anomaly is about 40 mV.

During the months of summer and fall of 2014, a major construction activity started along the geophysical profile. Bulldozers excavated the water transmission line down to a depth of 6 m (20 ft) on the McNeill Road. Two caves (Cave 4 and Cave 5)) were encountered at a depth of 5 m (17 ft) where the NP anomalies A and B are located. Picture of Cave 4 and Cave 5 are provided in Figures 13 and 14.

Cave 6
A cave feature (Cave 6) was confirmed in the sidewalls and floor of a wastewater line (WWL) trench and manhole excavation located on the Northern Edwards Aquifer Recharge Zone, a few miles to the north of McNeil Road (Figure 15).

The cave 6 represents a bedding plane cave that has developed into a groundwater flow channel. The feature lies at approximately 6 m (20 ft) below ground surface, and has exposed openings along approximately 22 m.

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**Figure 10.** Location of a geophysical profile—a mile long-along the McNeill Road and McNeil High School in north Austin, Texas. Two stars shown with red and white colors (A and B) are locations where significant NP anomalies are observed.

**Figure 11.** Resistivity and NP data from the west side of the study area. Note that a significant NP anomaly is detected across the creek and is shown with the letter A.
(71 ft) of the trench and manhole excavation sidewalls. The visible length of the cave is about 35 m (115.0 ft) in length, 3.5 m (12.0 ft) average width, and about 1.5 m (5 ft) in average height (see Figure 16).

After the discovery of the cave, geophysical surveys (resistivity and NP) were performed to map the karstic features. The purpose of the work was to define geology along the wastewater line and map potential karstic features. The length of the profile was extended 200 ft further north from the northern end of the trench.

The resistivity data is given in Figure 17. The cave locations on the western sidewall of the trench are exposed on the southern and northern ends, and are superimposed on the resistivity profile. A groundwater flow channel is

Figure 12. Resistivity and NP data from the east side of the study area. The letter B indicates a significant NP anomaly.

Figure 13. Cave 4 was observed where the NP anomaly A is observed (see Figure 10).

Figure 14. A void was encountered where the NP anomaly B is observed (see Figure 11). This void is enlarged to the north towards the McNell High School and became a cave (Cave 5).

Figure 15. Site map showing the location of the geophysical profile, and the geometry of the cave, which was defined by trenching. The length of the geophysical profile is about 122 meter (400 feet).

Figure 16. A picture showing the part of the cave which was encountered during the excavation.
observed from the northern cave to the southern cave. Resistivity values in the vicinity of the caves vary between 50 to 5000 Ohm-m. It is difficult to determine the cave locations based on the resistivity data.

Note that the high resistive pinnacle shown with a red color between the two caves on the resistivity section, based on the trenching, is not defined as a karstic feature. The NP data is provided in Figure 18, which indicates a strong but linear NP gradient towards the north. It is not possible to detect small NP anomalies along the profile with the superimposition of such a high gradient. The source of the high NP gradient could be due to the significant ground water flow from the north to the south.

The majority of the high gradient NP data was clipped out (a sort of regional removal) between stations 76 m (250 ft) to (121) 400 ft, and the rest of the profile is provided in Figure 19. The NP data indicates three NP anomalies as shown with letters A, B and C. The locations of these anomalies are correlative with the two cave locations exposed on the side wall of the trench. The resistivity data did not show the presence of the air-filled caves along the trench; however, the NP data did locate them with a good accuracy. The trench was completed up to the northern end of the geophysical profile without encountering any void as the NP data predicted.

Conclusion

It is clear from the ongoing discussion above that the 2-D resistivity data does not specifically determine where karstic features are located in the subsurface. However, it provides significant information on the near-surface geology and geological structure. In addition, the combination of 2D and 3D resistivity measurements illustrates the subsurface conditions in a sufficiently accurate manner as shown in Cave 1 case study.

The NP data, on the other hand, notably defines the location of karstic features. Thus the merits of integrating NP method along with the resistivity imaging, in order to reduce the ambiguity in the interpretation, are evident.

Figure 17. Resistivity data along the trenched wastewater line. Locations of caves encountered on the western sidewall of the caves are indicated as dashed red lines filled with white color. There is a groundwater flow from the northern cave to the southern cave.

Figure 18. NP data along the trenched wastewater line. Note that there is a strong NP gradient towards to the north.
Thus the best methods are chosen to be the NP and resistivity techniques over the Edwards Aquifer.

Acknowledgments
A research paper like this is a journey from the past, and I am grateful to the many people who helped me along the way. Thanks to Art Lange, my mentor on NP method, who has been instrumental for me to understand the NP technique and apply it correctly in the field. I dedicate this paper to him.

Special thanks to Alfred Hawkins, who has been my associate since late 1990s, for his help all those years and sharing the load of the fieldwork with me, and having a good time most of the time. I would like to thank Nico Hauwert, Sylvia Pope of City of Austin, for providing information on some of the data that is presented in this paper. I thank Janet Atkinson at City of Austin Water Utility for giving permission to publish the geophysical data on McNeil Road. Finally, but not the least, I thank Aaron Googins for allowing me to present some of the geophysical data on Cave 6.

References
ROLLALONG RESISTIVITY SURVEYS REVEAL KARSTIC PALEOTOPOGRAPHY DEVELOPED ON NEAR-SURFACE GYPSUM BEDROCK

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Abstract
Following flooding in September 2013, several areas in northern Eddy County, New Mexico (NM) were damaged by multiple sinkhole collapses. We conducted rollalong electrical resistivity (ER) surveys for subsurface cavities parallel to roads within and near the community of Lakewood, NM to guide road repairs. The rollalong method allowed us to generate exceptionally long, continuous ER profiles of the survey area. ER surveys attained a maximum exploration depth of 55 to 62 meters over a lateral extent of ~1000 meters, revealing an unconformable surface developed on gypsum bedrock, punctuated by shallow depressions. Subsurface stratigraphy, including clay-rich valley fill sediments, and mudstone and gypsum of the underlying Seven Rivers Formation, can be identified by vertical and lateral variations in electrical resistivity. The irregular bedrock surface of the Seven Rivers Formation reflects paleotopography developed on that surface prior to its burial by floodplain sediment. Some of the negative paleotopographic features are probably filled sinkholes, which may be associated with shallower karstic features not imaged on the profiles.

Background
In the aftermath of flooding in September 2013, several areas along Lakewood Road and Lake Road within the village of Lakewood, NM (Figure 1), were damaged by multiple small (1-2 m depth) sinkhole collapses (Figures 2 and 3). The National Cave and Karst Research Institute (NCKRI) was contracted to conduct electrical resistivity (ER) surveys parallel to the roads in an effort to identify subsurface cavities and guide road repairs.

The sinkholes in the study area, cover collapse sinkholes, form by the piping of soil and alluvium into underlying karstic cavities. Their position along the two roads is probably the result of drainage channels along both shoulders of the roads, which have promoted groundwater recharge in these linear areas for many years. The piping of unconsolidated materials created cavities in the alluvium that slowly stopped up toward the surface and into the soil. Heavy rains and flooding in September 2013 focused substantially greater flow down into the soil in the channels until the cavities became sufficiently large and unstable, causing them to collapse and breach the surface.

Geologic Setting
Effective remediation of cover collapse sinkholes requires accurate characterization of piping zones and
is highly variable because of paleotopographic relief developed on the underlying bedrock. Bedrock in the study area is represented by the Seven Rivers Formation, the middle unit of the Permian-age Artesia Group, which is composed of interbedded reddish-brown mudstone, siltstone, dolomite and gypsum. The Seven Rivers Formation is exposed along the McMillan Escarpment on the eastern margin of the Pecos River valley, where it is capped by dolomites of the Azotea Tongue (Kelley, 1971).

Subsurface dissolution of gypsum within the Seven Rivers Formation and other members of the Artesia Group has caused local and regional subsidence and has profoundly influenced topography along the margins of the Pecos River Valley. The presence of these highly soluble rocks has also contributed to the formation of sinkholes and caves (Land, 2003; Stafford et al., 2008). Lake McMillan, less than one kilometer east of Lakewood, was ultimately abandoned because of loss of water through karstic conduits within the underlying Seven Rivers Formation (Cox, 1967).

Based on a survey of well records in the area, depth to the top of the Seven Rivers Formation in the vicinity of Lakewood ranges from 26 to 49 m. One well intersected a cave within the Seven Rivers Formation at a depth of ~60 m.

**Methods**

The basic operating principal for electrical resistivity imaging involves generating a direct current between two metal electrodes implanted in the ground (current electrodes) in order to create a potential field. This potential field is sampled at multiple locations between two additional implanted electrodes (potential electrodes) in order to measure distortions around subsurface targets. The structure shown in 2D resistivity profiles and 3D models allows geologic and hydrologic interpretation and comparison with ground truth data derived from boreholes or other direct investigations. Previous work (e.g., Land and Veni, 2012; Land, 2013) has shown that resistivity imaging is one of the most effective methods for identifying water-filled and air-filled voids due to their strong contrast with surrounding bedrock.

Modern resistivity surveys employ an array of multiple electrodes connected with electrical cable. Over the course of a survey, pairs of electrodes are automatically measured...
by means of a switchbox and resistivity meter. The depth of investigation for a typical ER survey is approximately one-fifth the distance between the furthest measuring electrodes (effectively, the length of the array of cable in 2D surveys). NCKRI staff conducted six electrical resistivity surveys parallel to Lake and Lakewood Roads (Figure 3), using 112 electrode arrays at a 3-m electrode spacing, for a maximum deployment length of 333 m per array and an anticipated depth of investigation of approximately 65 m. This spacing and target penetration depth were selected to determine if substantially larger cavities occur in the subsurface than were indicated by the size of the observed sinkholes. Resolution of an ER survey is estimated by taking one half of the electrode spacing. The 1.5m resolution of this array is finer than typically used in engineering studies over karst terrains. This degree of resolution was chosen to better define both small and large voids formed in the gypsum bedrock that were of particular concern.

Four of the surveys employed a rollalong method to increase the length of the resistivity profiles. After data were collected for each 112 electrode array, the first half of the array was shifted forward to the far end for a 50% overlap. Although this method does not increase the depth of investigation, it permits a nearly seamless ER profile much longer than the length of the main array. While resistivity data were collected, a survey-grade global positioning system (GPS) survey was conducted of all electrode positions for each array. The elevation data were used for advanced processing to correct for variations in topography along the survey line. We used an Advanced Geosciences, Inc. (AGI) SuperSting R8/IP/SP electrical imaging system to collect the data and EarthImager-2D™ software to process the data.

**Results and Discussion**

**Lakewood Road, East of Railroad**

Two rollalong resistivity surveys were conducted parallel to Lakewood Road, one on each shoulder, extending west from the intersection of Lake and Lakewood Roads to the Burlington Northern and Santa Fe (BN&SF) railroad tracks (Figure 3). These surveys achieved a maximum depth of investigation of 62 m below ground level. Variations in electrical resistivity are influenced by subsurface stratigraphy and facies variations (Figure 4), and by the presence of subsurface cavities. Our interpretation of the resistivity data is informed by subsurface records from area water wells, including an augmentation well drilled by the New Mexico Office of the State Engineer (NM OSE) in 2006, located approximately 3.2 km southwest of the survey area.

Three distinct stratigraphic units can be identified on the resistivity profiles. A thin zone of relatively high resistivity occupies the uppermost ~5 m of the section and extends almost continuously across the profile from west to east. This high resistivity layer most likely represents air-filled porosity in unsaturated soil and gravel.

![Figure 4. Electrical resistivity profiles conducted along north and south shoulders of Lakewood road, east of BN&SF railroad tracks.](image-url)
Underlying this high resistivity layer is an interval ~30 – 50 m thick of generally lower resistivity (5 to 30 ohm-m). This layer probably represents a more clay-rich section of alluvial material of the Pecos River floodplain. Clays typically have lower resistivity than more coarse-grained sediments.

A zone of moderately high resistivity extends across the profiles ~40 to 50 m below ground level. Based on well records in the survey area, we interpret this interval to represent mudstone and gypsum of the Seven Rivers Formation, which crops out along the McMillan Escarpment on the opposite side of the Pecos River. Undulations in the top of the Seven Rivers Formation reflect the paleotopography of the unit prior to burial by floodplain alluvium. A prominent depression in the top of the Seven Rivers is visible on the north shoulder profile between 138 and 165 m. This feature may represent a buried sinkhole filled with more electrically conductive clay, as indicated by a pod of very low resistivity material above it.

**Lakewood Road, West of Railroad**

Because of the presence of the railroad tracks, it was not possible to conduct continuous rollalong surveys on Lakewood Road. For this reason two separate surveys were conducted on the north and south shoulders on the west side of the railway (Figure 5). These surveys achieved a maximum exploration depth of 60 m.

A thin layer of high resistivity is also present on the profiles west of the railway, reflecting air-filled pore space in sand and gravel of the soil horizon. This layer is underlain by 40 – 50 m of generally lower resistivity material of the Pecos River floodplain. However, several lenticular bodies of higher resistivity are also present within this section. One distinctive high-resistivity lens is present on the south shoulder profile between 110 and 140 m, ~15 m below ground level. A more laterally extensive high resistivity layer occurs beneath the west end of the north shoulder profile. Subdued reflections of both of these features are also present on the opposite shoulder profiles. Because of their geometry and their presence within the alluvial section, we interpret these features as lenticular bodies of coarse sand and gravel in the unsaturated zone; it is unlikely that they represent caves. In spite of the presence of several small sinkholes along the shoulders of Lakewood Road west of the railway, no obvious subsurface cavities are visible on the ER profiles.

The top of the Seven Rivers Formation is poorly defined in the resistivity survey data collected west of the railway. It is possible that the top of the Seven Rivers dips locally to the west and exceeds our depth of investigation in this area. Although regional dip is to the east, the topography of the Pecos Valley has been strongly influenced by local and regional subsidence due to subsurface gypsum dissolution (Bachman, 1984; Land, 2003), resulting in the presence of a number of sediment filled basins along the western margin of the river valley (Lyford, 1973).

**Figure 5.** Electrical resistivity profiles conducted along north and south shoulders of Lakewood road, west of BN&SF railroad tracks.
Lake Road

Two rollalong resistivity surveys were conducted along the west and east shoulders of Lake Road, extending north about one kilometer from the intersection of Lake and Lakewood Roads. Both surveys attained a maximum depth of investigation of 55 m.

The near-surface, high resistivity soil horizon is not as well defined on the Lake Road ER surveys. However, the lower resistivity, clay-rich alluvial section is clearly visible, extending along the entire length of both profiles (Figure 6). The Seven Rivers Formation is clearly defined by a high resistivity zone beneath the northern half of the profile ~25 m below ground level, and appears to be dipping gently to the south.

The top of the higher resistivity section that represents the Seven Rivers Formation is interrupted in several places, reflecting paleotopography developed on the top of the formation. One of these breaks occurs on the west shoulder profile between 600 and 650 m, where a sag in the top of the unit most likely represents a filled sinkhole. Subsidence over this feature appears to have caused a local near-surface accumulation of coarser sediment, as indicated by an overlying lenticular high resistivity zone. This feature is also visible on the east shoulder profile between ~630 and 670 m. Several less well-defined breaks in the Seven Rivers profile may also represent filled sinkholes.

A small but well defined pod of higher resistivity occurs at 305 m on the east shoulder profile, ~5 m below ground level, overlying a depression in the bedrock surface. A second high resistivity pod occurs directly beneath the first at ~20 m below ground level. These features are likely to represent subsurface cavities partially filled with sediment, or two segments of a fissure adjacent to or extending beneath the roadway. A 4-m long soil fracture was identified near this section of the ER profile during the February, 2014 reconnaissance survey (Figure 3, number 16).

The ER profiles show no unambiguous near-surface (<5 m) resistivity anomalies that would indicate shallow subsurface cavities, in spite of the presence of several shallow sinkholes and earth fissures along both shoulders of Lake and Lakewood roads, immediately adjacent to the survey lines. We attribute the absence of such anomalies to the array configuration and electrode spacing, which was designed to detect larger cavities at greater depths.

Conclusions

Electrical resistivity surveys conducted parallel to Lakewood and Lake roads near the community of Lakewood, NM attained a maximum exploration depth of 55 to 62 m. Subsurface stratigraphy, including a near-surface soil horizon, clay-rich alluvium of the Pecos Valley floodplain, and mudstone and gypsum of the Seven Rivers Formation, are indicated by vertical and lateral variations in subsurface electrical resistivity. The irregular surface of the Seven Rivers Formation on all profiles reflects paleotopography developed on that surface prior to its burial by floodplain alluvial sediment. Some of the negative paleotopographic features are almost certainly filled sinkholes, which may be associated with shallower karstic features not imaged on the profiles. The absence of a well-defined Seven Rivers section on ER profiles along Lakewood road west of the railroad tracks may be due to a broad zone of subsidence caused by subsurface dissolution of evaporites, accompanied by a thicker accumulation of overlying floodplain alluvium.

Figure 6. Electrical resistivity profiles conducted along west and east shoulders of Lake road.
References


INTEGRATION AND DELIVERY OF INTERFEROMETRIC SYNTHETIC APERTURE RADAR [INSAR] DATA INTO STORMWATER PLANNING WITHIN KARST TERRANES

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Abstract
As part of two USDOT-funded studies focused on the development of satellite-based Interferometric Synthetic Aperture Radar (InSAR) technology, the researchers integrated InSAR-derived point cloud data into the transportation design process to optimize the location of a stormwater management system in a karst terrane. After initial validation, the InSAR data (over 1.67 million data points comprising various “scatterers”) were brought into a GIS dataframe and georeferenced to locations of known sinkholes. This dataset was then used to evaluate karst hazard within a 40x40km data frame located in the Valley and Ridge Province of Virginia. The group identified systematic kinematic differences in scatterer behavior with respect to their proximity to mapped karst geohazards, and used this method to identify unknown karst features, revealing numerous previously unidentified sinkholes. After validating the data with quantitative field correlations, the group integrated the dataset into a traditional CADD-developed design, ported into a GIS environment, and utilized the resulting integrated dataset to optimize the location of stormwater management assets within a traditionally-developed roadway project. In the process, the group developed open-source data delivery, allowing greater flexibility, efficiency, and optimization of the infrastructure design and planning process conducted collaboratively over geospatial platforms. This data integration offers lifecycle cost benefits, improvements to the safety of the traveling public, and protection of the environment, particularly in groundwater-sensitive karst terranes. A case study of this approach is presented.

The views, opinions, findings and conclusions reflected in this presentation are the responsibility of the authors only and do not represent the official policy or position of the US Department of Transportation/Office of the Assistant Secretary for Research and Technology, or any state or other entity.

Introduction
InSAR Data and Potential Value
Synthetic Aperture Radar (SAR) is the extension of traditional radar data acquisition, in which the orbit of a satellite is used to synthetically mimic a much larger aperture (i.e., “synthetic” aperture), which allows for the delivery of images of very high resolution (Rosen et al., 2000). Each ground resolution element, or pixel, contains phase and amplitude data of the backscattered radar wave for each satellite flyover, or “acquisition”.

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By applying interferometric techniques to a time series of acquired images (a “stack”), it is possible to interpret the difference in reflected radar waves in a manner that reveals changes in topography over time (Power et al., 2006). These combined images, or interferograms, are generally termed InSAR or DInSAR when applied to ground motion. Pixels that exhibit stable radar signatures over time are referred to as permanent scatterers (PS). Techniques detecting PS are known under the general term of persistent scatterer interferometry (PSI). Often sets of neighboring pixels show behavior as a group without any individual pixel providing a stable reference; in this case, these pixels are combined in a larger geographic area, and are referred to as distributed scatterers (DS). In this work, the group made use of datasets derived using two specific techniques referred to as PSInSAR (Ferretti et al., 2001) and SqueeSAR (Ferretti et al., 2011). A further refinement to these techniques provides information about scatterers that either gain or lose radar reflectivity over a temporal subset of the stack. These are referred to as temporary scatterers (TS). Under ideal conditions, InSAR data can provide millimeter-scale records of vertical change (Morgan et al., 2011). The authors use the term InSAR as a general term for all SAR applications related to topographic change and infrastructure evaluation.

While SAR data has been available since the 1950s (Sherwin et al., 1962) and airborne InSAR was first used in the early 1970s (Graham, 1974), it was not until the 1990s that InSAR was used to investigate topographic changes such as those that occur after earthquakes (Massonnet et al., 1993). Many of those applications were for large-scale, slow-moving changes, such as slowly-moving landslides (Roering et al., 2009) or changes in rock-glacier mass (Strozzi et al., 2010). Applications to smaller phenomena, such as formations of sinkholes, activity on rock slopes, or distortions to bridges or rock buttresses, have not been targets of investigation for InSAR until quite recently.

The authors validated and evaluated the use of InSAR for such smaller-scale applications by bringing the entire InSAR dataset (PS, DS, and TS) into a GIS dataframe and correlating to control data. For karst hazards, these correlative datasets included published maps of sinkhole locations and karst terranes, and records of repaired sinkholes. For infrastructure, the displacement time series of the InSAR data were used to identify potentially compromised geotechnical assets, and the observations were quantitatively validated by field inspection (Vaccari et al., 2013).

The value of InSAR data, once validated, is evident to planners and designers. It allows generation of GIS-based geohazard, geotechnical, and surface kinematic databases. It also allows optimization of geotechnical planning in the light of a larger and more dynamic dataset than was previously available. In the Valley and Ridge Physiographic Province of Virginia, evaluating the InSAR data with regard to karst geohazard related to transportation planning and design has proven to be useful. It is worth noting that the cost of remediation, repairs, and maintenance of sinkhole occurrences alone was approximately $1,000,000 USD during the period of Virginia fiscal years 2012 to 2014, exclusive of any cost associated with economic harm caused by transportation disruptions. A cost-benefit analysis of the wide use of InSAR data is ongoing at the Virginia Department of Transportation, but the potential for significant cost and safety benefits is clear.

**Research Projects**

**RITA-RS-11-H-UVA and RITA-RS-14-UVA**

The authors are cooperative investigators in RITA-RS-11-H-UVA (RITA11) and RITA-RS-14-UVA (RITA14), two USDOT-funded projects titled “Detection & Bridge/Landslide Monitoring for Transportation Infrastructure by Automated Analysis of Interferometric SAR Images” and “InSAR Remote Sensing for Performance Monitoring of Transportation Infrastructure at the Network Level,” respectively. The Research and Innovative Technology Administration (RITA), now supplanted by the Office of the Assistant Secretary for Research and Technology (OASRT), coordinates the US Department of Transportation’s (DOT) research programs. RITA, and subsequently OASRT, is charged with developing innovative, interdisciplinary technologies to improve the US transportation system. The initial project, RITA11, focused on evaluating whether InSAR could be successfully used to detect and quantify surface change and thereby detect incipient sinkholes. In order to make this determination, the data were first broadly validated by comparison to geotechnical assets and field conditions. Validation was performed by a team of a Virginia-licensed professional geologists and engineers. The subsequent project, RITA14, expands the analyzed areas and focuses on integrating InSAR data into the
design process, integrating planning and InSAR datasets into a GIS dataframe in order to create a decision support system which is more efficient, more cost-effective, and better protects the environment.

**Selection of Area of Interest (AOI)**
The authors selected a 40x40 kilometer area of interest (AOI) for InSAR data acquisition. The area, represented in Figure 1, was selected based on geological diversity and the presence of numerous geotechnical assets. This offered the potential for the formation of sinkholes and other karst features, as well as deterioration of or distortions to assets within the AOI due to karst conditions. Numerous unmapped sinkholes were detected during this stage of the investigation (Bruckno et al., 2013).

One AOI, common to both RITA11 and RITA14, is centered at -79.222°W, 38.077°N in Augusta County, Virginia. It is tectonically complex, spanning the Valley and Ridge, and Blue Ridge physiographic provinces (Dietrich 1990). Geological ages ranging from Holocene to Precambrian (Bartholomew, 1977), with frequent unconformities, are represented within the AOI. The predominant tectonic framework consists of eastward-dipping thrust faults and decollements related to repeated orogenic cycles (Rader and Wilkes, 2001). The AOI contains carbonate, non-carbonate clastic, and metamorphic terranes, resulting in both rock slope stability hazard and severe karst hazard. The karst areas range in age from Cambrian to Devonian and formed during the Taconic and Acadian Orogenies and their associated divergent and inter-orogenic periods. Karst lithologies consist of limestone and dolostone, while non-carbonate clastic lithologies consist of interbedded shales, siltstone, conglomerates, and sandstone, and the metamorphic lithologies consist of charnockite, granulite gneiss, quartzite, greenschist, and blueschist-grade metabasalt. Figure 2 represents areas of karst geohazard within the AOI.

Several control datasets exist for sinkholes; Figure 3 is an aggregate dataset of known sinkhole locations compiled from the Soil Survey Geographic Database (SSURGO, 2015) and limited-release data from the Virginia Department of Mines, Minerals, and Energy.

**Selection of Satellite and Resulting InSAR Data**
COSMO-SkyMed, a constellation of four identical satellites built and operated by the Italian Space Agency, was selected for data acquisition. Each satellite is equipped with an X-band SAR operating at 9.6 GHz (Italian Space Agency, 2007). Between August 29, 2011 and June 16, 2014, 57 non-uniformly spaced SAR scenes were acquired and were processed by TRE-Canada, Inc. using the PSInSAR and SqueeSAR algorithms, which convert the data into subsidence. The resulting dataset consists of over 1.67 million PS, DS, and TS scatterers, as well as amplitude values for each 3x3 meter pixel within the entire AOI corresponding to each acquisition.

Figure 4 represents the processed InSAR scatterers, with PS, DS, and TS points all represented in blue. Heavily vegetated areas prevented backscatter from the ground.
Data Validation by Detection of Active Sinkholes

Because of the robust set of control data and maps of sinkhole occurrence, the research team was able to identify scatterers whose location coincided with locations of mapped sinkholes. Analyzing the timeseries of those scatterers, a typical example of which is illustrated in Figure 6, allowed the research team to create simple search tools that screened for unmapped sinkholes. This was accomplished by identifying scatterers with combinations of negative displacement, velocity, and acceleration.

During the data validation period of the RITA11 research, several unmapped sinkholes were identified using these methods. Figures 7 and 8 illustrate the growth of one such sinkhole.

Case Study
Integration of InSAR Data into Transportation Planning

The transportation planning process does not generally involve a sinkhole mapping program. Typically, a literature survey is conducted to evaluate karst features that may affect the project. However, the literature used for such purposes is often not current, digital versions may suffer from imperfect digitization, and the scale of such maps is often inappropriate for use in transportation planning. Integration of InSAR data into the process offers the opportunity to conduct planning and design decisions in the light of dynamic and recent data. The authors implemented this approach on a

and showed rapid temporal decorrelation; however, such areas tend to have limited human population and infrastructure, and are therefore of lesser value in terms of surface analysis.

Figure 5 illustrates scatterers within an area of new construction within the AOI. The scatterers represent both anthropomorphic and geomorphological features. Buffering on roads within GIS can eliminate apparent scatterers falling outside of areas of concern, and automated edge detection methods can remove scatterers coinciding with buildings from the dataset (Ferraioli, 2010).

Figure 3. Regions within the AOI (outlined in red) mapped as sinkholes by the VA DMME (in red), by SSURGO (in blue) and sinkholes repaired by VDOT 2000 to 2012 (in green) (ESRI ArcMap™).

Figure 4. Regions within the AOI (outlined in red) represented by scatterer data (represented by blue points) (ESRI ArcMap™).

Figure 5. Close-up of an area within the AOI represented by scatterer data (PS are represented in blue, DS in orange, and TS in green) (ESRI ArcMap™).
be optimized for soil type, topography, the consideration of local landowners, and, in karst terranes, the need to avoid groundwater contamination and active karst features. The mere presence of mapped sinkholes may be of no concern if the sinkholes are not subsiding, but actively subsiding sinkholes should be avoided.

Figure 10 illustrates those PS and DS scatterers near the construction project showing only the most negative velocity, and TS scatterers showing the greatest negative displacement over the data acquisition period (90th percentile of the dataset, or PS and TS velocities greater than -6.0 mm/year and TS displacements greater than -15.0 mm during the acquisition period). Scatterers coinciding with obvious anthropomorphic features, such as buildings and transportation infrastructure, were manually removed, so that the remaining scatterers reflect geomorphological subsidence.

Underlying the scatterer dataset is the aggregate sinkhole dataset with a multi-ring buffer extending to the maximum extent of anticipated sinkhole influence. The pattern of the InSAR scatterers showing only the most negative velocity or greatest subsidence coincides with the pattern of the sinkholes. This indicates not only the presence of sinkholes, but that the sinkholes are not yet in a state of post-collapse or meta-stability (the majority of sinkholes in the Virginia Valley and Ridge are subsidence, rather than cover-collapse, sinkholes).

Because survey control was available for the project, the CADD files were portable into an ESRI ArcMap™ environment, and all of the files, along with pertinent GIS files, could be georeferenced within a common coordinate system. From there they were ported to a Google Earth Pro™ environment, where they could be quickly assessed by planners, designers, and representatives of the public in open meetings; the data could also be shared and evaluated across remote offices using ArcGIS for Organizations™. This allowed regions of greater or poorer favorability for stormwater management basins to be evaluated: Areas near actively subsiding sinkholes, and areas near production wellheads, were to be avoided.

Figure 11 illustrates the areas of mapped sinkholes with a geographic buffer zone, the PS and DS scatterers showing the most negative velocity, the TS scatterers showing the most negative displacement,
Figure 9. Excerpt from Microstation™-drafted plans.

Figure 10. Mapped sinkholes overlain by scatterers showing the most negative velocity or displacement (PS are represented in blue, DS in orange, and TS in green) (ESRI ArcMap™).
and commercial water-supply wells with a geographic buffer. The proposed construction plans from Figure 9 are georeferenced to the image.

From the image, it can be seen that the region northwest of the proposed construction shows several problematic conditions with regard to stormwater management basin locations. Not only are there historical records of sinkholes in the region, but the InSAR data shows that the area is actively subsiding.

Figure 12 illustrates the scatterer behavior for the InSAR/Sinkhole cluster northwest of the proposed intersection in Figure 11. Several of the scatterers show a net displacement approaching 15 to 25 mm during the data acquisition period, suggesting a fast rate of subsidence.

The behavior of the InSAR scatterers within the geographic buffers of the wells also varies. The scatterers within the annulus around Well Cluster 2 show an average velocity of +0.1 mm per year, while the average velocity within the annulus of Well Cluster 1 show an average velocity of -1.16 mm per year (for clarity, these scatterers are not shown in Figure 11). Both wells are terminated in water-table aquifers, suggesting that the surface depression around Well Cluster 1 may be the result of a cone of depression around an overstressed aquifer; the areas near Well Cluster 1 are therefore less favorable for stormwater management basins than those near Well Cluster 2.

Areas where the geographic buffer for known sinkholes coincide with the selected (most negative velocity/displacement) scatterers, and overlap the geographic buffer for Well Cluster 1, are the least favorable areas for stormwater management basins and should be avoided; such an area is located northwest of the proposed intersection reconstruction. This is also an area of a topographic low (potentially the result of the area being a doline related to the sinkholes); absent other data, this area would naturally be seen as favorable for stormwater management basins. However, overtopping of a stormwater management basin in this area during weather events outside of the recurrence interval for which the basin was designed may result in the inadvertent construction of an injection well.

Areas to the northeast of the intersection are clearly better suited for stormwater management basins due to their distance away from mapped sinkholes correlated with areas shown to be subsiding according to InSAR data, mapped sinkholes, and Well Cluster 1. While stormwater management basins in this region may require more excavation in order to provide for positive drainage, a risk-reward analysis shows that avoiding potential groundwater contamination validates this decision.
Conclusion and Discussion

Protection of aquifers is particularly important in karst terranes, where there is often a direct hydrologic link between surface runoff and the water table via sinkholes. While records of sinkhole locations and sinkhole location maps are important tools in planning and design, the data contained in such maps are often outdated, may be found at a scale inappropriate to the planning process, and may suffer from poor digitization. Data derived from InSAR platforms, on the other hand, record a time-series of surface behavior which may be correlated with actual karst behavior. Where these data can be integrated into the design process, they offer lifecycle cost benefits, improvements to the safety of the traveling public, and protection of the environment, particularly in groundwater-sensitive karst terranes.

References


DETECTION OF VOIDS IN KARST TERRAIN WITH FULL WAVEFORM TOMOGRAPHY

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Abstract
This paper presents an application of time-domain surface-based waveform tomography for detection of voids in karst terrain. The measured seismic surface wave fields were inverted using a full waveform inversion (FWI) technique, based on a finite-difference solution of 2-D elastic wave equations and the Gauss-Newton inversion method. The FWI was applied to real experimental data sets collected from twelve test lines at a karst site in Florida. Two of the test lines were located next to open karst chimneys to image their extent. Ten other test lines were located in an open and flat area without any void indication from the ground surface to detect an unknown void. The inversion results show that the waveform analysis was able to delineate embedded low-velocity anomalies, a void, and highly variable bedrock both laterally and vertically. Locations of the low-velocity anomalies were consistent to the known open chimneys observed from the ground surface. The unknown identified void was confirmed by an independent standard penetration test (SPT).

Introduction
Embedded void detection in a site usually begins with non-destructive testing (NDT), as NDT data can provide general subsurface conditions over a large volume of materials. At suspicious locations (anomalies), more involved invasive methods such as the Cone Penetration Test (CPT) or Standard Penetration Test (SPT) are then conducted to obtain more detailed information. Various approaches have been developed and employed to characterize voids, ranging from gravity, resistivity, ground-penetrating radar and traditional seismic wave methods. These methods have both advantages and limitations in identifying and quantifying voids.

As reviewed by Plessix (2008) and Vireux and Operto (2009), the full waveform inversion (FWI) approach offers the potential to produce higher resolution images of the subsurface by extracting information contained in the complete waveforms rather than approaches using only the dispersive properties of Rayleigh waves or first-arrival signals. Nasseri-Moghaddama et al. (2007), for example, have clearly shown that the recorded responses at the surface can carry valuable information regarding the presence and characterization of anomalies, e.g., voids, below the surface. However, FWI is computationally intensive, requiring a full solution of the governing wave equation. Many algorithms for waveform inversion have been developed and applied to synthetic and real seismic data in large-scale (kilometer-scale) domains (Shipp and Singh, 2002; Ravaud et al., 2004; Sheen et al., 2006; Cheong et al. 2006; Brenders and Pratt, 2007; Choi and Alkhalifah, 2011; and others). In larger scale experiments surface waves can clearly separate from body waves and be removed in the inversion process. However, at shorter length scales (meter-scale), it is difficult to separate body waves from surface waves, and only a few studies of waveform inversion involving both body and surface waves have been performed for near-surface investigations on synthetic data (Ge’lis et al., 2007; Romdhane et al., 2011) or real experimental laboratory data (Bretaudneau et al., 2013).

A 2-D full waveform inversion (2-D FWI) technique (Tran and McVay, 2012) was reported which inverted both body and surface waves in the case of real experimental data. The technique includes forward modeling to generate synthetic wavefields and employs the Gauss-Newton inversion method to update model parameters.
portion of the site. The chimneys were formed due to sinkhole activities. Line 1 was at the grid line G6–G42, next to open chimneys 1 and 2. Line 2 was at the grid line 28A–28K, perpendicular to line 1, and next to open chimneys 2 and 3. Due to safety concerns, the two test lines were conducted about 1 m away from the chimneys.

<table>
<thead>
<tr>
<th>Line 1 (G6–G42)</th>
<th>Line 2 (28A–28K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Geophone spacing: 1.5 m</td>
<td>Geophone spacing: 1.2 m</td>
</tr>
<tr>
<td>Receiver spread: 34.5 m</td>
<td>Receiver spread: 27.6 m</td>
</tr>
<tr>
<td>Shot spread: 36.0 m</td>
<td>Shot spread: 28.8 m</td>
</tr>
</tbody>
</table>

Unlike line 1, the geophone spacing of 1.2 m (instead of 1.5 m) was used in an attempt to characterize smaller chimneys.

For line 1 data analysis, a proper initial model is required to avoid the inversion being trapped in local minima. For simplicity, an estimate of the initial model was established via a spectral analysis of the measured data. A linear increasing S-wave velocity from 200 m/s at the surface to 400 m/s to a depth of 18 m (half of test line length) over a length of 36 m was considered. The initial P-wave velocity for the domain was calculated from the S-wave velocities assuming that the initial Poisson’s ratio throughout the domain was 0.25. The mass density throughout the model was kept constant at 1,800 kg/m$^3$ for all inversions. Three inversion runs were performed for frequency ranges with central frequencies of 10, 15, and 20 Hz, beginning from the lowest frequency range. The medium of 18 m × 36 m was divided into about 1200 cells of 0.75 m × 0.75 m. During the inversion, S-wave and P-wave velocities of cells were updated independently, and each run was stopped after 20 iterations when the observed waveform data and the estimated waveform data were similar.

The final results are shown in Figure 2A for analysis of the data at 20 Hz. Locations of chimneys 1 and 2 are also shown in Figure 2A for comparison. The final inverted S-wave profile (Figure 2A, top) shows two low-velocity zones at distances 12 m and 21 m, along with high lateral and vertical variations in limestone boundaries ($V_s > 800$ m/s) at the bottom of profile. Evidently these anomaly
locations were the same as those of the chimneys (i.e., 12 m and 21 m). Note that the exact depths of chimneys were not measured due to safety concerns.

The inverted P-wave profile (Figure 2A, bottom) was consistent with the estimated S-wave profile. Chimney 1 of about 1.5 m diameter was also characterized in both S-wave and P-wave images. Chimney 2 of about 1 m diameter was not shown, due to 3-D effects. To characterize the smaller chimney, the test line may have needed to be closer to the chimney, and data at higher frequencies (20–40 Hz) may also have been required.

Data analysis for line 2 was similar to that of line 1. The medium of 14.4 m × 28.4 m was divided into about 1200 cells of 0.6 m × 0.6 m. Three inversion runs were performed for frequency ranges with central frequencies of 10, 15, and 20 Hz; and the final results for data at 20 Hz are shown in Figure 2B. Locations of chimneys 2 and 3 are also shown in Figure 2B for comparison. The final inverted S-wave profile (Figure 2B, top) shows a low-velocity zone at distance 20 m near chimney 2, along with high lateral and vertical variations in limestone boundaries (S >800 m/s) at the bottom of profile. A valley of low-velocity area was found at distance 8 m near chimney 3. The inverted P-wave profile (Figure 2B, bottom) was consistent with the estimated S-wave profile.

For further verification of the inverted profiles, S-wave velocity profiles from two different perpendicular lines that intersected are shown in Figure 3. The intersection was at distance 22 m of line 1 and distance 18 m of line 2. The similarity of two independent S-wave profiles suggested consistency and credibility of the FWI.
Four inversion runs were conducted for frequency ranges with central frequencies of 6, 10, 15, and 20 Hz, beginning from the lowest frequency range. The medium was divided into cells of 0.75 m x 0.75 m. During inversion, S-wave and P-wave velocities of cells were updated independently, and each run was stopped after 20 iterations. The final inversion results at 20 Hz are shown in Figure 4. The final inverted S-wave profile (top) shows a void embedded at about 6 to 9 m depth (S-wave velocity less than 50 m/s), along with high lateral and vertical variations in weathered limestone (S-wave velocity more than 600 m/s) boundaries at the bottom of profile.

The inverted P-wave profile (bottom) is consistent with the S-wave profile. To verify the seismic test results, a Standard Penetration Test (SPT) was conducted at the predicted void location (distance 18 m) by the sponsor (FDOT) three weeks after the seismic test, and the SPT ‘N’ values are shown in Figure 4B. It is interesting that a void exists at this location that was embedded at about 4 to 7 m depth, as the SPT ‘N’ values are zeros (void filled by air) or very low (void filled by raveled soil). Although the 2-D FWI showed the useful capability to locate the void, the predicted depth (6 to 9 m) is deeper than the
real depth of the void (4 to 7 m), this is mostly attributed to the discrepancy between the estimated waveform data (plane strain) and the measured data (non-plane strain). Non-plane strain measured data is due to the 3-D void and applied point loads (hammer strikes).

Conclusions
An application of Gauss-Newton inversion of full seismic elastic waveforms is presented for a highly variable (horizontal and vertical) site. The full waveform inversion successfully identified complex subsurface profiles including low-velocity embedded zones, a void, and highly variable limestone surfaces at the bottom of profiles. The inverted results are consistently identified known open karst chimneys in the unconsolidated sediments observed from the ground surface. The independent inverted S-wave velocity profiles at the intersection of two perpendicular test lines are similar, suggesting consistency and credibility of the full waveform inversion technique. The identified void was confirmed by an independent standard penetration test (SPT). For the cases presented, full waveform inversion is computationally practical, as the results obtained were all achieved in about three hours of computer time on a standard laptop computer.

References

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Figure 4. Northern portion: (A, top) S-wave and P-wave velocities (m/s) and (B, bottom) SPT ‘N’ value.


CHARACTERIZATION OF KARST TERRAIN USING GEOPHYSICAL METHODS BASED ON SINKHOLE ANALYSIS: A CASE STUDY OF THE ANINA KARSTIC REGION (BANAT MOUNTAINS, ROMANIA)

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Abstract  
To understand karst topography, we must determine both the nature and the factors that are defining dissolution processes in soluble rocks, as well as the drainage network resulting from these processes. The goal of this paper is to understand the underground drainage direction configuration and, also, the factors that are involved in surface water drainage of the Anina karstic region.

In this study we used two complementary geophysical methods, spontaneous potential (SP) and ground penetrating radar (GPR), applied in 5 sinkholes with a funnel shaped aspect. Four of these sinkholes are circular and one of them is elongated NW-SE direction. Three of the studied sinkholes are representing a chain of sinkholes orientated west-east.

SP data describe the surface drainage, indicating drainage direction and/or moisture accumulation points. The GPR investigation utilizes electromagnetic pulses for the investigation of subsurface dielectric properties. GPR offers an image of the underground, showing possible bedding planes, in this case mostly along north-south orientations. Besides, in two GPR profiles, we could identify an object that could be a cavity, in that point were on SP grid the values indicate small values, pointing out a link between those two geophysical results. Using SP and GPR methods we were able to show that the bottoms of these depressions are retaining more humidity and soil. In addition, the GPR profiles outlined several subsurface “objects”, at a depth ranging between 20 and 40 meters, which need a more thorough analysis.

Our future work is intended to enrich our field data using SP and GPR methods, to compare with our first results. Also, we intend to integrate electrical resistivity tomography measurements in our analysis for better subsurface characterization.

Introduction  
In Romania limestone represents almost 2% of its surface. The largest and most compact area of carbonate rocks in Romania is within the Reşiţa-Moldova Nouă Synclinorium, situated in the SW of the country, in the unit called Banat Mountains.

Karst terrain results from rock masses dissolution, having as a consequence an effective underground flow (Waltham et al., 2005). To understand karst topography, we must determine both the nature and the factors that are defining dissolution processes in karst soluble rocks as well as the drainage network resulted from these processes. (Ford, Williams, 2011).

The density and size of sinkholes indicate the degree of dissolution that geological substrate has undergone locally (Shofner et al., 2001). The fractures and their orientation in a karstic area give important knowledge regarding the drainage network, due to the fact that the karst system depends highly upon them (Chalikakis et al., 2011).

The study case of this paper is located in one of Banat Mountains’ subunits, Aninei Mountains. This approach is a comparative study using spontaneous potential (SP) and ground penetrating radar (GPR) as geophysical
methods. SP is a passive and an electrical geophysical method, which quantifies naturally occurring electrical fields at the Earth’s surface.

The self-potential surveying is based upon measuring the spontaneous or natural potentials developed in the earth by electrochemical actions between minerals and subsurface fluids or by electrokinetic processes involving the flow of ionic fluids (Sharma, 2002).

Also SP in the subsurface is caused by a number of processes that are not well understood at this time (Reynolds, 1997).

In recent years the SP method has found increasing use in geothermal, environmental, and engineering applications to help locate and delineate sources associated with the movement of thermal fluids and groundwater.

The spontaneous potential method has been used for many years in karstic areas (Stevanovic, Dragisic, 1998; Lange, 1999; Rozycki et al., 2006; Guichet et al., 2006; Jardani et al., 2007; Jardani et al., 2009, Jouniaux et al., 2009; Robert et al., 2011).

GPR is a non-destructive geophysical tool that can produce a continuous profile in cross section or record features underground without drilling, boring, or digging. GPR profiles are normally used to assess the location and depth of underground objects, and investigating the presence or the continuity of the natural subsoil conditions (Apel and Dezelic, 2005). The resulting GPR image (also called a radargram) is very similar to a seismic reflection profile. Acquisition of data by means of GPR is based on the propagation, reflection and distribution of high-frequency electromagnetic waves (generally from 10 to 1000 MHz) to the underground. Using GPR’s in karst areas partially covered with alluvial deposits is not very common, mainly due to alluvial deposits, clay content, which is hindering the penetration depth of GPR systems (Anchuela et al., 2008). Therefore, the results obtained will depend upon the type of soil and its degree of saturation, compaction, mineralogy, and also on the frequency of antennas used (Anchuela et al., 2009). If the study area contains clayey soils, it is recommended the GPR method should not be used in the sinkhole investigation (Zisman et al, 2013).

Studies to detect cavities using the GPR method were done by Chamberlain et al. (2000), Kadioglu and Ulugergerli (2012). Other goals for GPR applications in karstic regions are sinkhole detection and characterization: Anchuela et al. (2010), for sinkholes detection near Zaragoza (Spain), Anchuela et al. (2013) with a paper on the current development of sinkholes, Gutiérrez et al. (2011) combined GPR with different techniques for sinkhole characterization, and Nouioua et al. (2013) using GPR and ERT. Al-fares et al. (2002) developed a study for a karst aquifer structure involving also GPR measurements. A study that involves both methods used in this paper was done by Carpenter et al. (2013) near Cancun, Mexico.

In Romania, geophysical methods are not often used for karst investigations, even if there are many interesting karstic regions. There are two papers using resistivity methods, Mafteiu (1991) and Mitrofan et al. (2008), vertical electrical sounding (VES) using Schlumberger and pole-dipole arrays.

Mafteiu (1991) observed the vertical and horizontal plane of the fracturing effect that predetermines the development of the Cave of Padiș, and in the Padiș Plateau he identified the border between fissured limestone and compact limestone. Mitrofan et al. (2008) managed to delineate with success a concealed flow path in Hercules spring (Cerna Valley).

The goal of this paper is to analyze the terrain of the Anina karstic region to understand the underground drainage direction and the factors that are involved in surface water drainage. This study is based on field data collection during five field campaigns, from May 2013 to November 2014.

**Study Area**

The Anina karstic region is situated in the South-West of Romania, within the Banat Mountains, as shown shaded in yellow in Figure 1.

Geologically, the study area is located in the central part of the Reșița - Moldova Nouă Synclinorium, the largest and most compact, homogeneous structure covered by carbonate rocks in Romania (Orășeanu and Iurkiewicz, 2010).

We developed our study on the Mârghitaș Plateau, a suspended karstic plateau without surface water drainage, located in the northern part of the Anina karstic region (Figure 2).
There are many surface karstic landforms that may be seen: sinkholes, sinkholes valleys, blind valleys, dry valleys, karrens, and karren fields (Figure 3).

**Field Methods**

The characterization of karst regions requires specific knowledge of both surface and those forms of underground features, and application of the geophysical methods are an option to study the subsurface in connection with the surface landforms. One of these methods, which is also used in the analysis of the groundwater, especially in karst areas, is spontaneous potential (SP). The second method that completes our geophysical approach is GPR.

For SP data, we used two Petieau nonpolarizing electrodes, a fixed electrode and a mobile one. The measurements were made with a digital multimeter, Volcraft VC 850. We measured SP at 11 sites, repeating measurements 2 or 3 times, in different seasons and atmospheric conditions for comparison purposes. Our approaches for SP measurements are represented by profiles with N-S and E-W orientations and grids. Each
electrode was placed inside a hole, 10 cm deep in the soil and after 1 minute we noted the value indicated on the voltmeter (in mV) and then we moved the mobile electrode. The station spacing for the mobile electrodes was 5 m.

For the GPR method, we used a MALA RAMAC Georadar with two antennas, 50 MHz and 25 MHz. Because we developed our study in a karstic area, our goal was to identify voids on the radargrams below the locations of the SP anomalies. Because in the study area other geophysical studies are missing, we chose as a first approach to use an antenna that give a deeper penetration in the subsoil, trying to have better image regarding the underground in the Mărghitaș karstic plateau. Based on previous study on limestone and due to the fact that the RTA antennas are of compact type, no Common Midpoint (CMP) or Wide Angle Reflection and Refraction (WARR), thus an overall wave velocity of 0.12 m/ns was used for the depth conversion of the radar signal (Kadioglu and Ulugergerli, 2012; Apel and Dezelic, 2005). GPR results are represented by 11 profiles, 50 MHz frequency with a depth of 22 meters, 8 profiles with the 25 MHz antenna with a depth of penetration of 46 meters and 1 profile also with the 25 MHz antenna with a depth of 54 meters. In the next section of this paper, we present the results obtained. We focus on 3 sites using the 25 MHz antenna, being the most representative in these measurements. The measurements presented in this paper were made near the Mărghitaș Hotel (Figure 4).

Results
Site 1
SP results for the sinkhole in Site 1 were obtained in May 2013 and November 2013. Measurements obtained after the first campaign express mostly negative values, with some positive values. In the autumn 2013 all SP values were positive, excepting one borehole where we obtained a negative value.

The field measurements were interpolated based on the 5 m distance between holes, rasterized and contoured using ArcGIS 10 software developed by ESRI (http://www.esri.com/software/arcgis), obtaining the raster presented in Figure 5. Water flow in this sinkhole has similar direction as the tectonic fault orientation, N-S or NW-SE. This observation is highlighted in the contoured images, but is clearer in the data obtained in November 2013 (Figure 5b), where we can notice that the middle of the sinkhole is accumulating humidity (red) and the drainage direction is mostly north-south, as we will obtain also in GPR measurements for bedding planes orientation.

In May 2013 (Figure 5a) SP measurements indicate that on the boundaries of the sinkhole there is a direction of water infiltration, due to negative anomalies. At the bottom of the sinkhole, larger values suggest water retention, a function of the flat terrain and also based on the deep soil cover. In May 2013 the values where more different, alternating negative values (especially on the border of the sinkhole, where karrens are present) with positive values (in the middle of the sinkhole).

The SP values obtained in November 2013 are more homogeneous, due to weather conditions. The campaign was done after many months of uniform precipitations. In Figure 5b is more obvious the bottom of the sinkhole, where are the largest values, indicating the stagnancy of water (Artugyan and Urdea, 2014).
Site 2

Site 2 features a large sinkhole, with the east-west diameter of 70 meters and the north-south diameter by 60 meters. The SP measurements were made in two campaigns, October 2013 and November 2014. In October 2013 the SP values present negative values, with some positive anomalies, showing that at that period, after several weeks without precipitation, the

On the 25 MHz profiles, we notice that the sinkhole is very clearly observed, because we realized the profile longer than sinkhole’s diameters, to observe the difference in the radar signals. The first radargram (Figure 7) shows very well the bottom of the sinkhole and also the slopes, and at the end of it, we notice the buried karrens or small voids. Besides, we notice a continuous signal that we consider as a bedding plane of the area.

Figure 5a. Spontaneous Potential measurements in Site 1 in May 2013.

Figure 5b. Spontaneous Potential measurements in Site in November 2013.

Site 2

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Figure 6. Site 1 and the location of GPR profiles.
the highest values, indicating moisture accumulation and water retention due to thick soil. On N-S direction, we may observe a negative anomaly at 24 m from the start of the profile in both campaigns, which could indicate a void in the underground which quickly drains water from the surface. Even if on N-S direction the profiles are not very smooth, we can notice that except those negative anomalies at 24 m, the bottom of the sinkholes indicate the largest values, showing the tendency to retain moisture for a long time.

The aspect of this sinkhole (Figure 10) gives as the interpretation of the GPR results: in the middle there is a large accumulation of materials (organic, soil) and on the slopes the karrens are also observed on the GPR profiles. Surface is not well moisturized, being favorable for rapid flow into the underground. In November 2014 the values are positive, indicating that the drainage is more stable, the soil is more saturate with water. The middle of the sinkhole presents the highest values on E-W direction.

On the E-W orientation, in the middle of the sinkhole the SP presents values indicating accumulation, as we expect to obtain in the middle of these karstic depressions where the bottom is filled with thinner soil and the humidity presents higher values (Figure 8). The same situation is observed also in November 2014, but this time the measurements were realized after a week with higher precipitation in the area and the SP values are negative, indicating that the surface drainage is more unstable. Again SP values indicate that in the middle this geophysical method shows accumulating moisture. Also, we can notice that in both profiles there are two negative anomalies at 9 and 12 meters from the starting point of the profile and at 63 meters. These anomalies are showing that at that point the drainage is more rapid, being possibly linked to certain voids underground.

On the north-south orientation (Figure 9) the profile is more sinuous, presenting many negative anomalies, but we notice that in both campaigns at the point located at 24 meters there is a negative anomaly, possibly indicating a void in the underground where water is more rapidly drained. The north-south profile is not very expressive for this sinkhole, the bottom of it being not very obvious as for the east-west profile. This fact could indicate that on the north-south orientation the fractures in the bedrock are more developed, determining a certain behavior in water drainage. If we take into account the fault main orientation in the area, NNW-SSE, maybe we could find an explanation for the aspect of those N-S profiles.

We can observe that in both profiles in the middle of the sinkhole is well highlighted on E-W orientation, with

![Figure 7. GPR profile (25 MHz antenna) in Site 1 on east-west direction.](image)

![Figure 8. Spontaneous Potential measurements at Site 2 (east-west orientation).](image)

![Figure 9. Spontaneous Potential measurements at Site 2 (north-south orientation).](image)
Because is known that clay may perturb the GPR signals, we intend to employ in this site study electrical resistivity tomography, to describe accurately the underground and validate the GPR results.

Also, we observe that as on the first reflection radargram we could identify the continuous GPR signal, considering that it should be also bedding planes, this area being a strong faulted zone.

GPR profiles for Site 2 are designed to better explain the SP results and to give an image of the underground of this sinkhole. We can notice that GPR profiles are similar to both SP profiles, meaning that the W-E profile, shown in Figure 11, describes a smooth hyperbola for the sinkhole, with highest radar signals in the middle, but the second GPR profile, for N-S orientation (Figure 12), has many anomalies between the surface and 25 meters depth. These anomalies are also observed on the SP profiles.

**Site 3**

Site 3 is a chain of three sinkholes with west-east direction. Sinkhole 1 is the smallest one and less deep, the second one is the largest one, and Sinkhole 3 is the deepest one of this chain of sinkholes.

Sinkhole 3 consists of a circular sinkhole, a funnel shaped one, with a swallow hole in the middle. SP values were obtained at the beginning of October 2014, after several days of precipitation. The results show that in the middle, in the smaller sinkhole the water is staying for a long time, indicating moisture accumulation (larger values of SP data). Also, near the edges of the sinkhole, where karrens are present, the drainage is more rapidly, due to these rocks and thin soil. SP values indicate that the surroundings the swallow hole inside the large one presents the tendency of rapid flow into the swallow hole direction (smaller values of SP measurements) (Figure 13).

For GPR profiles, this site means a chain of three sinkholes (Figure 14) where we intend to observe on the radargrams the boundary of these karstic depressions

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**Figure 10. GPR measurements at Site 2.**

**Figure 11. GPR profile at Site 2 on west-east direction.**

**Figure 12. GPR profile at Site 2 on north-south direction.**
and if there could be certain cavities because the last 2 sinkholes (from West to East) present swallow holes.

The first radargram of this site (Figure 15), comprising all the three sinkholes is indicating very well the boundary into the underground of those depressions, the first sinkhole and the second one being more closely, and the boundary between the second one and the third one, that are separated by a dirt road. Also, we can notice that karrens present mostly on slopes of the first and the third sinkhole are observed on the GPR signals as buried rocks and also the radar signals indicate that in the middle of the third sinkhole, the largest and the deepest of these three depressions, could be a cavity or a void, based on previous GPR results in karstic areas (for example El-Qady et al., 2005; Gómez-Ortiz and Martín-Crespo, 2012). The funnel aspect of this sinkhole is favourable for a vertical cavity development. Again, we observed that countinous GPR signal that could be considered as bedding planes, but for the radargram presented in Figure 15, these are smaller than and not as obvious as in previous sinkholes.

We also obtained two radargrams on N-S orientation for sinkhole number 2 and 3 of this chain of sinkholes. Both radargrams present a signal between 30 and 40 meters depth for Sinkhole 3 (Figure 16), a deeper sinkhole, and between 20 and 30 meters for Sinkhole 2 (less deep than Sinkhole 3). We consider that object as bedding planes, being present on both radargrams that are parallel to the terrain at a distance of several meters. The velocity was established based on previous study that applied GPR on limestone (Kadioglu and Ulugergerli, 2012; Apel and Dezelic, 2005).

Combining the SP measurements, with the GPR results, we can point out that it could be certain void in the middle of this site. We observe that the signal present in the middle of the last sinkhole in Figure 14 was obtained in the N-S profile (Figure 17). For the second sinkhole we notice that there are signs that we include in the buried rocks, but we rise the question if is not also clay padding, due to the fact that under that 10 meters there is no GPR signal (Figure 18).

**Discussion**

The results of SP measurements indicate in most of the cases that there is a direction in the water circulation (based on the negative values of SP measurements), but we also obtained positive values during the dry season, most of them being measured during August and September, after large dry periods. We observe that on the karstic plateaus, starting from May the soil was very dry and hard, with very small absolute values of SP, but also with positive values in the middle of the dolines, suggesting moisture accumulation areas.

GPR radargrams indicate bedding planes at depths between 20 and 30 meters, all these profiles being along north-south orientation. On one of the radargram we
In this study we used two complementary geophysical methods, spontaneous potential (SP) and ground penetrating radar (GPR), applied in 5 sinkholes with a funnel shaped aspect. Four of these sinkholes are circular and one of them is elongated NW-SE direction. Three of the studied sinkholes are representing a chain of sinkholes orientated west-east. SP describes the surface drainage water indicating the tendency in the drainage direction or accumulation points. On the other hand, GPR describes the subsurface using the response of the materials or objects located in the underground to the signal sent by the radar antenna.

There are limitations in both methods, but they have been successfully applied in several sites for karst topography investigations (Jardani et al., 2007, 2009; Anchuela et al., 2008, 2009; Carpenter et al., 2013).

Conclusions and Future Work

GPR offers an image of the underground, showing possible bedding planes, mostly along north-south orientation. The north-south direction of the identified bedding planes are according to the main faults orientation of the studied area, NNE-SSW. Due to this aspect, we consider that the bedding planes are mostly observed on north-south profiles. Besides, in two GPR profiles, we could identify an object that could be a cavity, below and anomaly on SP grid.

There are two profiles that are pointing out some discontinuities and possible cavities. One of these profiles was measured over a chain of three sinkholes and this profile at between 6 meters and 25 meters depth, shows some anomalies that indicate differential radar signal that we associate to an object as micro-tectonic features. The homogeneous aspect of radargrams indicates that these zones are not influenced by karst activity.

Figure 15. GPR profile at Site 3 for a chain of 3 sinkholes, from west to east.

Figure 16. GPR measurements at Sinkhole 3 of Site 3.

observe a possible void or a cavity at 20 meters depth in the west to east profile. At the same depth we notice also on the north-south profile that the GPR signal point out an anomaly in the underground.

Figure 17. Sinkhole 3 of Site 3: GPR north-south profile.
Using SP and GPR methods we were able suggest that the bottoms of these depressions are retaining more humidity and soil. In addition, the GPR profiles outlined several subsurface “objects”, at a depth ranging between 20 and 40 meters, which need a more thorough analysis.

Our future work is intended to enrich our field data using SP and GPR methods, to compare with our first results. Also, we intend to integrate electrical resistivity tomography (ERT) measurements in our analysis for a better subsurface characterization.

The ERT measurements that we intend to use in the future should provide a complete image of the subsurface, and with interpreted air-filled voids on the radargrams corresponding to very high resistivity zones.

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INVESTIGATION OF A SINKHOLE IN OGLE COUNTY, NORTHWESTERN ILLINOIS, USING NEAR-SURFACE GEOPHYSICAL TECHNIQUES

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Abstract

A sinkhole measuring 40 m in diameter and up to 6.5 m deep occurs within the Nachusa Grasslands, near the town of Franklin Grove, northwestern Illinois. This area, dedicated to prairie conservation and restoration, is owned and operated by The Nature Conservancy. Several meters of unconsolidated sand, gravel, and clay overlie the St. Peter sandstone, beneath which lies karstic Prairie du Chien dolomite. Investigations included electromagnetic (EM) conductivity profiles, resistivity soundings, 2D resistivity, and ground-penetrating radar (GPR), supplemented by conductivity logs, soil cores, and tree core studies. These data indicate the sandstone averages about 5 m deep near the sinkhole rim and the sinkhole is about 115 years old. Nearby residential wells indicate an average static water level of 11 m below the surface, so the water table currently lies well below the sinkhole floor. GPR sections show abrupt termination of the bedrock reflector near the sinkhole rim, suggesting formation by collapse. Geophysical investigations also identified possible hydraulic conduits associated with the sinkhole. Specifically, GPR profiles, at 50 and 100 MHz, provide the highest resolution images of the subsurface and indicate possible conduits (soil pipes) near the sinkhole rim as diffraction hyperbolas 2-3 m below the surface. GeoProbe™ conductivity logs showing unusually low conductivity, and sudden probe drops, also suggest the presence of shallow soil cavities around the sinkhole. However, dye poured into various low spots on the sinkhole floor was never recovered, despite numerous sampling locations.

Introduction

Understanding sinkhole formation processes and age of formation are important in assessing land stability and in reconstructing geomorphic history. Identification of hydraulic conduits linking sinkholes with the underlying groundwater system is equally important in characterizing groundwater recharge and identifying potential contaminant pathways. This paper describes the use of near-surface geophysics, combined with tree-ring dating, to attack these two problems.

Site Description

This study is focused on an isolated sinkhole (locally referred to as the “Stone Barn Road sinkhole”) in Ogle County, northwestern Illinois. This area, dedicated to prairie conservation and restoration, is owned and operated by The Nature Conservancy and is now a bison preserve. The sinkhole lies in an upland area with shallow bedrock, approximately 15 km southeast of the edge of the officially defined “Driftless Area,” of the upper Mississippi River basin as shown in Figures 1 and 2. The nearest town is Franklin Grove, IL, about 6 km to the southeast; the unincorporated community of Lost Nation, with its golf course and artificial lake, lies about 1 km to the north. Figure 2 is a Lidar image of the sinkhole area. The data was acquired by a twin engine, fixed wing aircraft operated by Aero-Metric. Data was

Figure 1. Location of the study area (circle) relative to the Driftless Area (shaded) in northwestern Illinois (after Wilson, 2013).
acquired at an elevation of 1500 m above mean terrain with a horizontal accuracy of 0.27 m. Elevation accuracy (95% confidence interval) was 0.10 m (open terrain), 0.23 m (forested areas), and 0.079 m (man-made structures). A photo of the sinkhole is shown in Figure 3.

Geological setting

The study site lies approximately 7 km southwest of the Sandwich fault, a major structural feature in northern Illinois (Kolata et al., 1978). This normal fault juxtaposes Cambrian and early Ordovician sedimentary rocks to the south (the oldest rocks cropping out in Illinois) with late Ordovician sediments to the north. Average displacement along the fault is about 150 m.

Figure 2. Lidar image of the study site. Stone Barn Rd. sinkhole is circled.

Figure 3. Photo of the sinkhole floor in early spring.

Figure 4. A stratigraphic section of the study area (after Luczaj and Masarik, 2015). The St. Peter sandstone is a part of the Ancell Group.
At the site several meters of unconsolidated sand, gravel, and clay overlie the St. Peter sandstone, beneath which lies karstic Prairie du Chien dolomite. A stratigraphic section is shown in Figure 4. These will be discussed below according to their depth below the surface.

**Unconsolidated deposits (0-4 m depth)**
Using the U.S. Dept. of Agriculture (USDA) soil classification, less than 0.15 m of organic top soil exists overlying 4.1 m of sandy clay loam. This overlies 0.6 m of sandy loam, underlain by unconsolidated sand (likely top of sandstone bedrock) at the bottom of cores. No evidence exists of glacial till at the site.

**St. Peter sandstone (4 – 15 m depth)**
The St. Peter sandstone is a formation in the lower Ordovician Ancell Group. It is a fine-to-medium grained, well-rounded quartz arenite. This unit extends from Minnesota to as far south as Missouri and east-west from Nebraska to Illinois (Willman, 1975). Its commercial name, widely used as a fracking proppant, is “Ottawa sand.”

**Prairie du Chien dolomite (below 15 m depth)**
Early Ordovician dolomite lying on Cambrian strata and unconformably overlain by the St. Peter sandstone consists of the Shakopee dolomite, New Richmond sandstone, underlain by the Oneota dolomite. The Shakopee and Oneota are highly karstified and associated with hundreds of caves and sinkholes in the upper Midwest (e.g. Willman, 1975; Alexander, 1980; Ruhl, 1989).

**Geophysical Surveys**
Many studies and articles explore the use of geophysical methods to characterize sinkholes, or identify filled sinkholes (e.g. Carpenter et al., 1998; Al-fares et al., 2002; Ahmed and Carpenter, 2003; Dobecki and Church, 2006). In this study geophysical methods were used in a phased and sequential manner; first employing reconnaissance methods, such as EM conductivity surveys and resistivity soundings, followed by more detailed methods, such as GPR and 2D resistivity over anomalous or critical areas. Figure 5 shows all survey lines. Geophysical interpretations were verified and calibrated.

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*Figure 5. Image of the sinkhole showing geophysical survey lines.*
Figure 6. Conductivity logs obtained with the GeoProbeTM at various locations around the sinkhole. Possible air-filled voids and bedrock levels are noted.
using GeoProbe™ borings and conductivity logs. In addition to the geophysical surveys, tree core samples were used to determine the age of trees within the sinkhole. High-resolution Lidar images were also examined and a dye tracing experiment was conducted.

Calibration/verification of geophysical methods were achieved through GeoProbe™ borings and conductivity logs (Figure 6) outside the sinkhole that encountered sandstone bedrock beneath 2.7-5.8 m of unconsolidated sediment. Records of local private wells indicate pre-pumping water table depths of 9.1-22.6 m. Hand-auger and shovel digging within the bottom of the sinkhole revealed bedrock at about 0.5 m beneath the surface in places.

EM Conductivity Surveys
Several EM conductivity surveys were conducted around and within the sinkhole. This method was used as a reconnaissance tool to identify areas for more detailed surveys. These were performed with Geonics EM31 and EM34 conductivity meters, which have different depth responses: in the vertical dipole the EM31 has maximum response at about 2 m whereas the EM34 at a coil spacing of 10 m has a maximum response at about 5 m depth. Thus the EM31 response was too shallow to show bedrock variations, although it might show soil voids. The EM34 had sufficient depth penetration to see bedrock variations. The EM34 survey reveals elevated conductivity in a southwest-northeast trend and reduced conductivity in a north-south direction. The vertical dipole data is much noisier than the horizontal dipole data. The elevated conductivity in the southwest-northeast direction may indicate a deepening of the bedrock in that direction, perhaps coincident with a karst conduit incised into the subcropping bedrock surface. Relatively low conductivities occur in east-west and north-south orientations, suggesting bedrock highs or open air-filled fractures in those directions. The EM31 surveys also showed areas of sharply reduced soil conductivity that could be locations of air-filled voids in the soil.

Resistivity Soundings
Two resistivity soundings revealed the vertical structure of soil and bedrock around the sinkhole. One was performed directly east of the sinkhole and the other slightly south of the sinkhole rim. Sounding 1 and its layered model (from inversion using the program Resipx [Interpex, 1988]) is shown in Figure 7. It reveals low resistivity sediments overlying high resistivity unsaturated St. Peter sandstone overlying lower resistivity saturated St. Peter sandstone and Prairie du Chien formation.

2D Resistivity Profiles
Sinkhole Floor
Two-dimensional (2D) dipole-dipole array resistivity transects were made over the western portion of the sinkhole floor in east-west and north-south directions using an AGI Sting/Swift R1 system with 20 electrodes and a dipole width of 1 m. An east-west profile is shown in Figure 8 after inversion for true resistivity with the program Res2Dinv (Loke, 1998). The high resistivity is

Figure 7. Resistivity sounding and interpreted layered model obtained east of the sinkhole.
interpreted as bedrock whereas the low resistivity materials are interpreted as clay, silt and/or saturated materials. The east-west profile (Figure 8) shows abrupt termination of the high-resistivity bedrock on the sides, but also an apparent bedrock “shelf” that extends westward across the floor of the sinkhole. The much shorter north-south profiles are consistent with the east-west profile, but only show low resistivity materials below the west-central part of the sinkhole.

**Sinkhole Rim**
One 2D resistivity profile was made over the eastern sinkhole rim, as shown in Figure 9 (dipole width = 3 m). This profile shows a sharp truncation of high resistivity material (interpreted as bedrock) as the sinkhole rim is crossed.

**Ground-Penetrating Radar Profiles**
Ground-penetrating radar (GPR) profiles were performed at several locations around and within the sinkhole and across the sinkhole rim. Surveys were performed with a Sensors and Software pulse EKKO-IV unit equipped with 50- and 100-MHz antennas. Initial walkaway surveys were performed to establish the velocity of GPR waves, 0.08 m/s. This value is used to covert two-way reflector travel times in the profiles to depth.

**Sinkhole Floor**
One GPR survey was made across the western part of the sinkhole floor. The profile shows a myriad of diffractions, with no clear reflections. This is consistent with the 2D resistivity profiles that suggest an irregular bedrock surface, present in some places and absent in others.

**Sinkhole Rim**
The bedrock surface reflection exhibits a major change in character at the sinkhole rim as shown in Figure 10. In fact it is possibly truncated at the sinkhole rim, with the small wavelet remaining being an airwave reflection. This profile was made along the east side of the sinkhole along a profile more-or-less coincident with the 2D resistivity line shown in Figure 9.

**Evidence for Soil Piping and Subsurface Voids**
Geophysical data in various places suggests open voids may occur in the soil or bedrock. The Geonics EM31 profiles, in particular, show sudden decreases in conductivity, both northeast and southwest of the sinkhole, that could represent shallow air-filled soil voids and pipes. Borings were made in these same areas. “String-drops,” along with zones of extremely low conductivity were logged at SH01, SHNE1, SHNE03 and SHSW1, as noted in Figures 5 and 6. Figure 11 shows a GPR section with a large diffraction, apparently at the bedrock surface, possibly indicating a cave or void. Boring in the area of the diffraction also produced a string-drop.

**Discussion and Conclusions**
Our conclusions are that this is a soil-mantled collapse or caprock sinkhole (doline). Both these models involve
sudden roof collapse. Figures 12 and 13 below illustrate these models. In the case of the Stone Barn Rd. sinkhole several meters of soil mantle the bedrock, which is not shown in Figures 12 and 13. The caprock doline model, in particular, seems to be consistent with resistivity imaging of a bedrock “shelf” beneath the sinkhole floor.

Soil pipes and caves near the bedrock surface may also be present, and may provide hydraulic connections between the sinkhole and underlying or surrounding aquifers. A dye tracing test, however, failed to establish any sort of hydraulic connection between the Stone Barn Rd. sinkhole and underlying or surrounding aquifers.

Tree growth within the sinkhole is the only chronological data for dating the age of sinkhole formation. Due to nearly straight tree trunks with no evidence of “creep,” or curvature of the trunk, it is assumed the trees grew after the sinkhole had become dormant. We have inferred the youngest age of the sinkhole from the age of the old-

**Figure 10.** GPR profile across the east sinkhole rim. Abrupt change (possible truncation) in the bedrock surface reflection is circled.

**Figure 11.** GPR profile across what appears to be a cave (indicated by the diffraction circled in red) at the bedrock surface.

**Figure 12.** Collapse doline model (after Jennings, 1985).

**Figure 13.** Caprock doline model (after Waltham et al., 2005).
est trees. From tree cores taken, the oldest tree was found to be 115 years so the sinkhole probably formed suddenly at least 115 years ago.

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STUDY ON MONITORING AND EARLY WARNING OF KARST COLLAPSE BASED ON BOTDR TECHNIQUE

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Abstract
Brilliouin Optical Time Domain Reflectometer (BOTDR) is a newly developed measurement and monitoring technique, which utilizes Brillouin spectroscopy and Optical Time Domain Reflectometer (Jiang et al., 2006; Zhang et al., 2009; Xu et al., 2011) to measure strain generated in optical fibers as distributed in the longitudinal direction. This paper introduces the principle and characters of BOTDR technique firstly, and makes an example of karst collapse monitoring at section K14 of highway from Guilin to Yangshuo. Discussion includes how to use this technique in underlying karst collapse monitoring in karst highway; environmental factors, like temperature and vehicle dynamic load; how to affect the monitoring results; and how to choose optical fiber type and paving region. At last, we compare the results between using BOTDR and geological radar, and conduct the safety diagnosis on the experimental road. The application achievements demonstrate that BOTDR is a viable technique for the karst collapse monitoring.

Introduction
Brilliouin Optical Time Domain Reflectometer (BOTDR) is a new-type photoelectric monitoring technique. Using spectrum technology and optical time domain measuring technology, BOTDR technique takes optical fiber as a monitor to incessantly measure the external physical parameter located along the geometric path of optical fiber and gain the changing spatial-temporal property of the measured object in order to realize the monitoring on the deformation of rock-soil body. Compared with traditional measuring methods, optical fiber can serve as conductive medium as well as sensing medium. Meanwhile, the features of optical fiber texture such as explosion proofing, anti-electromagnetic interference, resistance to corrosion, heat-resistance, small size and light weight allow the implementation of long-distance measurement, so it is more adaptive to severe environment. At present, a large number of studies have been done on structural engineering and geotechnical engineering field, both at home and abroad, and many technology inventions and significant progress have been made (Linker et al., 2009; Assaf et al., 2010). However, there are not many studies aiming at monitoring karst soil caves. The karst collapse event is uncertain and elusive in space, and abrupt in time. Traditional monitoring methods mainly use means such as groundwater dynamic conditions monitoring and regular geophysical prospecting, which cannot realize long-term regional monitoring. While the features of BOTDR technique allow the possibility of automatic, networked, and regional monitoring and early warning. Combined with the highway construction from Guilin to Yangshuo our research group started the study with BOTDR technology for monitoring and early warning of karst collapse so as to better apply BOTDR technique to karst collapse monitoring and early warning.

The basic principle of BOTDR monitoring
The BOTDR basic principle is the utilization of the linear relation between: the frequency shift change of Brillouin scattering light in optical fiber and the axial strain received by the optical fiber; and the obtainment of the axial strain of the optical fiber (Shi et al.,2005; Liu et al.,1998; Haruyoshi et al,1997; Figure 1).

The drift distance between axial strain of optical fiber and the frequency of Brillouin scattering light is showed by formula (1) below:

\[ V_B(\xi) = V_B(0) + \frac{\partial V_B(0)}{\partial \xi} \xi \] (1)
where $\Delta V_B(\varepsilon)$ is the drift distance of Brillouin scattering light frequency when strain happens in optical fiber; $\Delta V_B(0)$ is the drift distance of Brillouin scattering light frequency when there’s no strain in optical fiber; $k_{1B}/k_{2B}$ is proportionality coefficient which is about 0.5GHz/% (strain); $\varepsilon$ is the axial strain of optical fiber. Compared with other scattering lights, another prominent advantage of Brillouin scattering light is that its association between frequency shift variation and temperature is much smaller than its association between frequency shift variation and strain (0.002%/\degree C). Thus, when measuring the Brillouin frequency shift related to strain, the influence of temperature on Brillouin frequency shift is often neglected when the temperature difference is under 5\degree C.

Brillouin Optic Time Domain Reflectometer (BOTDR) is developed with optical light converging technology and Coherent detection technology, realizing the detection of Brillouin scattering light in high sensitive optical fiber with OTDR technology and Brillouin scattering technology. We can calculate the strain value of optical fiber according to the frequency variation value corresponding to the maximum intensity value of Brillouin scattering light before and after the strain on optical light. We can see from the principle that monitoring rock-soil body with BOTDR technology and placing optical fiber into monitoring object, and monitor the object comprehensively like implantation into “nervous system” (Wu et al, 2005).

BOTDR is a leading edge technology which has advantages as follows compared with traditional monitoring technology:

1) Distributed: On the end of optical fiber, we can get the information of stress, temperature, vibration and damage of any point. Regional and comprehensive monitoring of monitored object can be realized by mesh layout and decrease omission.

2) Long distance: This technology can monitor a distance of up to 80km, and optical fiber can not only serve as sensing element but also information transmission channel, which will help realize remote monitoring.

3) Adaptability: traditional geophysical engineer monitoring methods mostly use strain gage monitoring technology whose lifespan and veracity are greatly influenced by severe monitoring environment. The optical fiber is made of fiberglass, which is not only more durable than metallic material but also less disturbed by electromagnetism and thunder.

In this study, the BOTDR used is the newest AQ8603 optical strain analysis meter launched by Japanese Amtium Company. See major performance indexes in Table 1.

**Study area**
The highway from Guilin to Yangshuo stretches 66 km from Chongkou in Guilin in the north to Gaotian in Yangshuo and in the south via Lingui County, Yanshan District and Yangshuo County (Figure 2).

![Figure 1. Strain dependence of Brillouin frequency shift change](image)

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**Table 1. AQ8603(BOTDR) Technical Index**
The monitoring section is located between highway K14–K15 which is the administrative region of Lingui County. This area is hoodoo plains geomorphic type with sea level elevation between 151 to 155 m. The Quaternary soil system is slope and alluvial red clay. Soil composition is very heterogeneous and usually less than 5 m thick. The soil layer in the saturated zone is often soft and plastic, and may reach 10 m. The bedrock is mainly Devonian-age dolomitic limestone of the Guilin Group, which has well-developed rock joints, fractures and karst (Figure 3).

The research area is located nearby subterranean rivers’ discharge area where the underground waters are shallow and create seasonal wet land. The underground water is mainly karst underground water. Peak cluster area

**Figure 2. The location of the study area**

**Figure 3.** Location and geological map of the study area in Guilin City 1.limestone, 2.dolomite limestone, 3.alluvial red clay, 4.stratigraphic boundary, 5.the direction of groundwater flow, 6.the highway
and hilly area receive water from atmosphere rainfall. Apertures are rich on the surface of bed rock and karst underground water has close relation with surface water.

**Technical proposal**

Karst caves grow in this area - there were 14 caves found in merely 100 m of the monitoring section. Engineering treatment was used in construction period. In order to prevent the influence of newly developed cave to the safe operation of the highway, this section was chosen to be the monitoring and early warning section of highway roadbed caves.

The monitoring length is 90 m. The exploring application of this technology is done with the use of distributed optical fiber sensing technology to monitor and pre-warn about the roadbed caves under the highway. Through the analysis of the inbuilt environment’s temperature, vibration and other elements’ influence on monitoring results, by geological radar’s scan to verify the reliability of monitoring result and finally present a security evaluation of this monitoring section.

**The selection of optical fiber types**

Generally, optical fiber contains fiber core, cladding, coating layer, restrictive coating, etc. Different structure and external bonding method will influence the sensitivity of measurement (Liu et al., 2006). Specific to working conditions like highway roadbed construction, according to existing conditions and experiments, the choice of Corning’s single-mode tight tube optical fiber is ideal in practical application. Its major parameters are: core diameter/ cladding diameter/ tight tube diameter are 83/ 125/ 900μm; minimum tension long-term 3.0N; minimum bending radius long-term 5.0 cm; compressive strength 200N/m; operating temperature -20−80°; operating wavelength 1.3~1.5μm; refractive index difference: 0.36%; effective group refractive index: 1,4681 (1550 window).

**Laying scheme of roadbed optical fiber**

At present, there are several laying methods of optical fiber: fixed point bonding, Ω shape fixed point bonding and comprehensive fixed bonding. According to the operating condition of this project, through many experimental comparisons, the method of excavated cement mortar bonding optical fiber is chosen. Firstly excavate and lay certain thickness of cement mortar. Then pouring in proper cement mortar after optical fiber is placed in. Finally a cement beam is formed along the optical fiber and roadbed filling can be done.

Lay 4 parallel optical fibers along the roadbed. Record the locations and lengths of the optical fibers in detail and introduce them into monitoring box to supervise. Due to the severe operating environment, 4 break points appeared in the process of subgrade compaction after the optical fibers are laid. See Figure 4 for the locations.

**Study and analysis on influencing factors**

In order to study the influence of factors like seasonal temperature fluctuation on highway and vibration (traffic passing) on strain of optical fiber, dynamic measurement was conducted from 2011 at intervals of 1-2 months. Chosen GX2 optical fiber as object of study, effective monitoring length was 90m, K14+660m was corresponding to the 40m of monitoring data, K14+ 570m was corresponding to the breakpoint on 130m. Monitor

![The optical fiber flat laying distribution.](image)

*Figure 4.* The optical fiber flat laying distribution.
The influence of environment temperature on strain curve

The monitored fracture surface is located in Guilin with low latitude and mid-subtropical monsoon climate. This area has a moderate climate, abundant rainfall, long frost-free season, ample light, rich heat, long summer and short winter, four distinctive seasons and rain heat during the same period. The annual average temperature is 17.8°. The hottest month is July, in which the average temperature is 28°; the coldest month is January, in which the average temperature is 5.8°.

Relevant research (Li et al., 2013) showed that filled subgrade temperature field has a hysteretic nature compared with atmosphere temperature field. Combined with the environment condition of Guilin, we took the data collected in January as basis reference, then compared and analyzed the data collected in March and August. Figure 5 is the difference value figure of the data collected in January of GX2 optical fiber under different environment temperature conditions.

Based on the principle of distributed optical fiber sensing technology, we can see that the offset of Brillouin scattering is influenced by not only axial strain of optical fiber but also temperature. For the Corning tight tube optical fiber used in this research, the influence of the changing temperature on Brillouin scattering light frequency spectrum offset is 1.43Hz/°C, which is about 30με. The red bold line in the figure is the strain curve of the monitored optical fiber in August; the red fine line is the strain curve of the monitored optical fiber in March. The maximum strain difference is 200με; minimum difference is 10με, averagely 45.91με higher and change rate is approximately 0.005%, averagely 1.5° of temperature change.

The influence of vibration on strain curve

Different period of time has different number of vehicles passing by. According to observation on-site, 12:00-13:00 is low ebb of traffic, and 15:00-16:00 is peak time of traffic. According to the observed data of two periods in a day, we analyzed the sensitivity of BOTDR on optical fiber vibration.

Taking GX2 as research section, we addressed the initial value of the observed data in peak time and low ebb of optical fiber circuit. We placed the two treated strain curves in Figure 6 and compared them.

We can see from Figure 4 that red line is a little bit higher than black line overall, which means that the monitoring curve in peak time is higher than its counterpart in low ebb. According to the figure, between 40m and 55m, the strain is different; while after 55m, the differences of strain are minor, the two monitoring lines are almost identical. The average strain difference is 4.3με, that is to say, the vibration caused by passing vehicles has little influence on the result of optical fiber strain monitoring.

Figure 5. The different temperature environments of strain curves
Evaluation on the health of roadbed
Analyses on the operation of monitoring data
The monitored section of roadbed was laid 4 optical fiber, we took optical fiber GX2 for analysis. Long-term monitoring data shows that the monitoring of optical fiber tends to be stable after the stable period of construction (Figure 7).

The Figure below is the whole dynamic monitoring data of 2011. The data are all absolute strain value of GX2 optical fiber, of which 0-40 m is reserved section, 40-60 m is the section between monitoring box of optical fiber and 5 m underground, 60-130 m is the key monitoring section. Through the analysis, we can find that:

1) As is placed in monitoring box, 0-40 m has larger temperature fluctuation; 40-60 m is switch-in section which is influenced not only by environmental temperature because of its location in shallow ground but also temperature field change by seasonal water-level fluctuation due to its proximity to water culvert (the fluctuation of strain curve diminish from the near to the far); 60-130 is 5 m underground, which is less influenced by environmental temperature and is more identical with each other.

2) The test has shown that the strain caused by temperature under 2°C can be omitted. The average deviation of two strain curves who have the biggest discrepancy in a year is 1.5°, which can be omitted in this monitoring result.

Geological radar scanning assessment
Geological radar is a kind of geophysical prospecting using ultra high frequency electromagnetic wave to explore underground media. It can explore the developing status of karst cave down to 10 m effectively. Potential karst collapse can be prevented by regular scan.

The diagram below (Figure 8) is the geological radar exploring profile of the K14+550-K14+660 section of Guiyang highway in October 2011. Corresponding to GX2 optical fiber, the length of the scanning profile is 130m. it can be seen clearly from the diagram that there is no cave developing in this part of profile and the highway roadbed was operating safely.
The 4 optical fiber laid is under dynamic monitor by BOTDR technology. Except the seasonal entire change caused by temperature in reserved section and switch-in section, strain curves are mainly identical in key section with the overall average strain difference less than 0.03% which can be omitted. Meanwhile, by regular geological radar scanning, it is proved that there was no karst cave developing under the highway roadbed and the roadbed was healthy.

**Conclusion**

The health condition of highway roadbed is a most important index for the safe operation of roads. By laying distributed optical fiber monitoring system under test section of highway roadbed, we studied the change of strain field under roadbed with BOTDR technology. Optical fiber type and laying scheme were chosen according to on-site observation and experiments. The influence of environmental factors like temperature and vehicle vibrating load are analyzed. The result shows that the laying of distributed optical fiber monitoring system is successful and the influence of environment temperature and vehicles vibration on the monitoring result of strain field under roadbed can be omitted. Finally, we used geological radar to scan roadbed and further showed that the application of BOTDR technology is practical in roadbed health monitoring and has important application prospect.

**Figure 7.** Strain curve of optical fiber GX2

**Figure 8.** 3# measuring line(corresponding to optical fiber GX2) result of geological radar exploring
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Liu J, Li Z. 1998. SM UL TANEO measurements of strain and temperature in smart structures using fiber grating and raman scattering. Acta Aeronautica et Astronautica Sinica 19 (7(S)).
Abstract
We discuss measurements of the precursory and post-formation ground displacement in the vicinity of the Bayou Corne, Louisiana, sinkhole made using interferometric synthetic aperture radar (InSAR) and data from the L-band UAVSAR instrument. Large precursory movement was observed at the sinkhole site and shown to be predominantly horizontal in direction, in contrast to sinkhole precursors previously detected with InSAR, all of which indicated vertical deformation. Here we discuss how two opposing imaging directions were used to determine the precursory horizontal movement, and use the same technique to look at the progression of post-formation ground displacement around the expanding sinkhole during the interval 2012-2014. We find that the Bayou Corne sinkhole has expanded asymmetrically about the initial location, and show that expansion has tracked ground movement observed with InSAR along the margins of the water-filled central subsided area. This work shows that InSAR applied to images acquired from multiple directions can be used to image incipient sinkhole formation over large areas and to track the expected direction of expansion. We discuss how geologists can best use the InSAR technique to quantitatively monitor ground movement associated with sinkholes, particularly in areas where radar rapidly decorrelates, e.g., in Florida or Louisiana. These results demonstrate that InSAR could be used in sinkhole warning systems across a much broader geographical area than previously demonstrated, and for identifying both precursory and post-formation surface movement.

Introduction
The Bayou Corne sinkhole, located 50 miles south of Baton Rouge, Louisiana, formed catastrophically sometime between the late evening of 2 August 2012 and early morning of 3 August 2012, as a result of sidewall failure of a brine cavern mined near the western edge of the Napoleonville salt dome. Figure 1 shows radar images of the area both before and after sinkhole development. The Bayou Corne sinkhole formation was unusual in two ways: Firstly, the collapse occurred in the side wall of a solution-mined cavern that was located deep below the surface; the cavern extended from 1.0 km to 1.7 km depth below the surface (LDNR, 2013). Collapse of the thin side wall between the edge of the dome and the cavern interior enabled outside material to move into the empty volume of the cavern, and breached the sheath around the dome, bringing the normally isolated halite within the dome in contact with external materials, including water. This is the first reported sidewall failure within a salt dome, and because it occurred far below the surface, material in lower strata began to fill the cavern without initially causing observable movement at the surface. The second unusual feature of the Bayou Corne sinkhole is that its sudden formation, likely some months after the breach occurred, was preceded by horizontal movement at the surface (Jones and Blom, 2014), indicating that material at the surface flowed towards a chimney-like feature leading down the side of the salt dome to the breach location. The abrupt change that occurred on 2 August 2012 must have been initiated by unblocking of the path to the cavern, consistent with the change in observed quake activity at that time (Nayak and Dreger, 2014; Ellsworth et al., 2012). The Bayou Corne site is located in the Atchafalaya Basin of the Southern Mississippi River Alluvium Major Land Resource Area (MLRA) of Louisiana (Weindorf, 2008), an area with Holocene deposits. The soils in the backswamps are clayey alluvium with poor to very poor drainage and very slow permeability (Weindorf, 2008). The fact that the surface soil is largely mud is likely the reason why the direction of precursory movement did not resemble the vertical deformation reported by sinkhole researchers working in much drier locations (Paine et al., 2012; Conway and Cook, 2013; Nof et al., 2013; Rucker et al., 2013).

Previously, we reported use of InSAR performed on data acquired with the Uninhabited Aerial Vehicle Synthetic Aperture Radar (UAVSAR) to determine that the catastrophic collapse was preceded by precursory ground
movement of up to 26 cm extending across an area of radius ~250 m about the center of the initial sinkhole (Jones and Blom, 2014). The precursory movement was observed in interferograms formed from images acquired on 23 June 2011 and 2 July 2012, indicating that the ground started moving at least one month before catastrophic collapse occurred. The ground movement was evident in two different UAVSAR flight lines that imaged the Bayou Corne area from opposite directions. With two interferograms from different look directions, we could determine the 2-dimensional vector deformation of the surface, as explained in more detail below. The two interferograms were used to resolve horizontal and vertical precursory movement, showing that the precursory movement was predominantly horizontal. Clays deform easily when wet, and can flow either horizontally or vertically. At the Bayou Corne sinkhole site, the horizontal flow was observed over a much larger area than the initial sinkhole (Figure 2), so vertical surface change almost certainly occurred but was below the level detectible with our instrument.

In the sections that follow we discuss the data and methods used for pre-event and post-formation sinkhole monitoring at the Bayou Corne site, explain why the observed precursory displacement was interpreted as being horizontal in direction, compare the post-formation interferograms to ground survey sinkhole plan measurements available through the Louisiana Dept. of Natural Resources (LDNR), and discuss the most recent InSAR results which cover the period after the April 2014 ground survey.

**Method**

Deformation measurements using InSAR require two or more synthetic aperture radar images acquired at different times. Image pairs can be analyzed via interferometry to determine the difference in the signal phase between the two acquisitions, which relates to the amount of surface deformation that occurred if stringent observational criteria are met and if the surface did not undergo significant decorrelation during the interval between image acquisitions.

Interferometry can measure changes in distance between the platform and surface at cm to mm scales, thus measuring elevation changes. For side-looking SAR instruments the measured displacement can be a combination of horizontal and vertical motion. Principal advantages of InSAR over point techniques such as GPS are geographically complete regional coverage and ability to measure small changes in elevation over time using a time series of images. Principal disadvantages include data availability and processing issues, and loss of ability

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**Figure 1.** Radar intensity images acquired with the UAVSAR synthetic aperture radar on 2 July 2012 (upper), one month before the Bayou Corne sinkhole formed, and on 28 October 2014 (lower), nearly 26 months after the sinkhole first appeared on the surface. The star in the lower image marks the location of the wellhead connected to the failed cavern and the ‘X’ marks the location used as the zero phase reference for the interferograms in Figure 4. In these images, the town of Bayou Corne is at the upper left and the Napoleonville Salt Dome extends to the east of the failed cavern.
to create interferograms due to surface changes in roughness or water content between observations (temporal decorrelation). Longer wavelength radars, like the L-band UAVSAR (23.8 cm), experience less impact from decorrelation than shorter wavelength SAR instruments.

The UAVSAR data used in the work reported here was acquired over the Bayou Corne area in eight deployments dating from June 2011 to October 2014. Interferograms formed from sequential acquisitions of two UAVSAR flight lines (IDs 14013 (looking to the NE) and 32018 (looking to the SW)) were used in this analysis. The specific dates of acquisition are given in Table 1. The earliest data used was acquired on 23 June 2011, over a year before the sinkhole formed; second set was acquired on 2 July 2012, one month before surface collapse; and the subsequent collections occurred 2-3 times per year through 2014.

Figure 1 shows the UAVSAR radar intensity images of the Bayou Corne area around the sinkhole site on 2 July 2012, a month before the sinkhole formed and at the time when large precursory movement was evident, and on 28 October 2014, the date of the last acquisition reported here. The location of the failed cavern is indicated in this figure and the sinkhole lies to the northwest, just beyond the edge of the Napoleonville salt dome. In the earlier acquisition, the site where the sinkhole formed (approximately centered on the water filled area in the later image) is not distinguishable from the background area. In July 2012, the backscattered intensity from the surface in that area is consistent with scattering from soil, not still water, which would be radar dark as in the October 2014 image.

Application of InSAR techniques in agricultural areas and other areas with highly variable vegetation and soil moisture is notoriously difficult. Certainly, on the face of it, the Bayou Corne site seems a poor candidate for InSAR. However, we were able to successfully form interferograms in large part because we used L-band radar, which has a longer wavelength (24 cm) than the contemporaneously operating satellite SARs, which at X- and C-band (3.3 cm and 5.5 cm, respectively) could not obtain coherent images over the Bayou Corne site. In addition, the ground movement at the sinkhole site was very large, and hence was visible over the instrument and speckle noise and the phase variance caused by temporal decorrelation. Furthermore, we consider it likely that double bounce scattering from ground-to-tree-to-antenna, with the trees being stable scattering surfaces, contributed to our unexpected success in obtaining coherent InSAR phase across many months.
Results
Sinkhole Precursory Movement
We used two 375-day temporal baseline (23 June 2011 – 2 July 2012) interferograms that imaged the area at an incident angle of 60 degrees but from opposite directions to look for precursory surface movement. Figure 2 shows a photograph of the area acquired during that time interval (on 16 Nov 2011) prior to the sinkhole formation indicating the location of the well head attached to the cavern that failed, and one of the interferograms showing precursory deformation covering the same area. The interferogram was multilooked by a factor of 3 x12 during UAVSAR processing, georeferenced to 6 m pixel spacing, then smoothed with a 3 x 3 boxcar filter. Precursory deformation is seen extending across an area ~500 m across. The deformation shows a distinctive 2-lobe pattern, which indicates movement of the surface away from the radar in the nearer lobe and towards the radar in the further lobe. Because InSAR only measures a change in distance in the line-of-sight direction, we cannot resolve horizontal and vertical motion from a single interferogram.

Figure 3 shows the line-of-sight movement obtained from the two interferograms, one looking at the sinkhole from the northeast and the other looking from the southwest. Because we had two interferograms we were able to resolve vertical movement from one dimension of the horizontal movement, specifically that in the direction of the projection of the line-of-sight vector on the surface. The striking double-lobe pattern is apparent in both interferograms and the direction of change in the line of sight distance differs in sign with roughly equal magnitude when measured from the two opposite directions. The cartoon at the bottom of Figure 3 shows how this indicates that the movement was primarily horizontal, with inflow towards the center of the pattern. We note that this observed pattern (increasing distance from one observer, with decreasing distance to an observer looking from the opposite direction) precludes the possibility that the observed signal was due to a change in the level of standing water on the surface, which would have had the same sign for both observations. Without another perpendicular imaging geometry, it is not possible to determine through InSAR measurement both dimensions of the horizontal movement. InSAR cannot measure displacement perpendicular to the line-of-sight direction, which is why the horizontal displacement shows the 2-lobe pattern: InSAR cannot capture radial horizontal movement without a 3rd image looking perpendicular to the two that we have. However, radial inflow towards a sink location is likely to have occurred.

Sinkhole Progression
The size of the Bayou Corne sinkhole, where the sinkhole proper is defined as the area with standing water depth > 10 ft (3 m), has been tracked by LDNR since its formation. The sinkhole, initially ~10,000 m² in surface extent, had expanded by 15 April 2014 to ~130,000 m², with >223,000 m² of area with water depth > 2 ft (0.6 m), designated the “sinkhole plus subsided area” in the surveys of the area supplied to the LDNR and posted on their Bayou Corne incident website (http://dnr.louisiana.gov/index.cfm?md=pagebuilder&tmpl=home&p id=939).

We used the 15 April 2014 Bayou Corne sinkhole survey (Louisiana Department of Natural Resources, 2014), which showed the full sinkhole extent on several dates between 4 Oct 2012 and 15 April 2014, to determine the direction and timing of expansion of the Bayou Corne sinkhole for comparison to the UAVSAR post-sinkhole-

<table>
<thead>
<tr>
<th>Date</th>
<th>Days to Sinkhole Formation</th>
<th>Image Interval (Temporal Baseline) [days]</th>
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<tbody>
<tr>
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<td>na</td>
</tr>
<tr>
<td>2 July 2012</td>
<td>T−32</td>
<td>375</td>
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<tr>
<td>26 Oct 2012</td>
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<td>24 July 2013</td>
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<td>97</td>
</tr>
<tr>
<td>9 Apr 2014</td>
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<td>162</td>
</tr>
<tr>
<td>28 Oct 2014</td>
<td>T+816</td>
<td>202</td>
</tr>
</tbody>
</table>

Table 1. UAVSAR Acquisitions– UAVSAR acquisition dates, time between acquisition and sinkhole formation, and the temporal baseline between the SAR images used to form sequential interferograms.
formation interferograms. For the ground surveys, depth measurements were taken with sounding rods to determine the sinkhole depth. No earlier or later surveys were available from the LDNR website, so the ground measurements cover a subset of the time period covered by the UAVSAR data set.

The interferograms formed from sequential acquisitions (dates in Table 1) of the UAVSAR flight line viewing the region from the southwest are shown in Figure 4. These are plotted in radar coordinates (slant range, also known as line-of-sight distance, vs. azimuth, or flight track position) and cover the outlined region in the intensity images of Figure 1. The interferograms are multilooked a factor of 3 \( \times \) 12 (slant range x azimuth) and have pixel size of 7 m x 7 m. The location of the zero phase reference used is outside of the plotted area and indicated by an ‘X’ in Figure 1. The center of the initial sinkhole that formed on 3 Aug 2012 is indicated on each interferogram.

Once the sinkhole formed, the open water areas have very little radar backscatter (dark on interferogram) and InSAR cannot be performed. In areas surrounding the open water, much of the area experiences too much temporal decorrelation to be used for InSAR (phase changes randomly from pixel to pixel, giving a speckled look to the color figures). This information is useful in delimiting the extent of major surface change between the two acquisitions forming the interferogram. In addition, along the margin in some areas the phase remains coherent and shows fringed pattern indicating coherent surface movement. In the earlier interferograms (1, 2, and 3 in particular) the interferograms show some double-lobed characteristics so it is clear that surface flow towards the sink continued after sinkhole formation, although it was not as clearly dominant as in the precursory ground movement.

Based on the ground surveys, between 4 Oct 2012 and 15 Aug 2013 the sinkhole expanded in all directions, with the largest expansion to the SSW and more expansion to the west than to the east or north. Interferograms 2 and 3 most closely cover this time period. The deformation patterns in interferograms 1, 2, and 3 of Figure 4 all show greater extent of the deformation pattern to the SW, with the largest extent in the same direction as measured in the surveys. By 24 July 2013, the end date of interferogram 3, the deformation to the SW had decreased. Between 15 Aug 2013 and 15 April 2014 the ground surveys measured an abrupt slowing of expansion to the SSW and continued expansion to the north and east, with the largest expansion in the NE quadrant and the next largest expansion to the NW. This time pe-

Figure 3. Precursory deformation derived from the two UAVSAR flight lines imaging the Bayou Corne site. The upper (middle) image shows movement in the line-of-sight direction for a look direction pointing to the NE (SW). Positive values (red) indicate an increase in distance, i.e., movement away from the radar. A point on the ground appears to move towards one direction and away from the other. The cartoon at the bottom shows how this pattern of sign change corresponds to inward flow towards the center between the two lobes.
Figure 4. The six post-sinkhole-formation interferograms from UAVSAR line ID 14013 covering the region outlined in white in Figure 1, formed from sequential UAVSAR images of the Bayou Corne area acquired during the period from 2 July 2012 to 28 Oct 2014 (Table 1). The track and look directions are indicated on interferogram 1, which also shows the location of the nearby intact caverns within the salt dome as lying below the white outlined area. The center of the original sinkhole that formed in Aug. 2012 is indicated with a star on all interferograms.
period best overlaps with the UAVSAR interferograms 4 and 5 in Figure 4. These show a much reduced deformation area with few clear fringes to indicate the location of coherent ground movement. Exceptions are interferogram 4, which shows a signal along the west edge of the open water area, and interferogram 5, which shows fringes along the shore in the NE quadrant. These results generally agree with the expansion directions measured in ground surveys.

The last interferogram (6) in Figure 4 shows change during the interval from 9 April 2014 to 28 October 2014. There is no indication from the measurements of significant movement in the vicinity of the nearby, intact caverns, whose wellheads lie within the white outlined area shown in interferogram 1 of Figure 4. Bright targets in this area remain stable in all interferograms, including the latest. If failure of a nearby cavern were preceded by horizontal flow, it would be evident in this interferogram. This latest data indicates that sloughing to the NE between the open water of the sinkhole and the nearby wellheads continues but at a reduced rate. In addition, inflow from the SE continues with potential expansion of the sinkhole to the east of the wellpad of the failed cavern.

Discussion

In this paper, we discuss the precursory and post-formation ground movement in the vicinity of the Bayou Corne sinkhole and show the pattern of expansion and the direction of ground movement around the edges of the sinkhole. Using radar intensity images acquired before and after sinkhole formation, we show that the precursory ground deformation was not apparent from the intensity images alone. Using interferograms acquired from opposing imaging directions, we show that the precursory movement was horizontal in direction and could not have been an artifact of water level change for two reasons: (1) Water level change would not show an increased level in one area (one lobe in observed pattern) immediately adjacent to an area with decreased water level (other lobe) of pattern; and (2) water level change would have caused an area on the ground to have the same change in line-of-sight distance from the two imaging directions.

The interferograms of the area during the period of sinkhole expansion show a radar-dark open water area surrounded by an incoherent area which generally has a coherent border. In the first year post-formation, inflow into the sinkhole came from the direction away from the Napoleonville salt dome, consistent with flow of the sediments into a chimney or disturbed rock layer leading to the cavern. During the period 24 July 2013 to 28 October 2014, the size of the incoherent area around the sinkhole steadily decreased. The direction of maximum movement derived from the interferograms generally matches sinkhole expansion measured with sounding rods for the State of Louisiana and posted on the LDNR Bayou Corne incident website. In the last interferogram (Interferogram 6 of Figure 4), documenting change between 9 April 2014 and 28 October 2014, there is little incoherent area around the open water and indications that slow movement is still occurring. There is an extended region around the sinkhole (yellow in image) showing change contained within the borders of a levee built to isolate the sinkhole from the Bayou Corne waterway, the nearby highway, and the outskirts of the town. Without further information, we cannot distinguish whether this signal is water level change or vertical surface movement.

This work shows that sinkhole monitoring with InSAR is possible even in areas with wet soil and significant vegetation. The major sinkhole that formed near Bayou Corne, Louisiana, in August 2012 was the result of mine collapse, unlike the sinkholes that occur in karst. Nonetheless, the methods of identifying and monitoring the Bayou Corne sinkhole have general applicability to all types of sinkholes. For example, the natural environment of Florida has much more in common with Bayou Corne than with Holbrook, AZ, or the Dead Sea in Israel, both arid areas where vertical deformation associated with sinkholes have been observed [Conway and Cook, 2013; Nof et al., 2014]. In areas with similar sediments and soil conditions to the Louisiana site, horizontal movement could precede collapse as an indicator of potential ground failure. Using L-band SAR (i.e., UAVSAR or ALOS-PALSAR), it is possible to detect movement at least in some locations that would not maintain temporal coherence nearly as long if using shorter wavelength SAR instruments, e.g., at X-band (TerraSAR-X, COSMO-Skymed) or C-band (Radarsat, Sentinel). However, even at L-band, the cumulative movement had to be large (several centimeters or greater) to be apparent against the background decorrelation in these InSAR-challenging areas. We had better success using a longer wavelength SAR and longer intervals between acquisition rather than shorter wavelengths and shorter intervals because there was less movement over the shorter intervals and, for the shorter wavelength SARs, too much temporal decorrelation occurred for the small ground movement to be visible above the phase variance from noise sources.

The InSAR technique can measure surface deformation from satellites or aircraft over time as a series of snapshots. Proper analysis permits detection and monitoring of surface deformation, including that from developing
sinkholes. Radar image swaths can typically cover hundreds of square kilometers, and thus have the distinct advantage of being able to cover very large areas. InSAR detection of surface deformation due to sinkhole development has now been documented in multiple studies, thus showing promise for incorporation of the method as one part of a sinkhole warning system.

Acknowledgements
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References


THE APPLICATION OF PASSIVE SEISMIC TECHNIQUES TO THE DETECTION OF BURIED HOLLOWS

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Abstract
Pilot studies involving the use of passive seismic techniques in a range of geological settings and applications, e.g., mapping bedrock, studies of soil erosion and Quaternary mapping have shown that it is a versatile, non-invasive and economic technique. This paper presents the findings of three case studies that trialled the use of passive seismic techniques for the detection and characterisation of buried hollows in carbonate rocks, comprising: i) a buried hollow in the Cretaceous Chalk at Ashford Hill in the Kennet Valley, a tributary of the River Thames, UK; ii) buried karst in the foundation excavations for wind turbines in Carboniferous Limestone at Brasington, Wirksworth, Derbyshire, UK, and iii) defining the extent of solution hollows that host terrestrial Miocene deposits, near Friden, Newhaven, Derbyshire, UK. Whilst case studies ii) and iii) are focused on areas of buried dolines, the geological context of the Ashford Hill site is more complex; comprising a deformation hollow with an uplifted “pinnacle” of chalk bedrock at the centre. The data were collected using a (Tromino), a three-component, broadband seismometer to measure background ambient noise (microtremors induced by wind, ocean waves, industrial machinery, road and rail traffic, etc.). The Tromino is small, portable with an operating range of 0.1 Hz to 1,024 Hz and interpreted using proprietary software (Grilla), which subjects the data to Fourier transformation and smoothing. Where possible, estimated shear wave velocities used in the Grilla Software modelling, based on peaks identified on the H/V spectrum, have been calibrated using...
borehole data or parallel geophysical techniques. In each case, the karst features were defined by Nakamura’s horizontal to vertical (H/V) spectral ratio technique, where microtremors are converted to show impedance contrasts (velocity x density), or a pseudo layered seismic stratigraphy of the near surface along each profile. An additional benefit of the use of this technique is its depth of penetration and potential for defining the structural and lithological context of the hollows, thereby contributing to the process understanding associated with their formation. To this end the technique has helped define discontinuity (fault, joint or bedding) guidance of the hollows.

Introduction
In 2012 the British Geological Survey (BGS) acquired a modern passive seismic measurement unit (Tromino) to complement their existing suite of geophysical equipment. During the subsequent two years, this equipment has been tested in a range of geological and geomorphological settings. The Tromino measures passive seismic (acoustic, elastic wave) noise, which comprises both natural (<1 Hz) and man-made (>1 Hz) frequencies, and extracts information from it through time domain and spectral techniques, i.e., HVSR (horizontal to vertical spectral ratio, Nakamura, 1989). The ratio of the averaged H/V frequency spectrum is used to determine the fundamental site resonance frequency and interpreted using regression equations to estimate the thickness and depth to bedrock. The instrument comprises a single, light-weight, mobile seismometer incorporating a compact 3-component electrodynamic sensor (velocimeter) with an operating frequency of between 0.1 Hz and 1,024 Hz, above which, anelastic absorption resulting from rock internal friction causes strong attenuation (Castellaro et al., 2005). One advantage of the Tromino is the efficiency with which data can be collected. The duration of recording is scheduled in accordance with the lowest frequency of interest allowing for at least 10 repetitions to achieve stability and further time to allow for the variation in resonance with time. In practice this requires a recording time of 10 – 15 minutes (Micromed s.p.a., 2009), but enables significant ground coverage within a day. The Tromino’s operation is based on guidelines and recommendations made by the Site Effects Assessment using Ambient Excitations project (SESAME European Research Project, 2004). The instrument is placed on natural ground (artificial surfaces are more rigid and dampen the H/V curve) and orientated parallel or orthogonal to linear features and structures, because of the polarization or directional resonance effects on ground motion. The technique is applicable where a soft layer overlies a hard substratum (e.g., Amorosi et al., 2008; Grippa et al., 2011). Interpretation of the results requires an understanding of a number of influential factors, e.g., the potential for trapped waves (waves without vertical penetration due to reflection); the increased complexity of the seismic response to irregular geometry in the basement deposit, or the presence of faults, cavities and irregularities in the topography (Panzera et al., 2013). Recorded data returns from the Tromino are analysed using Grilla software, which subjects the data to Fourier transformation and smoothing (Triangular Window method, Micromed s.p.a) to derive a plot of frequency against H/V, e.g., Figure 1. Interpretation of the results requires a measure of “shear-wave velocities”, which can be obtained from the literature or from laboratory testing. Here we present three case studies undertaken in the context of karst geohazard research in the BGS.

Case Study 1: Detection of a Buried Hollow at Ashford Hill, Hampshire
Buried hollows encountered in the Chalk can be a hazard for construction in the London region where they comprise closed depressions that are associated with ground deformation and may be infilled with softened and/ or an unpredictable range of sediments (e.g., Banks et al., 2015). Some of the buried hollows in London extend deep into the bedrock geology and are in-filled with disturbed superficial deposits and reworked bedrock. They can be up to 500m wide and more than 60m deep. It is suspected that a “chalk pinnacle” (Hawkins, 1952) at Ashford Hill, approximately 100km to the south-west of central London, provides a rural analogue for the London hollows. Consequently it became the focus for comparative Tromino and electrical resistivity geophysical surveys (Figures 2, 3, and 4) in a quest to identify successful non-intrusive tools for the identification of similar features elsewhere.
The isolated pinnacle of Newhaven Chalk (Cretaceous White Chalk Group) present at Ashford Hill is known to rise in the order of 50m above its normal level. It is immediately surrounded by the stratigraphically younger Eocene Lambeth Group (interbedded shelly clays, clays, silty clays and silts that are underlain by black pebbly sands). The Lambeth Group strata (locally referred to as the Reading Beds) are overlain by the London Clay (Palaeocene Thames Group). This feature occurs in the buried valley of a tributary (Baughurst Stream) to the River Enbourne, itself a tributary of the River Kennet. In this setting alternative hypotheses were postulated with respect to the mode of formation, the principal ones being either as a consequence of confluence scour or due to syn-sedimentary subsidence (Collins et al., 1996). Within the valley, the bedrock geology is overlain by river terrace deposits and alluvium with patches of head deposits along the valley sides. The area is subject to flooding during periods of high groundwater levels (rising in the chalk) and the centre of the feature is largely covered by peat associated with a pool of standing water for much of the year. This unique setting warranted designation as a Site of Special Scientific Interest (SSSI) by Natural England in 1986. A series of boreholes was undertaken in the 1940s. The majority penetrated the London Clay and extended approximately 3m into the underlying Lambeth Group which meant that the extent of the “pinnacle” was not fully constrained (Hawkins, 1952). The grid of boreholes formed the basis for the schedule of monitoring points for the Tromino (Figures 2 to 4) approximately 85m apart. The electrical resistivity was undertaken along two lines.

Figure 2. Case study 1: Context for Ashford Hill Tromino Section 3 showing geology and borehole grid (BGS, 2000). BGS©NERC. Contains OS Open data ©Crown Copyright and database rights, 2015.

Figure 3. Case study 1: Ashford Hill Tromino Section 3 plot of impedance (H/V) against depth. BGS©NERC. 2015.
Excavation of the foundations for four wind turbines at Carsington Pasture, near Wirksworth, in the Derbyshire Peak District, UK exposed buried, sediment filled hollows in the bedrock. The hollows are attributed to a form of karst that results from hypogene processes on a carbonate platform edge. The bedrock geology (Aitkenhead et al., 1985) comprises dolomitised Carboniferous limestones that have been subject to lead-zinc-barite mineralization as a consequence of the expulsion of fluids from over-pressurized mudstones that occupied the adjacent Widmerpool Gulf (Figure 5). The carbonate sediments are interbedded with thin layers of volcanic dust that have weathered to clay (wayboards). Infill sediments comprised variable mixes of vari-coloured silty clay, sand, sand and gravel and clayey sand (Jones and Banks, 2014).

Excavation of the foundations commenced on May 8, 2012. Difficult ground conditions were encountered that arranged approximately north-south and northwest-south-east across the “pinnacle” (Figure 4). The H/V 0.4 contour and the low impedance characterising the organic alluvial soils. The “pinnacle” is mapped as an ovoid shape of approximately 120m along the north-west axis and 80m along the north-east axis on the 1: 50 000 scale geological map. Tromino survey sections 2 and 3 (Figure 3) suggest that the feature is approximately 100m by 40m. Comparison of the Tromino results with the electrical resistivity profiles (Figure 4) shows the benefits of using multiple techniques in geophysics, as each technique measures different properties. Whilst the Tromino provides valuable information with respect to the bedrock boundaries, the electrical resistivity provides good resolution in the near surface and is strongly related to the moisture content of the ground, e.g. the low resistivity associated with the high moisture contents in the vicinity of the chalk “pinnacle” towards the north-west end of line 2 and the northern end of line 1 (Figure 4).
necessitated remedial engineering measures and delayed the project by 12 to 14 months. This provided a time window within which it was possible to trial the Tromino in another karst setting.

At Carsington Pasture, Tromino surveys were carried out across two of the turbine foundation excavations (T1 and T2). Here we describe the profile across the excavation for T1. The BGS used the Tromino to measure the seismic noise at 5 m intervals along an approximately northerly alignment across the excavation for the foundation of Turbine 1 (T1, Figure 5). Recording times were limited to 8 minutes and ground elevations were determined using conventional levelling techniques. The configuration of the impedance contrasts (Figure 6) suggests lithological guidance on the distribution of the higher impedance ground (non-dolomitised limestone with a lower porosity) with displacement being indicative of the presence of a fault, as indicated by the subvertical linear zone of low impedance (Figure 6). This may be sediment filled and appears to be dissolutionally enlarged, particularly at a depth of about 25m. Although there was no borehole evidence to validate these findings, their value was described by the Resident Engineer for the project as being representative of what was actually encountered during excavation for the remedial engineering measures (10m cut from the near surface zone of low impedance towards the southern end of the section and fill with imported crushed limestone).

Panzera et al. (2013) concluded that it is not possible to identify a unique spectral response to cavities. It was noted that the response (amplification or deamplification of the vertical component of motion) appears to be influenced by cavity size such that only cavities having a height of greater than about 4m show significant H/V spectral peaks. Intrusive investigation would be required to validate the extent of the dissolutional enlargement evident in Figure 6.

Case Study 3: Enhancing Geological Knowledge Associated with the Buried Hollows that Host Terrestrial Miocene Deposits in the Peak District, UK

Terrestrial Miocene deposits are relatively rare in the UK and their preservation in deep hollows in the dolomitised limestone of the Peak District Carboniferous platform is of significant scientific interest in terms of the associated flora and implications for understanding the climatic

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**Figure 5.** Case study 2: geological context for the Carsington Pasture survey (BGS, 1983). BGS©NERC. Contains OS Open data ©Crown Copyright and database rights, 2015.

**Figure 6.** Impedance (H/V) against depth for northerly transect across T1. BGS©NERC. 2015.
again suggests lithological guidance on the distribution of the higher impedance non-dolomitised limestone. In this profile a subvertical zone of high impedance is suspected to be indicative of a mineral vein, as borne out by the occurrence of barite in borehole DP1. This borehole encountered a void at 22.65m depth, which can also be identified in the impedance plot (Figure 8).

Conclusions and Potential for Further Development in Karst Environments

Each of the three case studies demonstrates the value of using passive seismic techniques to establish the elevation of bedrock, which is particularly beneficial in karst terrains where knowledge of the depth to bedrock is critical for engineering design. The case studies include both softer Cretaceous Chalk and Palaeozoic limestone bedrocks. Case studies 2 and 3 indicate that passive seismic techniques appear to provide an efficient means of discriminating between lower impedance dolomitised and higher impedance non-dolomitised limestones in the platform carbonates in the Southern Peak District, UK. Faults and mineral veins have also been successfully identified using the Tromino. The implication is that this would be an effective technique for investigating the extent of buried karst in areas scheduled for development, particularly given the efficiency with which data can be collected. A useful extension of this project would be i) the trialling of the application of passive seismic techniques to characterise artificially filled buried hollows, e.g., landfill, and ii) the application of more closely spaced monitoring to

and environmental conditions at the time of deposition (Boulter et al., 1971; Pound et al., 2012). These deposits were formerly exploited for refractory minerals and many worked out excavations occur as a series of remnant depressions. Two depressions at Kenslow, near Friden in the Peak District, UK (Figure 7) became the focus for trialling the Tromino in a karst setting. At this location the bedrock geology comprises dolomitised limestone (Asbian age Bee Low Limestone Formation of the Peak Limestone Group). A series of borings that had previously been undertaken by the minerals company (borehole logs available via the BGS national geoscience data centre) provided a target for ground truthing of the larger pit.

The passive seismic noise was monitored at approximately 20m centres, coincident with the former line of boreholes at the base of the pit. Recording times were limited to 8 minutes. The impedance plots were generated using a Vs value of 225m/second, a typical value for sand. The configuration of the impedance contrasts (Figure 8)
provide data to help constrain the processes associated with karstification. For each of the case studies, it is considered that the laboratory determination of shear velocities offers the potential for better modelling and resolution of depth in the superficial deposits.

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References
USING ELECTRICAL RESISTIVITY IMAGING TO CHARACTERIZE KARST HAZARDS IN SOUTHEASTERN MINNESOTA AGRICULTURAL SETTINGS

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Abstract

Much of the Driftless Area of southeastern Minnesota is underlain by karstified carbonate bedrock. Land use in this karst terrain is dominated by agriculture, including row crop and dairy operations. The karst in this region is often mantled with up to 15 m of soil and unconsolidated sediments. As a result, underlying karst hazards such as incipient sinkholes are often hidden until they are suddenly revealed by the collapse of subsurface voids.

Regionally, the economics of the dairy industry is causing a trend toward the consolidation and expansion of existing operations. As concentrated animal feeding operations (CAFO) or feedlots expand, state and local agencies are charged with enforcing regulations designed to protect environmental and water resources in agricultural areas. One of the key challenges in reviewing and siting expanded dairies is identifying potential karst hazards, particularly where they might undermine manure storage facilities or where they occur on crop-lands where manure is applied. Uncertainty about the location of karst hazards relative to proposed feedlot facilities is one of the reasons that feedlot expansions in the Driftless Area are often controversial. Recently, the Minnesota Pollution Control Agency enacted strict guidelines that severely limit bedrock removal in order to facilitate the construction of manure storage containment facilities. The purpose of this rule is to ensure that a minimum separation is maintained between intact bedrock and the containment liner to provide the opportunity for attenuation within the soil of contaminants that could potentially leak if the containment structure is compromised.

Electrical Resistivity Imaging (ERI) techniques have been employed to screen for karst hazards during the planning phase of feedlot expansions, and where present, to more accurately characterize the nature of the karst hazard. Because depth-to-bedrock is highly variable in the karst terrain of southeastern Minnesota, ERI has also been a useful tool to characterize this spatial variation under proposed manure containment sites. In this study, ERI was performed using a 56-channel AGI Supersting™ system with post-processing of the data in EarthImager™ software. Dipole-Dipole and Wenner electrical resistivity arrays have been the most useful for identifying karst hazards. Electrode spacing of 3 to 5 m has provided a good balance between depth-of-image and the spatial resolution necessary to locate and identify karst hazards. Soil boring data, which is typically collected during pre-construction site investigations, is critical to the interpretation of ERI data. Although individual sites vary, most surface materials in southeastern Minnesota have resistivities that fall within predictable ranges: 20-80 ohm-m for soils, 80-100 ohm-m for epikarst and weathered residuum, and >100 ohm-m for bedrock. Karst voids in the subsurface typically display resistivities greater than 1000 ohm-m, providing good contrast with the resistivities of the surrounding bedrock.

ERI has been an effective tool in identifying karst hazards in agricultural settings of southeastern Minnesota. In addition to improving pre-construction site assessments, ERI has also helped to reduce potential controversy surrounding the karst hazards of proposed projects by providing more certainty about the underlying geology.
THE COST OF KARST SUBSIDENCE AND SINKHOLE COLLAPSE IN THE UNITED STATES COMPARED WITH OTHER NATURAL HAZARDS

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Abstract
Rocks with potential for karst formation are found in all 50 states. Damage due to karst subsidence and sinkhole collapse is a natural hazard of national scope. Repair of damage to buildings, highways, and other infrastructure represents a significant national cost. Sparse and incomplete data show that the average cost of karst-related damages in the United States over the last 15 years is estimated to be at least $300,000,000 per year and the actual total is probably much higher. This estimate is lower than the estimated annual costs for other natural hazards; flooding, hurricanes and cyclonic storms, tornadoes, landslides, earthquakes, or wildfires all of which average over $1 billion per year. Very few state organizations track karst subsidence and sinkhole damage mitigation costs; none occurs at the federal level. Many states discuss the karst hazard in their state hazard mitigation plans, but seldom include detailed reports of subsidence incidents or their mitigation costs. Most state highway departments do not differentiate karst subsidence or sinkhole collapse from other road repair costs. Amassing of these data would raise the estimated annual cost considerably. Information from insurance organizations about sinkhole damage claims and payouts is also not readily available. Currently there is no agency with a mandate for developing such data. If a more realistic estimate could be made, it would illuminate the national scope of this hazard and make comparison with costs of other natural hazards more realistic.

While fatalities, or even injuries, are rare, karst subsidence and sinkhole collapses damage man-made structures and cost the nation many millions of dollars each year. In addition, land values in sinkhole-prone areas can become depressed, impacting not just land owners, but also counties and municipalities that are dependent on real estate and property tax revenue.

Unlike other natural hazards, sinkhole collapse is generally not dependent on extreme weather events and can occur, sporadically, across the country every year. Man-made infrastructure and buildings and transportation arteries have expanded onto karst terrain that was formerly rural and sparsely developed. Features such as water supply pipes, storm drains, and sewers, already built over karst, are failing at higher rates as they age, particularly if they have not been maintained.

Introduction
Karst subsidence in the United States, particularly catastrophic sinkhole collapse, is a significant natural hazard with national scope. Although the potential for subsidence or sinkhole occurrence is aerially variable, areas underlain by relatively soluble carbonate and evaporite rocks exist in all 50 states (Figure 1).

Sinkhole collapse tends to occur more often in the eastern part of the country, where there is generally higher rainfall and where local geologic settings are conducive to formation of cover-collapse sinkholes, such as soluble rocks overlain by variable thicknesses of sediment or soil. The New England states are less prone to sinkholes because many sinkholes which may have formed in the bedrock surface have been removed by glacial scouring.

State and local governments may be reluctant to track and report sinkhole damages, because such information may lead to reduced property values and a reduction in the property tax base.
There is a growing awareness among national natural hazard management agencies, insurance organizations, highway departments, and karst scientists that the costs of the karst hazard, in terms of property damage, is significant. There has been, however, little political or institutional support for data collection, scientific studies, or mitigation programs partly because there is very little data documenting the totality of karst subsidence and sinkhole costs at national, state, or even municipal levels. Also, the karst hazard tends to be manifested by scattered and sporadic individual events that are local in scope. Although karst collapse is sometimes very costly locally, they often fail to reach thresholds where they become worrisome to the general public and their civic leaders. Although there have been some spectacular, but limited, collapse incidents, there has never been a presidentially declared sinkhole disaster. In aggregate, however, the costs induced by sinkhole collapses across the United States are significant and may be large enough to compare with other natural hazards. States that are aware of their sinkhole hazard tend to keep incomplete records of karst incidents.

The purpose of this paper is to attempt to outline an estimate of the cost of karst subsidence and sinkhole hazard in the United States and to emphasize the lack of hard data available at almost all scales. The intent of this paper was to try and describe the problem and elucidate the need for data, and not to create an exhaustive national database.

Definitions
Impediments to public and official acknowledgement of the true magnitude of karst hazards include a lack of understanding of the processes involved, as well as non-standard definitions of collapse features. It is common for the news media to report any type of collapse, regardless of cause, as a sinkhole. More often than not these turn out to be collapses in man-made fill and are caused by leaking pipes or drains rather than related to karst processes. Some reports even describe highway potholes, caused by separation of paving materials by freeze-thaw and other processes, as sinkholes. In the context of this study, the term sinkhole refers only to those depressions that result from natural karstic processes.
Karst
The term karst has traditionally been used to refer to regions of exposed or only shallowly buried soluble bedrock with an abundance of surface landforms, such as sinkholes, sinking streams, springs that reflect the presence of subsurface voids (caves) (Ford and Williams, 2007). All 50 of the United States as well as most of the territories and islands contain areas of karst (Figure 1). About 18% of the ground surface of the United States is underlain by soluble rocks and sediments with potential for sinkhole development (Weary and Doctor, 2014).

Subsidence
Subsidence is the lowering of the ground surface, either as a gradual and slow process or as a sudden and rapid collapse. Karst subsidence of broad areas is caused by karstic processes of dissolution at the surface or in the subsurface and is so slow that it generally does not constitute a threat to the health and wellbeing of people and animals. This subsidence is often so slow that it does not affect structures, although in some cases it can cause foundation cracking and tilting. There are other kinds of subsidence associated with groundwater withdrawal, mining, and other activities, but they are not addressed in this report.

Sinkholes
Sinkholes are closed topographic depressions caused by a lowering of the earth surface by dissolution of the bedrock or by collapse of the surface into a void produced by solution or removal of subterranean materials. A karst sinkhole is produced by natural processes of solution of the bedrock, sometimes followed by collapse of overlying sediments or soil. For the remainder of this paper, the term ‘sinkhole’ will refer only to karst sinkholes.

Sinkholes may be divided into several types based on their morphology and the processes involved in their formation. These sinkhole types are: (1) dissolution, (2) cover-subsidence, (3) cover-collapse, and (4) bedrock collapse. See Galloway et al. (1999) for simple descriptions of the first three types. The fourth type, bedrock collapse, is rare and not discussed in that publication or in this report. Most injury and damage resulting from karst subsidence in the U.S. is caused by the cover-collapse type of sinkholes. This is because these sinkholes are relatively common, they often occur without warning as sudden ground failures, and since they occur where there is a soil or sediment mantle over the bedrock their precise locations are difficult to anticipate.

Sinkholes can form over various periods of time. Many were formed many years before the present and, although they indicate solution of the bedrock, they may not pose a threat of future collapse. These depressions might be termed ‘topographic sinkholes’ or possibly ‘inactive sinkholes’. Sinkholes that have formed in very recent time, particularly as sudden collapses may be termed ‘active sinkholes’ and can indicate instability in the soil or sediment cover of the areas in which they are found. Most sinkholes in state sinkhole databases tend to be based on topographic maps and are of the ‘inactive’ or ‘topographic type’.

Karst hazard
In the context of this report, karst hazard refers to both subsidence and collapse caused by natural karst processes or by other processes working on natural karst features.

Annual cost of karst subsidence in the United States
FEMA (1997) conservatively estimated losses to all types of ground subsidence, including karst, to be at least $125 million per year in the U.S., a very low figure indeed.

Florida is generally accepted as the state most adversely affected by karst collapse. Other sinkhole-prone states include: Texas, Alabama, Missouri, Kentucky, Tennessee, and Pennsylvania. This list is attributed to the U.S. Geological Survey (USGS) where it appears on a few informational webpages (for example, see the USGS Water Sciences School webpage at: http://water.usgs.gov/edu/sinkholes.html). This list appears to be ad hoc, anecdotally-based, and there is no substantive reference for it. There is inadequate sinkhole data available to authoritatively evaluate the relative ranking of these states, or even whether they actually rank within the top seven of all states for occurrence of damaging sinkholes. Many states other than those on the list are also adversely affected by karst hazards to various degrees.

The U.S. Disaster Mitigation Act of 2000 mandates that states must have in place a FEMA-approved standard State Hazard Mitigation Plan to remain eligible for pre- and post-disaster federal hazard mitigation funding. The plan lists significant potential natural hazards that may be expected to impact that state. A review of each state’s multi-hazard mitigation plan, most of them 2013 documents, reveals that 29 of the 50 states discuss karst subsidence as a potential hazard and the remainder apparently do not consider karst hazard as significant enough to discuss in their plan (Figure 2). Interestingly, several states known to have abundant karst features, such as Indiana, Illinois, and Arkansas, do not discuss the karst hazard in their hazard mitigation plans.
**Methodology**

Because of the lack of comprehensive statistics on karst subsidence/sinkhole collapse damage costs, it is impossible to generate a reasonably accurate estimate of the annual cost due to this nationwide. An examination of readily available cost data from the most sinkhole-affected states can provide a lower boundary. The actual costs are very probably much higher than the sum of those reported. Any of the cost estimates contained in this report should be considered uncertain as the data are incomplete and many of the data sources used were secondary in nature.

If there was an authoritative estimate of sinkhole damage cost within a state, that figure was used with citation. This was true only for Florida, Kentucky, and Virginia. For each State where estimates of sinkhole damage costs were not available such costs were generated from an analysis of publicly available data. Such sources include:

1. Sinkhole incident and cost information described in

![Figure 2. Map showing states that discuss karst as a hazard in their State hazard mitigation plan. This figure is based chiefly on 2013 updates to most plans. These plans can usually be found at each state’s emergency management agency website.](image-url)
in each States’ hazard mitigation plan; (2) An internet search for reports of sinkhole damage incidents and costs by city or county; (3) An internet search for sinkhole repair reports from each State department of transportation; and, (4) An internet search for sinkhole incident reports and costs statewide for each year for the period 2000–2014. Cities and counties listed as impacted by sinkhole hazard in the state hazard mitigation plans were used as keywords for specific searches. These internet searches were cursory, employing a popular search engine and the keywords: [city or county name], sinkhole, cost, damage, and state. I generally examined only the first several pages of search returns. Extended links from these search results were often followed in an attempt to find additional information on incidents reported or alluded to. I examined each incident report and used my judgment, based on the geologic setting and the source’s description of the ‘sinkhole’, to decide whether to include it in the state sinkhole cost estimate. Many incidents were obviously not karst sinkholes and were not included. Individual incidents identified in the searches are listed by state in Table 1.

No attempt was made to adjust reported cost amount for inflation, the raw numbers were used. All costs identified for the years 2000–2014 were summed and the average cost per state per year calculated. These averages were then summed to produce a rough minimal estimate of cost to the United States per year.

Other karst-related damage or remediation costs, like ongoing rehabilitation of leaking dams built over karst such as the Wolf Creek Dam, Kentucky, estimated at $300 million (Carey and others, 2008), the Rough Creek Dam, Kentucky at $149,800,000. (U.S. Army Corps of Engineers, 2014), the Center Hill Dam, Tennessee, at $364 million (U.S. Army Corps of Engineers, 2015), and Clearwater Lake, Missouri, for $240,663,000 (U.S. Army corps of Engineers, 2012) were not included in the calculation. The estimate is principally limited to

<table>
<thead>
<tr>
<th>STATE</th>
<th>YEAR</th>
<th>COST</th>
<th>DESCRIPTION</th>
<th>REFERENCE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alabama</td>
<td>2007</td>
<td>$1,100,000</td>
<td>Hanson and Oldcastle (Quarry owners) settle lawsuit with City of Opelika</td>
<td><a href="http://www.aggregateresearch.com/articles/13175/Hanson-and-Oldcastle-settle-11-million-lawsuit-with-City-of-Opelika.aspx">http://www.aggregateresearch.com/articles/13175/Hanson-and-Oldcastle-settle-11-million-lawsuit-with-City-of-Opelika.aspx</a>.</td>
</tr>
<tr>
<td>Alabama</td>
<td>2007</td>
<td>$100,000</td>
<td>City of Madison House over sinkhole. Cost of geotechnical evaluation only</td>
<td><a href="http://blog.al.com/huntsville/2010/07/the_comback_of_the_former_sin.html">http://blog.al.com/huntsville/2010/07/the_comback_of_the_former_sin.html</a></td>
</tr>
<tr>
<td>Alabama</td>
<td>2008</td>
<td>$350,000</td>
<td>City of Tarrant sinkhole repairs</td>
<td><a href="http://www.al.com/birminghamnews/stories/indexssf/base/community/1224663393133860.xml&amp;coll=2">http://www.al.com/birminghamnews/stories/indexssf/base/community/1224663393133860.xml&amp;coll=2</a></td>
</tr>
<tr>
<td>Alabama</td>
<td>2010</td>
<td>$3,000,000</td>
<td>Morgan County, I-65 sinkhole repair</td>
<td><a href="http://www.fhwa.dot.gov/pressroom/fhw1221.cfm">http://www.fhwa.dot.gov/pressroom/fhw1221.cfm</a></td>
</tr>
<tr>
<td>Alabama</td>
<td>2013</td>
<td>$200,000</td>
<td>City of Birmingham, baseball stadium sinkhole repair</td>
<td><a href="http://blog.al.com/spotnews/2013/01/massive_sinkhole_at_birmingham.html">http://blog.al.com/spotnews/2013/01/massive_sinkhole_at_birmingham.html</a></td>
</tr>
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</table>

Table 1. Sinkhole incidents and associated costs from the years 2000–2014. Websites referenced were accessed 6/03/2015.
<table>
<thead>
<tr>
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<th>COST</th>
<th>DESCRIPTION</th>
<th>REFERENCE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alabama</td>
<td>2015</td>
<td>$9,400,000</td>
<td>Repair of sinkholes in Alabama Route 21</td>
<td><a href="http://www.annistonstar.com/the_daily_home/dh_news/article_fba2867e-b195-5726-b2bc-b0c68546542e.html">http://www.annistonstar.com/the_daily_home/dh_news/article_fba2867e-b195-5726-b2bc-b0c68546542e.html</a></td>
</tr>
</tbody>
</table>
| Indiana | 2014 | $11,000,000 | Repair of sinkholes under Monroe County Airport                                 | http://www.airportimprovement.com/article/monroe-county-airport-repairs-airfield-sinkholes |}

Maryland  2003  $2,000,000  Frederick County, new Design Road sinkhole repair  http://www.brunswickmd.gov/wp-content/uploads/2012/01/Frederick-County-Hazard-Mitigation-Plan-Final1.pdf


Missouri  2004  $650,000  Lake Chesterfield drained, sinkhole repaired  http://www.semissourian.com/story/2022983.html

Missouri  2006  $50,000  City of Nixa, Scrivener sinkhole. Cost for city to close hole; not including house and car losses  http://articles.kspr.com/2013-07-18/sinkhole_40662901

Missouri  2010  $30,000  City of Nixa, new sinkhole repair, developer’s estimate to fill  http://articles.ky3.com/2010-12-02/sinkhole_25004905

Missouri  2014  $1,200,000  City of Cape Girardeau, sinkhole road repairs  http://www.semissourian.com/story/2140640.html


Pennsylvania  2004  $6,000,000  Stockertown, rebuilding of Route 33 bridges over Bushkill Creek  http://www.lvpc.org/pdf/hazardMitigation/hazardMitigation.pdf

Pennsylvania  2004  $300,000  City of Easton, St. John Street, street repairs. Several damaged buildings not included  http://www.lvpc.org/pdf/hazardMitigation/hazardMitigation.pdf


Table 1 Continued. Sinkhole incidents and associated costs from the years 2000–2014. Websites referenced were accessed 6/03/2015.
structural damages and remediation and not to personal property loss, such as cars and other valuable items.

Because of the limited scope of this investigation, only states with large areas of karst and located in the eastern part of the U.S. were examined for damage cost reports. Karst subsidence and sinkhole collapses in the western part of the country certainly occur, but are less common. Some of the western states are increasingly encountering karst, such as Colorado, because of human development expansion into areas underlain by evaporite rocks. Some western regions are also affected by gradual subsidence as a result of natural dissolution of deeply-buried salt deposits.

**Alabama**

Large areas of Alabama are underlain by carbonate rocks and are susceptible to sinkhole collapse. One of the largest recent sinkhole collapses in the United States occurred in Shelby County in 1972 with the sudden formation of a 325 feet long by 300 feet wide and 120 feet deep “Golly Hole”. Fortunately this collapse occurred in a rural area and no injuries or significant property damage occurred. The Alabama Emergency Management Agency (2013) reported recent sinkhole collapse of buildings and infrastructure in, and near, the cities of: Sylacauga, Opelika, Valley Head, Huntsville, Auburn, Phenix City, Montevallo, Alabaster, Gadsden, Birmingham, Tuskegee, and Trussville.

For the years 2000–2014 there are reports of at least $15,566,250 in sinkhole damage repairs, a 15 year average of $1,037,750, per year (Table 1).

**Arkansas**

While sinkholes certainly occur, particularly in the Ozark Plateaus region in the north part of the state, no records were found documenting sinkhole collapse damage costs in Arkansas.

<table>
<thead>
<tr>
<th>STATE</th>
<th>YEAR</th>
<th>COST</th>
<th>DESCRIPTION</th>
<th>REFERENCE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pennsylvania</td>
<td>2014</td>
<td>$844,422</td>
<td>Pennsylvania Department of Transportation project to make sinkhole repairs on U.S. Route 422 near Palmyra</td>
<td><a href="http://www.dot.state.pa.us/Penndot/Districts/D8news.nsf/a2a8ee9f2c47a-24b8525783a004f735a/b7eef5ea2f6e84f85257d240052d2f8?OpenDocument">http://www.dot.state.pa.us/Penndot/Districts/D8news.nsf/a2a8ee9f2c47a-24b8525783a004f735a/b7eef5ea2f6e84f85257d240052d2f8?OpenDocument</a></td>
</tr>
<tr>
<td>South Carolina</td>
<td>2014</td>
<td>$4,167,280</td>
<td>City of Georgetown, repairs to municipal buildings damaged by sinkholes</td>
<td><a href="http://gtweb.epp.dc.publicus.com/article/20141231/GTT06/141239982/1129">http://gtweb.epp.dc.publicus.com/article/20141231/GTT06/141239982/1129</a></td>
</tr>
<tr>
<td>Tennessee</td>
<td>2012</td>
<td>$39,450</td>
<td>Unicoi County, Love Chapel Elementary School sinkhole, temporary stabilization and relocations costs only. School was eventually abandoned.</td>
<td><a href="http://www.johnsoncitypress.com/article/102376">http://www.johnsoncitypress.com/article/102376</a></td>
</tr>
<tr>
<td>Tennessee</td>
<td>2013</td>
<td>$100,000</td>
<td>City of Knoxville, private home sinkhole damages</td>
<td><a href="http://www.wbir.com/news/article/264022/2/Sinkhole-nearly-bankrupts-West-Knoxville-homeowner">http://www.wbir.com/news/article/264022/2/Sinkhole-nearly-bankrupts-West-Knoxville-homeowner</a></td>
</tr>
</tbody>
</table>

Table 1 Continued. Sinkhole incidents and associated costs from the years 2000–2014. Websites referenced were accessed 6/03/2015.
Florida
Insurance sources in Florida reported over $84 million in sinkhole losses plus adjustment expenses in 2009. The Florida Office of Insurance Regulation (2010) reported that insurers had received 24,671 claims for sinkhole damage in Florida between 2006 and 2010 totaling $1.4 billion, an average of $280 million per year for those five years. These figures do not include annual costs incurred by the Florida Department of Transportation in highway repairs due to karst subsidence. In Florida the loss partly comes from gradual subsidence damage to homes and other structures, as well as incidents involving outright collapses. Some portion of the insured damages, such as foundation cracking, are probably caused by non-karst processes such as differential settling, expansive soils, or poor building practices.

Georgia
There were several news media reports mentioning sinkholes, but all were related to failed storm water drainage structures. The Georgia Emergency Management Agency (2014) discussed the geology and geography of potential sinkhole areas in the state, but has no data on specific incidents or costs.

Indiana
Indiana is underlain by large areas of karst and has many known sinkholes. Sinkhole damage and threat of further collapse to the Monroe county airport, first noticed in the 1990’s was repaired for $11 million in 2014 (Scott, 2014; Table 1).

Illinois
No reports of recent sinkhole damages were found for Illinois. Most sinkholes in Illinois are in rural areas in the southern part of the state.

Kentucky
Kentucky is one of the few states that have attempted to keep a sinkhole collapse database and to quantify the cost of sinkhole collapses. Cover-collapse sinkholes in Kentucky cause about $20 million in damage per year.
(Currens, 2012). This figure includes Kentucky Department of Transportation highway repairs of sinkhole damage.

**Maryland**
Frederick County has been a hotspot for sinkhole collapse incidents in Maryland in recent years, as there is a combination of geology conducive to sinkhole formation and urban development of the greater City of Frederick area (Brezinski, 2007). The Frederick County hazard mitigation plan (2010) lists several examples, but only one entry includes a dollar amount: $2,000,000. A sinkhole collapse of a road in Washington County in 2007 cost $217,141 to repair (Table 1).

**Minnesota**
No cost amounts for specific incidents were found, although the Minnesota state hazard mitigation plan (2014) lists several incidents of sewage lagoon collapses into sinkholes.

**Missouri**
Missouri is extensively underlain by Paleozoic carbonate rocks in the Ozark Plateau region. Because of extensive areas of thick soil or residuum over soluble rocks, some parts of Missouri are particularly vulnerable to cover-collapse sinkholes. The Missouri State Hazard Mitigation Plan (2013) designates counties with the most topographic sinkholes, but does not list specific collapse or subsidence incidents. Those counties are: Cape Girardeau, Christian, Dent, Greene, Howell, Oregon, Perry, Shannon, St. Genevieve, St. Louis, and Texas. A USGS Fact-sheet on sinkholes in Missouri (Kaufman, 2007) listed a few notable collapses but does not include any cost statistics. An examination of the Missouri Department of Insurance website did not reveal any sinkhole-related statistics.

Easily identifiable incidents with costs for the period 2000–2014 added up to only $1,930,000 or an annual average of $128,666 (Table 1). Based on an extensive bedrock geology that is prone to karst sinkhole development, this amount surely underestimates the actual cost of sinkhole damages in Missouri.

**New Jersey**
The 2014 State Hazard Mitigation Plan lists several sinkhole collapse incidents since 2000 but no cost data are available. Most natural sinkholes occur in the northwestern part of the state, particularly in Warren County (New Jersey, 2014).

**North Carolina**
The 2013 North Carolina State hazard mitigation plan discusses the geology of natural sinkholes which are concentrated in the southeastern coastal plain part of the state. No sinkhole damage costs were listed (North Carolina, 2013).

**Pennsylvania**
Abundant sinkhole locations in Pennsylvania include the Saucon Valley of Lehigh County, the greater Harrisburg metropolitan area in Dauphin and Cumberland Counties, and the Nittany Valley in Blair, Centre, and Clinton Counties (Pennsylvania Emergency Management Agency, 2013).

Easily identifiable incidents with costs for the period 2000–2014 totaled $28,404,422 and averaged $1,893,628 per year (Appendix I).

**South Carolina**
The South Carolina state hazard mitigation plan (2013) discusses sinkholes, natural and man-made, but contains no loss data. Sinkholes that damaged municipal buildings and private property in the city of Georgetown in 2011 caused at least $4,167,280 in damages (Table 1).

**Tennessee**
Tennessee has numerous sinkhole incidents and, in addition to Florida, is the only state to require all insurance providers to offer sinkhole insurance. The 2013 Tennessee State Hazard Mitigation Plan lists only one historical sinkhole incident and identifies broad areas of the state, based on the mapping by Weary, (2008) as having potential for subsidence or sinkholes (Tennessee, 2013). There is also a map of land subsidence hazard relative risk index by county that was generated in a GIS. The index ranges from 1 to 6, with 6 being the highest. The index was apparently based on area of each county underlain by potentially karstic rocks.

Easily identifiable incidents with costs for the period 2000–2014 totaled only $696,450 or an average of $46,430 per year (Table 1). Very few incident reports were available in this karst-rich state, so the amounts obviously underrepresent the actual costs.

**Texas**
Karst terrain in Texas straddles the divide between karst in humid climate regimes and karst in arid and semi-arid climate regimes of the United States (Weary and Doctor, 2014). It has extensive areas of karst and numerous natural sinkholes, but apparently has few incidents of cover-collapse failures that damage property. The Texas state hazard mitigation plan (2013) includes sinkholes as a subcategory of land subsidence. It lists only 2 specific sinkhole incidents: one in Wink (the Wink Sinks), Winkler County in 1980, and the other at Daisetta, Liberty
County in 2008. Both of these incidents are associated with dissolution of evaporite rocks at depth and were probably influenced by drilling and fluid injection.

**Virginia**
The Virginia hazard mitigation plan lists several historical sinkhole collapse incidents but provides little cost information (Virginia, 2013). The Virginia Department of Transportation (VDOT) has some record of cost data associated with sinkhole repairs to highways and roads and it has been roughly estimated that about $8 million dollars were spent on sinkhole repairs over the years 2000–2014 (B. Bruckno, VDOT, 2015, personal communication).

**West Virginia**
The West Virginia statewide standard hazard mitigation plan categorizes karst hazard as a type of land subsidence, but lists no specific incidents or costs (West Virginia, 2013). It does suggest that the West Virginia Department of Transportation collect data for future reports on highway damage costs. Sinkhole hazard in West Virginia occurs mostly in the eastern parts of the state, particularly in the Greenbrier Valley and in the Eastern Panhandle.

**Results**
Over the last 15 years (2000–2014) sinkhole collapse and karst subsidence has cost on average, at least $304,316,761 per year, based on the sum of costs listed in Table 1, as well as the costs reported for Florida, Kentucky, and Virginia.

**Comparison of karst subsidence costs with other natural hazards/disasters**
Natural hazards are natural processes which cause loss of life, injury, or health impacts to people; property damage; social or economic disruption; or environmental damage. A natural disaster is a disruption of the functioning of a community or a society involving widespread human, material, economic, or environmental losses and impacts that exceed the recovery ability of the affected community or society (Holmes and others, 2012). A disaster usually occurs as a single event or series of events.

Since data are quite sparse, comparing the annual costs of karst subsidence damage to the better-known natural hazards is difficult. Most of the direct cost of a subsidence event may be the repair or replacement of damaged structures, repairs to infrastructure, particularly roads and bridges, and loss of real estate values.

Natural hazards and disasters in the United States result in direct costs averaging many billions of dollars annually. These include: floods (about $10 billion per year) (The Association of State Floodplain Managers, 2013), hurricanes and cyclonic storms (about $10 billion per year; Pielke and others, 2008), tornadoes (about $5 billion per year; Simmons and others, 2013), landslides ($3.5 billion per year; Kjekstad and Highland, 2008; Shuster and Highland, 2001), earthquakes (about $2.5 billion per year; Vranes and Pielke, 2009), wildfires (about $1.5 billion per year in federal suppression costs alone; National Interagency Fire Center, 2014), volcanic eruptions (rare but costly), and tsunamis (rare, but potentially very costly). Based on the information collected in this report, sinkhole damages in the United States average at least $300 million per year. Figure 3 illustrates the relative annual cost of natural hazards and disasters, including sinkholes, in the United States.

**Discussion**
The estimated annual cost of sinkhole damage, $304 million per year, is a very conservative minimum figure that is based on the relatively few incidents that are documented in this report. Complete information about the cost of damage by karst subsidence incidences is not available. These include: unreported property damage to private, commercial, and government buildings; and most of the cost of highway repairs (except for the losses reported in Kentucky and a few other incidents in other States).

If a full accounting were possible for the total actual costs of sinkhole damage, it is likely that annual damages would be significantly higher than the amount reported here.

In compiling information for this report, no karst cost information was found from several states that have karst and therefore must have at least some occurrence of sinkhole damage. Addition of these losses will increase the national cost estimate.

State geological surveys generally are good possible sources for obtaining karst subsidence and sinkhole collapse cost information aware of the extent of karst lands and the potential of karst hazard within their respective states. Some of these surveys keep records of sinkhole incidents, but most do not have the funding or political mandate to obtain and keep records programmatically. Perhaps a survey initiated by an organization such as the American Association of State Geologists (AASG) could poll the state geological surveys for known incidents and cost estimates.
As mentioned earlier in this report, each state has a hazard mitigation plan, which includes a format for karst, sinkhole, or ground collapse hazard description and historical incidence. These descriptions are usually generalized and detailed data are generally not easily available. Many of the states rely on their geological surveys to complete this portion of their mitigation plan and so these surveys may it be better primary sources. At the federal level perhaps FEMA could encourage greater reporting detail in those state plans.

In some states insurance organizations and regulators collect karst subsidence and sinkhole collapse damage information such as Florida did in a data call to their property insurers (Florida Office of Insurance Regulation, 2010). This data call represented an investment in time and money and was driven by the mounting losses to sinkhole insurance claims in Florida, a situation that has not occurred in other states. Without such a driver, it is difficult to mandate similar calls elsewhere, but voluntary surveys may be a possibility.

There is some information available through State Departments of Transportation, generally from archived news releases and public contract awards. An organization such as the American Association of State Highway and Transportation Officials (AASHTO) could sponsor a specific survey of State highway maintenance engineers. Once again, there is the problem of highway department’s generally not recording sinkholes as a unique type of damage repair and because of the need for a geologist or engineer to separate karst incidents from other types of collapse.

Of course there is also potential karst information and informational contacts available through the many karst scientists in in the United States. The Geological Society of America (GSA) recently added a Karst Division (2014) which could access a large contact list for information and suggestions on defining the karst hazard. Likewise, the National Cave and Karst Institute (NCKRI) could also reach out to its contacts and friends for information.

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HAZARD OF SINKHOLE FLOODING TO A CAVE HOMININ SITE AND ITS CONTROL COUNTERMEASURES IN A TOWER KARST AREA, SOUTH CHINA

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Abstract
Zengpiyan Cave, one of the most important cave hominin sites of the Neolithic in the South of China, was listed on the national register of cultural preservation sites in 2001. Large quantities of precious material in the Zengpiyan site were unearthed since the beginning of the trial excavation in 1973. These materials include hominin skeletal remains, fire pits, human burials, stone implements, tools fashioned from mollusk shells and animal or plant fossils.

According to the historical record, ancient people lived in caves in the karst plain of Guilin. They moved out of the caves approximately 7,000 years ago. These cave hominin sites provide important material for understanding ancient environmental change and human prehistory. However, the exploratory shaft of Zengpiyan had no appropriate treatment after the initial excavation. Groundwater flooded the exploratory shaft due to frequent rises in the water table, resulting in collapse of the exploratory shaft shoring as well as some other serious damages. Even though some rescue and protection measures were taken, for example, slope supporting and backfill treatment, they failed to eliminate the hidden trouble caused by rapid fluctuation of groundwater levels during the rainy season. Rapid urbanization is also affecting this region. Infrastructure construction of the city changes the hydrogeological conditions of the karst, increasing the area of impervious surface and risks of urban flooding. Moreover, an increase of extreme climate events may lead to frequent flooding of the site by groundwater. Therefore, a focused hydrogeological investigation was carried out to study the status of the site and karst development, and
the mechanism of groundwater movement at local and regional scales. These surveys include borehole drilling, electrical resistivity surveys, computed tomography (CT) scanning technology, dye tracer tests, groundwater monitoring, and hydrochemistry analysis. The results show that the site is located in the seasonal fluctuation zone of the groundwater. Water level in the karst aquifer is sensitive to rainfall. Continuous rainstorms lead to synchronous rising of the groundwater level in both the cave and the aquifer. In addition, surface runoff and urban sewers cannot discharge smoothly, resulting in surface water backflooding into the cave entrance. Therefore, controlling the recharge of groundwater and the influx of surface runoff, and dredging a groundwater discharge channel are all important in order to reduce the damage of flooding to the archaeological sites. Based on these detailed investigations and research results, countermeasures for flood control and archaeological site protection were put forward.

We recommend that the engineering measures should combine curtain grouting, drain construction, and effective water resources management for the entire basin. Even though the measures are feasible, we can’t promise a perfect damage control of the ruins by water due to the complex hydrogeological conditions in the covered karst area.

**Introduction**

Guilin in Southern China not only has a unique karst Fengcong and Fenglin landscape in the world, but also a long cultural history. Evidences of human activities date back 10,000 years B.P. According to the investigation from the Chinese Cultural Relics Department, 71 caves in Guilin have evidence of ancient human activities. These sites provide abundant artifacts and human remains for archaeological excavation (Wei, 2011). However, karst development and recent human activities raise challenges for environmental protection of these cave archaeological sites. Karst develops well in Guilin due to geology and climate conditions. Relative groundwater dynamics and generated corrosion and dissolution processes are active, resulting in frequent karst collapse occurrences, and threatens the preservation of cave relics.

The Guilin basin is a covered karst region. The interface between the surface soil and the sub-strata is located within the seasonal fluctuation (vadose) zone of the groundwater aquifer. Fluctuations in the groundwater elevation and its erosive capacity can not only cause numerous caverns within the bedrock, but can also carry off soil particles by erosion, the main driving force of karst collapse. Some cave relic sites in Guilin are also located within the groundwater seasonal fluctuation zone. It is inevitable that they will be disturbed by this groundwater fluctuation, with high risk of geological hazards from erosion and collapse.

**Study Area**

The Zengpiyan Cave was one of the residences and cemeteries of ancient residents in the Guilin area from 12,000 to 7,000 years ago. The cave is located in an isolated carbonate hill in a tower karst plain in Southwest Guilin. The surface of this plain has an elevation of 154 m above sea level. This cave, developed in the sparite and calcirudite from the Rongxian Group of the Devonian system (D₃r), is composed of a dry cave and a water cave (Figure 1). The dry cave is where the ancient humans lived and has a 10 m wide and 4 m high entrance (if not stated, ‘cave’ mentioned in this paper always indicates the dry cave). The entrance of the cave faces south, and the rest of the cave extends to northeast for 20 m. The altitude of the cave bottom coincides with the outside ground. A three meters thick cultural layer is deposited in the dry cave, carrying rich information from 12,000 B.P to Song Dynasty (960-1279 A.D.). In 1973, fossil remains of pre-historic human, stone implements, bone tools and pottery pieces dating some ten thousand years B.P. were unearthed within the cave. The cave has since become recognized as one of the most important repositories of human relics from the Neolithic Age in South China. The elevation of the cave floor ranges from 154 to 154.8 m. Archaeological excavations produced seven main vertical test pits, the deepest to a depth of 3 m. Evidence of intense dissolution of the limestone bedrock could be found in the bottom of the pit.

Groundwater levels in this area fluctuate in the range of 149.59 to 154.93 m elevation. Therefore, the cave is located in the groundwater seasonal fluctuation zone. Immersion and flow of the groundwater causes a series of problems including cave sediment softening, erosion and removal, collapses of the test pits, bottom collapse, and cave pollution. Controlling the groundwater level and its rate of fluctuation as well as water pollution are important for the protection of the site.
The hydrogeological investigation in this karst area aims to study the groundwater storage pattern and movement. It also characterizes the spatial variance of karst development, and connections between karst features and groundwater behavior, in order to provide the basis for engineering design and implementation to modify groundwater movement. Although direct groundwater encroachment is observed only at the cave archaeological site, it is influenced by groundwater movement over a much broader area. Therefore, determination of the range of groundwater flow paths is a crucial task. This study selects two scales of exploration scope, including regional scale and local scale. The regional scale study aims at understanding the formation condition and groundwater flow paths by geological analysis and hydrogeological investigation. The catchment area of hydrogeological unit and its boundaries are then determined. The local scale study aims at studying the specific locations of karst conduit development and the source and direction of groundwater which has direct connection with Zengpiyan Cave. The local scale groundwater investigation was performed using diverse methods, including geological analysis, hydrogeological investigation, and hydrochemical analysis. The summary of the regional scale groundwater investigation will be presented here. Figure 1 shows the schematics of the Zengpiyan Cave system. Modified from Qin et al. (2011).
Results and Discussion

Hydrogeological Investigation

The hydrogeological unit where the Zengpiyan cave is located extends from clastic rock mountains in the northwest, flowing through the plain in the basin, and then ends up in the Lijiang River to the southeast of the cave. However, the lateral boundaries of the flow unit are not clear. There are two surface rivers outside Zengpiyan Cave in the north (Nanxi River) and in the south (Nanwan River), respectively (Figure 2), but they can’t be the groundwater discharge boundary due to their insufficient incision into the aquifer. Within this unit, the mountainous area is much smaller than that of the plain. Most flooding generated from the mountains is discharged by streams and ditches, and at baseflow period, most of the mountain runoff is recharged to groundwater in the plain. Rainfall in the plain recharges the groundwater. With urbanization, surface runoff generated by imperious surface flows to stormwater drains. Waste water from the urban area is delivered to sewage treatment plants. Pools in the plain can retain some rainwater, but their total area is relative small in size. Most of the discharge in the Nanxi River is coming from domestic wastewater. The Nanwan River originates from a spring, but it is silted up seriously during its course, resulting in limited water discharge capacity. Therefore, most runoff is discharged to the Lijiang River from the aquifer. The riverbed of the Lijiang River has a thickness of 50 m and is composed of sand gravel. Springs exist in the riverbed but are buried by sand and gravel. The spring’s outlets in the riverbed were discovered through interviews with local peoples. In the karst plain, natural groundwater outlets are located in the caves within the peaks, such as Zengpiyan. The depth of hand dug wells in the rural areas is usually less than 10 m, but they can still satisfy household water demand.
Zengpiyan Cave is located in a karst tower, close to the upstream part of the hydrogeological unit. No surface stream exits in the neighborhood, and caves are the important water storage systems. Groundwater level fluctuates frequently, reflecting good recharge and discharge conditions. Because the groundwater depth ranges from 2-4 m (basically lies in the vicinity of the interface between the covering layer and the karst aquifer), pools in the covering layer have close relationship with groundwater. The pool in the Zengpiyan Cave and the surrounding aquifer have good relationship with other pools in the hydrogeological unit. A dye tracer test shows that groundwater flows from the north and west of Zengpiyan to the water cave (Figure 1). The water then flows to pools to the south. Borehole drilling and CT scanning indicate that karst features are more developed near the cave entrances than in other parts of the karst terrain. Hydrochemical analysis suggests that an abnormal zone which has significantly different water chemistry is found in the cave entrance and in the more intensive developed karst zones.

Hydrogeological investigations indicate that Zengpiyan Cave and its surrounding area belong to a concentrated flow zone in the aquifer of the Guilin Basin, and their groundwater storage is considered to be the largest in the basin.

**Groundwater Concentrated Flow Zone**

Delineation of a groundwater concentrated flow zone was the most significant finding of this hydrogeological investigation. Determination of this zone was achieved by using data from hydrogeological analysis and pumping tests. A pumping test was done in a blue hole to the southeast of Zengpiyan Cave in 1960. At the same time as the test, groundwater levels in the Zengpiyan boreholes and blue holes to the northwest of the cave were observed. According to the contours of the water table drawn down during the pumping test, an elliptical cone of depression trending in a SW direction along its long axis was produced. The radius of influence in the SW direction is much bigger than the other directions, indicating that water transmissibility from the SW direction is greater, forming a groundwater concentrated flow zone in the aquifer (Figure 2). The Zengpiyan site is located in this groundwater concentrated flow zone.

Groundwater is solutionally aggressive in the concentrated flow zone, and karstification, including cave development is intense in the aquifer there. For example, a cave is developed in the peak, which is 400 m away from Zengpiyan to the north. In the rainy season, the lowest parts of the caves regularly collect water, suggesting these caves are most likely located in the groundwater level fluctuation zone. A 6 m high cave was found in the borehole in the north of Zengpiyan, and according to CT electromagnetic wave, the cave is 28 m wide. For 21 boreholes drilled around Zengpiyan Cave, 10 boreholes contained caves, with the highest being 10 m high.

Groundwater discharged from springs also reaches the concentrated flow zone. In some places, groundwater rises to the top surface after rainstorm, indicating that the groundwater recharge comes from higher elevations, and then flows into the plain after moving long distance. Strong groundwater activities often result in frequent karst collapse.

Within the groundwater concentrated flow zone, boreholes have large pump test flows, suggesting that the aquifer has high water storage there. Groundwater exploitation in the Guilin Basin usually uses well drilling and water pumping from caves. For example, cavern water like that at the Zengpiyan Cave is common in the Guilin Basin, and commonly used for drinking.

Water discharge can be divided into three levels of productivity according to the specific capacity for boreholes by pumping tests. When specific capacity is lower than 1 L/(s•m) (the unit means the water inflow when water table decreases by one meter), it means the water yield of borehole is small. This often happens when borehole is located in a part of the aquifer where karst is undeveloped, indicating that karst media is primary in fissure and fracture. When the borehole flow ranges from 1-10 L/(s•m), the flow belongs to medium level, indicating that there are some dissolution fractures existing in the aquifer where borehole is located. Water yield is abundant in the aquifer when the specific capacity is bigger than 10 L/(s•m), suggesting that karst is well developed in the aquifer.

Flow of water pumping from caves is greater than from boreholes. For example, pumping test in the boreholes around Zengpiyan Cave showed that the scope of groundwater depth declining is less than 1 m when the pumping load is 2.8 L/s. When the pumping load reaches 10 L/s, groundwater declining is still less than one meter within an hour. The coefficient of aquifer transmissibility...
in the concentrated flow zone was 9-26 m/s according to pumping test, and the groundwater flow velocity was 10-40 m/d as calculated by tracer tests.

**Treatment Scheme of Water Damage**

The results of a hydrogeological survey provide the basis for mitigation of water damage in Zengpiyan Cave. The treatment concept is to reduce the risk of groundwater flooding, erosion, and pollution of the site through engineering measures. The objectives are to control the groundwater level to a threshold value and to control its fluctuation velocity within a permissible range.

Prior to engineering measures to control groundwater movement, karst water types should be determined first. Karst water types are classified as subterranean stream (water-bearing conduits), concentrated flow zone, and general karst fissure water (Table 1). The subterranean stream has the strongest hydrological dynamics, and thus it has the lowest engineering success rate. If the subterranean stream is blocked, there is a risk of water leakage and collapse. Besides, the upstream of underground stream often becomes flooded after the stream is blocked. For groundwater concentrated flow zone, karst collapse is highly probable. Another characteristic is that groundwater is usually abundant in this zone, with good interconnection of karst media, resulting in high water transmission. Curtain grouting in part can reduce water inflow, but the accompanying environmental change should be considered and evaluated. By contrast, engineering treatment in karst fissure areas, in general, is relatively simple.

Eight treatment measures for relieving groundwater hazard are proposed, including monitoring, evaporation, dewatering, pumping, impermeable layer setting, curtain grouting, and environmental treatment. These eight measures should be conducted coordinately, and prioritized according to minimize damage to the environment. A small scale experiment should be performed first, to confirm its effectiveness before further investment scale starts.

**Environmental Monitoring**

The aims of monitoring are to learn groundwater hydrological dynamics and hydrochemistry at all times, to observe possible immersion, scour, and pollution events; to understand the impact of surrounding human activities on the cave relics and environment; and to prevent new problems from occurring. Monitoring includes: atmospheric environment change (namely, air temperature, precipitation, humidity, wind speed, pressure, and CO$_2$ content); and, groundwater environment (namely, water level, flow, fluctuation velocity, water temperature, and water quality) in both cave drip water and in groundwater.

**Water Draining by Evaporation**

Enhance the water draining rate of groundwater by means of increasing evaporation. Because groundwater depth in and surrounding Zengpiyan Cave ranges from 2-4 m, tree roots can use groundwater directly. Using trees to enhance groundwater discharge is a measure of not only environmentally friendly, but also economical. Specifically, increase the green area in and surrounding Zengpiyan Cave, using tree species which have high water consumption during plant growth period.

**Water Draining by Open Channel**

The aim of water draining by open channel is to discharge surface runoff and part of groundwater during raining periods, in order to reduce the groundwater level, and to solve the problem of groundwater level exceeding the warning level. According to interviews with local people, surface runoff is discharged inefficiently during rainstorms. So much so that surface runoff can flow back into the caves, which is a key factor of rising groundwater level during rainstorms. The main reason for inefficient surface runoff is that there are still some areas without rainwater conduit in the upstream parts of the Zengpiyan Cave hydrogeologic unit. Surface runoff flows to the low-lying areas, and then recharges groundwater. Under extreme conditions, rainwater conduits nearby can’t

### Table 1. Karst water types and corresponding treatment scheme.

<table>
<thead>
<tr>
<th>Karst water types</th>
<th>Specific capacity/ (L/s·m)</th>
<th>Karst developed</th>
<th>Engineering technical measures</th>
<th>Geological hazard</th>
</tr>
</thead>
<tbody>
<tr>
<td>Subterranean stream</td>
<td>&gt;10</td>
<td>very heterogeneous</td>
<td>blocking, water diversion by tunnel</td>
<td>flooding, dam falling</td>
</tr>
<tr>
<td>Concentrated flow zone</td>
<td>&gt;10</td>
<td>relatively heterogeneous</td>
<td>Curtain grouting, pump and drain of water, or water diversion by open channel</td>
<td>collapse</td>
</tr>
<tr>
<td>General karst fissure</td>
<td>&lt;10</td>
<td>even</td>
<td>pumping</td>
<td>none</td>
</tr>
</tbody>
</table>
Environmental Management

Zengpiyan Cave is located in the urban Guilin City. In addition, the aquifer is a highly vulnerable karst aquifer. Urban development is changing the environment of the Zengpiyan site, including groundwater-level variation and atmospheric environmental changes. Along with an engineering treatment, an integrated and long-term scheme should be set up to protect Zengpiyan Cave. The risk of groundwater pollution is a challenge for protection of the site because the official protection area is very local. However, pollutants originate from faraway areas and from many directions, which are not clear in most conditions. It is very difficult for the Administrative Department of Zengpiyan to deal with these kinds of problems. These kinds of issues need societal awareness, including scientists, local people, and government management departments.

Conclusions

Groundwater damage reduction for the Zengpiyan Cave site requires reducing the fluctuation speed of groundwater and decreasing the intervals of groundwater flooding in the cave. Because the groundwater influence on the site is of regional-scale, it is difficult to change the natural dynamics of groundwater through local-scale engineering. Moreover, site protection requires changes in the natural environment where the site is located. Eight treatment measures are proposed by this study, in which curtain grouting has the highest impact on the environment. So curtain grouting needs to be carefully planned and conducted if it is selected. Groundwater modeling is an effective tool to learn the control mechanism of groundwater dynamics, and to evaluate the engineering measure’s effectiveness. However, when groundwater modeling is used for groundwater hydrological dynamics in a karst aquifer, high precision hydrogeological information is needed, especially when the study area is small and factors which affect groundwater movement are unclear. Water pumping is an important measure to control groundwater fluctuation, but experimental verification is necessary. On the other hand, pumping test in a large scale may have a high risk of causing karst collapses. With small pumping rate used in this study, groundwater decline is less than 1 m so analogical measures have been used to make up the deficiency of water pumping, which provide information for evaluating the engineering measures.

Water Pumping

Water pumping means when groundwater table exceeds the warning level, groundwater should be pumped directly from the caves, in order to protect cave relics. Or before the rainstorm comes, groundwater in caves can be pumped out to make room for rainwater. It has been proven feasible to lower the groundwater table by water pumping through pumping test performed and reference to nearby experiments. However, in order to raise the efficiency of water pumping, lining the bottom of caves with an impermeable layer is suggested.

Impermeable Layer Setting

Setting an impermeable layer in the bottom of the caves can reduce the connection between cave water and groundwater. The function of this layer can be demonstrated in two aspects. Firstly, an impermeable layer in the bottom of caves can retard groundwater recharge to cave during normal conditions and control sudden groundwater rises or declines in the caves. Secondly, when groundwater is pumped manually, it can quickly enhance the efficiency to reduce water table in the caves. Non-polluted and recycled impermeable material should be chosen. Quartz sand is suggested as the best material.

Curtain Grouting

Curtain grouting is suggested to cut off the groundwater source in Zengpiyan Cave. From the viewpoint of engineering technology, it is feasible to take this measure (Mai and Zeng, 2009; Bai, 1996). We reference a successful case in nearby Wanfu Road which had used curtain grouting in drainage of a foundation pit. groundwater inflow decreased ten times when the grouting depth reached 25 m and used a double row of holes in 2 m intervals (Wang et al., 2006). Therefore, curtain grouting can intercept groundwater flow in concentrated runoff zone effectively (Tan, 2013). But due to high risk, heavy pollution, and irreversible engineering; curtain grouting is not suggested if there are better measures (Liu et al., 2014).
The Zengpiyan site suffers from groundwater fluctuation, which seriously threatens the preservation of the cave archaeological materials. Greater knowledge of groundwater dynamics, including relationships between hydrogeology and human activities are needed to combat this problem. To analyze the mechanism of groundwater hazard to cave relics, to provide feasible measures for relic protection, to change groundwater dynamics by engineering measures, to coordinate the relationship between relic protection and groundwater movement, all of these issues are new innovative research orientation in hydrogeology and engineering geology.

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CASE STUDIES OF ANIMAL FEEDLOTS ON KARST IN OLMSTED COUNTY, MINNESOTA

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Abstract
A unique area of Olmsted County is located a few miles southeast of Rochester by the small community of Predmore (Figure 1). Surface geology within the Orion Sinkhole Plain is dominated by a large array of sinkholes and limited soil cover over carbonate bedrock of the Ordovician Stewartville and Prosser Formations. Dye trace studies completed by Eagle and Alexander (2007) have demonstrated that a large portion of the plain’s groundwater discharges into springs that feed two local trout streams. Land-use in the area is mixed. For generations, local farmers have relied on livestock for stable income and profit. To put the 8,000 acre region into perspective, there are approximately 3,600 animal units located at 12 facilities which produce an estimated 74 million pounds of manure per year (United States Department of Agriculture / Natural Resources Conservation Service, 1995) and 10 million gallons of manure contaminated runoff. (Larsen et al., 2014) 349 known karst features exist of which 316 are sinkholes (Alexander et al., 1988).

Following snowmelt and rain in March of 2013; an incident occurred where an area well was potentially impacted. Investigation revealed manure contaminated runoff was entering groundwater in a newly discovered sinkhole (Larsen, 2013). Local citizen concern grew for groundwater quality.

Developing relationships with landowners and livestock producers became necessary for protection of water resources and has facilitated research, education and action. A newly formed sinkhole that seasonally receives feedlot runoff was studied with ground penetrating radar for repair. Two producers in the region are implementing manure management techniques that are more stringent then regulation. The Wiskow dye trace was completed in spring of 2014. The study identified discharge springs that discharge into the Mill Creek trout stream from two vulnerable sinkholes (Johnson et al., 2014). Four springs and four previously unknown sinkholes were identified and mapped. A manure contaminated runoff storage area was constructed in the fall of 2014 by a livestock producer located at a headwater spring of Mill Creek. A filter strip and large manure contaminated runoff system is being designed for construction in 2015.

Building great relationships with producers has been successful in Olmsted County. Livestock producers are making investments and taking action. Producers are an essential component of the mid-western economy and assistance with information, funding and resources will help protect the environment and keep farms profitable for future generations.

Introduction
A large array of sinkholes known as the Orion sinkhole plain is found south of Interstate 90 and north of the...
According to site evaluations, 39 million liters (10 million gallons) of manure contaminated runoff (Larsen, et al., 2014). 349 known karst features exist of which 316 are sinkholes (Alexander, et al., 1988).

Although karst geology is widespread across Olmsted County, unique challenges exist for the protection of water resources by livestock producers in this area. There are engineering complications for the design of manure storage structures that may hold millions of gallons of manure and manure-laden runoff. Proper investigation and siting must take place so that the risk of soil collapse or seepage from underneath the structure is minimized. Operational challenges exist for livestock producers who are applying manure in the study area not just in proximity to sinkholes. The entire area is underlain by a karst system with widespread dendritic, underground drainage that discharges to springs and streams. Proper manure management techniques must be developed and implemented to minimize the risk of manure contaminated runoff entering the groundwater system.

Surface geology of the Orion Sinkhole Plain is dominated by a large number of sinkholes and limited soil cover over carbonate bedrock of the Ordovician Stewartville and Prosser Formations (Figures 2 and 3). Dye trace studies completed by Eagle and Alexander (2007) and Green (2004) have demonstrated that a significant portion of the groundwater from the sinkhole plain discharges into springs that feed two local trout streams (Figure 5).

Land-use in the area is mixed. For generations, local farmers have relied on livestock for stable income and profit. The highly variable soil types are not well suited for farmers to depend exclusively on a cash grain cropping system. Within the 3,200 hectare (8,000 acre) region, there are approximately 3,600 animal units. Twelve facilities produce an estimated 34 million kilograms (74 million pounds) of manure per year (United States Department of Agriculture / Natural Resources Conservation Service, 1995).

Figure 2. Map of the Orion Sinkhole Plain with Case Studies and Feedlots Identified (Larsen, 2015).
The objectives of this study are to 1) show case studies and the problems encountered in the study area; and 2) to illustrate beneficial relationships between livestock producers, engineers, scientists and regulators that have had positive results and are essential. There is not a complete understanding of how current rules and engineering practices are performing in such dynamic karst environments. In order to enhance these standards and practices, parties must interact in a collaborative manner so that information, knowledge and feedback is shared.

Case Study 1 – Concern for Groundwater and Dye Tracing to Determine Groundwater Flow

Concern for groundwater contamination resulting from livestock production overlying karst systems has occurred many times in Olmsted County. Citizen trepidations commonly arise when new animal operation proposals are being considered. For example, in the late 1970s, two large swine facilities were proposed in the county. A contentious discussion between farmers, regulators, and other property owners focused on karst systems and groundwater quality. At one of the proposed sites, an Eyota area farmer constructed an un-permitted concrete block manure storage pit (which was excavated into the Galena Limestone). Following construction of the storage pit, a nearby private well in the Galena Limestone allegedly experienced “dark-colored” water and the renters’ wife was hospitalized after ingesting that water in the fall of 1978 (Landherr, 1979, written communication).

More recently, an incident occurred on March 11, 2013, where a private well on the northern edge of the Orion sinkhole plain was potentially impacted with manure-contaminated runoff. A series of climatic events created a “perfect storm” for runoff. Rainfall occurred early in the winter of 2012 – 2013 when there was limited snow cover in the region. As the winter progressed, little additional snow fell and

**Figure 3.** Illustration of the stratigraphy in the Orion Sinkhole Plain (Alexander, et al., 1988).

**Figure 4.** A pail of foaming water collected from the contaminated well. Foam can commonly occur in water contaminated with human or animal waste (Schmidt, 2013).
temperatures dropped dramatically. The rain and snowmelt that occurred on March 8 – 10 was not able to infiltrate the ground that was sealed by ice and thick frost. Flowing water carried large amounts of manure contaminated runoff from farmland throughout southeast Minnesota.

During the week prior to March 11, 2013, a farmer applied cattle bedding pack on a field southern Eyota Township in compliance with all applicable federal, state and local laws. The field does not contain any obvious sinkholes. For the most part, soil is thin, the field has less than 5 feet of soil cover overlying the Prosser Limestone that is visible in outcrops within the field. Soil texture is mixed with sands and sandy loam inclusions.

According to a neighboring landowner, their well water became discolored and odorous following the manure applications and subsequent runoff event (Figure 4). The well, which is 43 meters deep, was constructed in the early 1950s has 6 meters of steel casing. It was drilled in the highly fractured and porous Ordovician Prosser Formation. It does not penetrate the lower Decorah Shale, which serves as a regional aquitard (Figure 3).

County and Minnesota Pollution Control Agency response lasted many weeks. Clearly significant amounts of runoff left the application field and moved thousands of meters into drainage ways and neighboring properties. The movement was tracked, documented and mapped with GIS tools and software. Vulnerable wells in the surrounding area were monitored and tested. It was not until a later snowmelt event on April 3, 2013, when manure contaminated runoff was discovered entering groundwater (Larsen, 2013).

Runoff entered the ground water through a subtle, newly discovered sinkhole. The swallow hole was only 5 centimeters in diameter; however it was accepting large amounts of runoff.

The Wiskow dye traces were completed in spring of 2014 as a result of the manure contaminated runoff incidents of March 2013. Many participating agencies were involved. The traces were completed by Jeff Green and Scot Johnson of the Minnesota Department of Natural Resources (DNR) Ecological Water Resources Division, Martin Larsen of the Olmsted County Soil and Water Conservation District and E. Calvin Alexander, Jr. of the University of Minnesota Earth Sciences Department.

Two sinkholes were identified in the Prosser formation to be used in the trace. One, the small newly identified sinkhole on the Wiskow property; (MN55:D00983) located at 557550 E / 4865569 N ± 3.9 m. On May 6, 2104, 1,095 grams of 35 wt. % Uranine C dye (often called fluorescein) was introduced into the sinkhole. A skid loader excavated a basin in the area of the very small swallow hole. 1,100 liters (300 gallons) of fresh water was first injected into the bowl. Dye was injected next (Figure 6) followed by 2,300 liters (700 gallons) of water. Initially, drainage was very slow and the water ponded for nearly an hour; thereafter, it quickly flushed into the ground. (Johnson et al., 2014).

Another sinkhole, approximately 2 kilometers south of the Wiskow property was located on the Applen property to introduce dye (MN55:D00282) found at 557516 E / 4865569 N ± 3.3 m. The sinkhole is approximately 21 meters in diameter, 9 meters deep with a 1 meter swallow hole in the bottom. Injection included 750 liters (200 gallons) of water followed by 435 grams 17.7 wt % Rhodamine WT dye and 3,000 liters (800 gallons) of additional water (Johnson et al., 2014).

Twelve springs resurfing from the Prosser and Cummingsville Formations in the area were identified
for sampling. Two vulnerable domestic wells were monitored.

Rhodamine WT was traced from the Applen sinkhole (MN55:D00282) to the newly mapped Gasner Farmyard Spring (MN55:A00566). The dye took less than one day to reach the spring at a minimum velocity of 922 meters (3,025 feet) per day. Dye was detected in the spring for the remainder of the sampling period.

Uranine C was traced from the Wiskow sinkhole (MN55:D00983) to the newly mapped Pagel Spring (MN55:00567). The dye took two days to reach the spring at a minimum approximate velocity of 278 meters (913 feet) per day.

Rhodamine WT or Uranine C was not detected at any other sample location. Surface water entering the karst system through the sinkholes studied in the Wiskow Dye trace does not appear to leave the Prosser Formations. Dye was not detected in the lower Cummingsville Formation springs.

Case Study 2 – Studying a Sinkhole for Pollution Abatement

In the spring of 2013, a new sinkhole formed in proximity to an existing feedlot in the Orion Sinkhole Plain. Initially the owner tried to fill the sinkhole on his own; however in the spring of 2014 the sinkhole was reactivated at the surface. The farmer then contacted the Olmsted County Soil and Water Conservation District for assistance in correcting the potential pollution hazard.

In July 2014; Natural Resources Conservation Service (NRCS) personnel and an Olmsted County Soil and Water Conservation District employee met on site to perform a detailed survey of the sinkhole. Soil borings were taken to determine the depth to bedrock around the perimeter of the sinkhole. Multiple cross sections were taken with ground penetrating radar (Figures 7 and 8).

NRCS recommended the sinkhole be fully excavated to bedrock and locate the conduit system. The conduits would be covered with select stone drain material followed by a non-woven geotextile liner. The largest portion of the sinkhole filled with compacted engineered soils.

Case Study 3 – Blue Ridge Confined Animal Feeding Operation

Schoenfelder Farms owns a large confined animal feeding operation (CAFO) named Blue Ridge. Blue Ridge is centrally located in the sinkhole plain. The feedlot was initially permitted by the Minnesota Pollution Control Agency (MPCA) in 1991. The Schoenfelders’ have been researching solutions to modify the site so that it meets current pollution control laws. According to federal
feedlot laws, all CAFOs must comply with a “zero runoff” status and, therefore must eliminate or collect all water leaving the feedlot. Many proposals have been presented by Schoenfelder Farms and their private engineers to the MPCA, but location challenges have delayed completion of environmental corrective action for three years. In attempts to find a suitable site for a large manure storage structure, many locations have been evaluated at Blue Ridge.

Three primary challenges exist for locating the proposed storage area. Minnesota feedlot rules require the following:

- Due to the proximity of sinkholes within the plain, a liquid manure storage area (LMSA) cannot be constructed on most of the Schoenfelder property. According to Minnesota feedlot rules LMSAs must not be located within 91 meters (300 feet) of any sinkhole (Minn R. 7020.2005 Subp. 1.).

- The total size of the LMSA may not be larger than 950,000 liters (250,000 gallons) when four or more sinkholes exist within 305 meters (1,000 feet). The required volume for the facility is significantly more than 950,000 liters, and many sites would not allow for the 22 million liter (6 million gallon) structure that is needed at the Blue Ridge Site (Minn R. 7020.2100 Subp. 2. A).

- Minnesota feedlot rules require a minimum vertical separation to bedrock of 3 meters (10 feet) when the site has a capacity of 1,000 or more animal units (1 slaughter steer is equal to 1 animal unit). Most of the sinkhole plain has soil cover of less than 1.5 meters (Minn R. 7020.2100 Subp. 2. B. (3)).

Since the feedlot is located at the northern edge of the plain, the only potentially suitable location is 1,000 feet north in an open field. One small corner of the Schoenfelder property appears to allow for separation to sinkholes and vertical separation to bedrock. In the fall of 2014, test pits were dug with an excavator to determine separation to bedrock (Figure 9).

As of March 2015 research is continuing to find a solution that meets Minnesota Feedlot Rules and is a viable option for the Schoenfelders’.

**Case Study 4 – A Constructed LMSA to Protect Surface and Groundwater**

A manure contaminated runoff storage area was constructed in the fall of 2014 by a livestock producer with 275 animal units located at the headwater spring of Mill Creek (Figure 10). The structure was built to protect the stream from manure contaminated runoff. The LMSA also stores manure so that the producer can apply manure in a seasonal manner that maximizes crop nutrients and minimizes environmental risk. (Manure application considerations are described further in the next section.) Prior to construction manure contaminated runoff discharged directly into the creek.
a problem arise with the concrete liner. Perimeter drain tile is also installed around the footing of the LMSA. The purpose of the drain tile is two-fold. It controls groundwater around the LMSA and collects any potential leachate. The drain tile water is also inspected at the same location as the interior form-a-drain (Fryer, 2015, written communication).

Considerations of Manure Application on Karst

Manure application on the land surface is not necessarily a bad thing if it is done carefully and all laws, regulations and best management practices are followed. Crop nutrient prices have escalated exponentially in the past five years and manure contains costly nutrients that crops require. Detailed nutrient management plans assist producers with application method, timing and rates applied.

Following the severe runoff incidents in March 2013; the Olmsted SWCD developed manure application standards that are being presented to livestock producers in Olmsted County. In most cases the standards exceed Minnesota feedlot rule requirements. The standards are intended to limit the risk of manure contaminated runoff reaching groundwater in karst areas.

In regards to manure applications on snow covered or frozen soils; the watersheds of open sinkholes appear to be most vulnerable (Figure 11). Collaborating to identify and communicate these risks has been productive. In general the consensus with producers is that 91 meter (300 feet) may not adequate for protection, and that the entire watersheds of sinkholes should not receive manure applications.
Do not winter apply manure when weather forecasts include rain and snow melt.

Regarding manure applications, producers should use standards above 7020 rule requirements, including more restrictive setbacks during winter applications completed after December 31.

Avoid application to aged snow pack that is crystalline in nature and contains high moisture content.

Modify tillage practices to reduce the impact to exposed bedrock outcrops. Contour tillage should be used at all times.

Regarding newly acquired application acres:

- Producers should meet owners of adjacent properties before spreading manure at new sites, explain manure management practices and ask about well types and locations so that pro-active steps can be taken to reduce groundwater contamination risk.
- Keep manure application rates low at first when spreading at a location that is new to the operation or has had little or no recent manure history.
- When new land is acquired, talk to previous operators, if available, about abandoned wells, shallow bedrock, sinkholes or places where water disappears into the ground.
- Provide educational workshops to all livestock producers within active karst and close bedrock soils to communicate that vigilance and management is essential for protection of water resources.

More recommendations were developed:

- In karst areas (where carbonate bedrock is present), producers should identify and understand:
  - Areas where bedrock limits tillage or drain tile installation or drainage ditch depth.
  - Presence of sinkholes, springs, closed depressions in or near fields that may fill and drain rapidly. Sinkhole application setbacks are required by feedlot rules (Minn. R. 7020.2225, Subp 6. A-C.)
  - Areas where runoff water or tile discharge disappears into the ground.
  - Known or potential locations of abandoned wells or cisterns. Feedlot rules contain well setbacks including wells that are abandoned. (Minn. R. 7020.2225, Subp 6. A-C.)
  - Locations of wells used for any private, water supply wells
  - Historical water quality problems in the area, such as short- and long-term water quality/quantity problems related to the ground -water system.

**Continued Challenges**

The challenges continue for livestock producers protecting water resources in karst systems. Preventing all contamination of ground water is expensive and very difficult to achieve. However, it does not alleviate the responsibility that landowners have to greatly reduce the potential for contaminants entering aquifers.

Acute, catastrophic events may occur when karst features are unknown, when they develop quickly, or when they are underestimated by engineers and regulators. Tools such as electrical resistivity imaging will continue to be more efficient in detecting hidden karst features. However, implementing the new geophysics tools may not be required for construction of a manure storage area.
When engineered soils are placed on top of the bedrock, the material between the stored manure and bedrock is homogeneous with known compaction and density. There is, however, a risk that the interaction between surface and groundwater flow is impacted, creating preferential conditions for sinkhole development.

**Conclusion**

The four case studies outlined provide excellent educational, informational and examples for future engineering and operational proposals.

Case Study 1 is a detailed example how risk of water pollution exists, even when all applicable manure application guidelines are followed. The dye tracing study that followed the runoff incident indicates groundwater flow and resurgence.

Case Study 2 shows that dynamic karst terrain is difficult to predict, and that karst features may appear quickly. Survey and studies are valuable for remediation if the new or existing karst features present an elevated risk of groundwater contamination.

Proposed LMSAs may be 90 meters in dimension and constructed over many vertical bedrock joints. Currently, Minnesota feedlot rules require:

> “A minimum of two soil borings within the boundaries of the proposed manure storage area for the first one-half acre of surface area. A minimum of one additional location is required for each additional one acre of surface area for the manure storage area.” (Minn. R. 7020.2100, Subp 4. A. (2))

The soil borings are only one small picture of the karst system below and rarely provide enough information to indicate the presence of karst features.

Figure 13 provides an illustration of an LMSA constructed over carbonate bedrock to meet current Minnesota feedlot rules. Clearly there is a risk of missing potential dangers beneath soil covers greater than 10 feet.

Bedrock removal, in order to create required separation has occurred in many instances in Olmsted County (Figure 14). The outcomes of removing bedrock in order to achieve separation are not clear. Potential advantages may be that a clean bedrock surface can be inspected for areas of preferential flow or enlarged solution. When engineered soils are placed on top of the bedrock, the material between the stored manure and bedrock is homogeneous with known compaction and density. There is, however, a risk that the interaction between surface and groundwater flow is impacted, creating preferential conditions for sinkhole development.

**Figure 12.** Fracture traces in the Ordovician Prosser Formation. Rochester Township, Olmsted County (Peter et al., 1976).

Large, weathered joints that commonly cut through many bedrock formations are typically spaced 9 – 30 meters apart (Alexander, et al., 1995). These joints are usually sediment filled as seen in Figure 12. They may also be enlarged through solution and become an integrated conduit system capable of carrying large amounts of water; therefore increasing probability of sinkhole formation.

**Figure 13.** Illustration of an LMSA constructed over carbonate bedrock with an undiscovered void (Larsen, 2015).

The soil borings are only one small picture of the karst system below and rarely provide enough information to indicate the presence of karst features.

The outcomes of removing bedrock in order to achieve separation are not clear. Potential advantages may be that a clean bedrock surface can be inspected for areas of preferential flow or enlarged solution. When engineered soils are placed on top of the bedrock, the material between the stored manure and bedrock is homogeneous with known compaction and density. There is, however, a risk that the interaction between surface and groundwater flow is impacted, creating preferential conditions for sinkhole development.
Case studies 3 and 4 are examples of real investment into environmental protection from multi-generation family farms. The setting and engineering of practices may be extensive, costly and take time.

Ongoing manure application and engineering challenges need to be continually discussed and researched between livestock producers, engineers, scientists and regulators so that pollution reduction practices can be designed, implemented and funded.

Building excellent relationships with producers has been successful in Olmsted County. Livestock producers are making investments and taking action. Producers are an essential component of the mid-western economy and assistance with information, funding and resources will help protect the environment and keep farms profitable for future generations.

References

Figure 14. Limestone bedrock in process of removal from the floor of an LMSA (Larsen, 2014).

EVAPORITE GEO-HAZARD IN THE SAURIS AREA (FRIULI VENEZIA GIULIA REGION - NORTHEAST ITALY)

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Abstract
Evaporite sinkholes represent a severe threat to many European countries, including Italy. Among the Italian regions, of the area most affected is the northern sector of Friuli Venezia Giulia Region (NE Italy). Here chalks had two main depositional periods first in the Late Permian and then during the Late Carnian (Late Triassic). Evaporites outcrop mainly in the Alpine valleys or are partially mantled by Quaternary deposits, as occur along the Tagliamento River Valley. Furthermore, evaporites make up some portions of mountains and Alpine slopes, generating hundreds of karst depressions.

This paper presents the preliminary results of the research activities carried out in Sauris Municipality where sinkhole phenomena related to the presence of gypsum are very common.

Field investigations were devoted to recognition, mapping and classification of evaporite sinkholes. To recognize sinkhole phenomena, the preliminary steps included the analysis of historical documents collected in archives, the analysis of aerial photos and Airborne Laser Scanning (ALS) surveys. The integration of the above-cited activities allowed a preliminary identification of the phenomena, which were later validated by detailed field surveys.

All the collected data populate a geo-database implemented for a project funded by the Geological Survey of Friuli Venezia Giulia Region. The objective of this project is to inventory and classify the sinkholes associated to evaporite rocks.

Introduction
Subsidence phenomena associated to the presence of evaporite rocks are common in Europe. Evaporite sinkholes affect the central and northern part of England (Cooper, 2008), Lithuania (Taminskas and Marcinkevicius, 2002), NE Spain (Gutiérrez, 1996; Gutiérrez and Cooper, 2002; Guerrero et al., 2002; Guerreiro et al., 2004; Gutiérrez et al., 2008) and Albania (Parise et al., 2004; Parise et al., 2008).

As reported by Nisio (2008) and by Caramanna et al. (2008), sinkholes occur also in Italy where are distributed along the whole peninsula, especially in some regions, such as Sicily, Lazio, Campania, Alto Adige, Puglia, and Friuli Venezia Giulia.

The Friuli Venezia Giulia Region is located in the NE part of Italy and covers an area of 7,858 km². Here are present 221 municipalities of which approximately 40 coexist with geo-hazard problems associated to the existence of outcropping or mantled karstifiable rocks. Limestones and dolostones represent approximately 24% of the whole regional territory, whereas evaporite rocks do not exceed 1% (Figure 1). In all those areas, where evaporites and limestones outcrop or are mantled by quaternary deposits or other rock types, sinkholes
Evaporite sinkholes were recognized and classified by means of traditional activities such as desk analysis and field surveys.

Study Area

The Sauris territory has an area of about 42 km$^2$, and a mean elevation of 1,212 m with an inhabitant density of 10 persons/km$^2$.

Evaporite sinkholes were recognized and classified by means of traditional activities such as desk analysis and field surveys.

Study Area

The Sauris territory has an area of about 42 km$^2$, and a mean elevation of 1,212 m with an inhabitant density of 10 persons/km$^2$. Even if the average elevation is not so high, in the northern side of Sauris, some peaks as
Bivera Mountain reach an altitude of 2,474 m. In this municipality, evaporite rocks are not so common at the surface because they are overlying Werfen Formation or mantled by Quaternary deposits. The presence of sinkhole phenomena is historically known in this area (Calligaris et al., 2009). Tens of sinkholes were recognized on the valley bottoms, over the top of the ridges and in the middle of the slopes. Their presence partially compromises urban expansion and consequently affects land use planning. In the villages and settlements, this situation represents a severe geo-hazard for human facilities and inhabitants (Figure 2).

**Geological Overview of the Area**

From a geological viewpoint, Bellerophon Formation and Werfen Formation are the most important. The Bellerophon Formation is a regionally extensive unit, outcropping from Slovenia to Veneto Region. The intensity of the alpine tectonic deformations strongly affected its stratigraphic continuity. In fact, it is difficult to find a comprehensive section and it frequently appears cataclastic (Venturini et al., 2006).

It is difficult to define the original thickness of this unit due to the presence of several thrusts and high rates of gypsum dissolution (Buggisch and Noè, 1986). The unit is mainly outcropping in a wide belt coincident with the valley bottoms.

In the study area, the main discontinuity is the E-W oriented Sauris Fault (Figure 3), which follows the regional structural trend. This fault is 40 km long from West to the East of the Region reaching the Tagliamento River where is present its further eastern portion. The above-cited fault permits Permian units to overlap the Triassic rocks. The presence of gypsum at the base of the northern overthrusted units facilitated the process. The Sauris Fault is also crossed by secondary faults.

The Bellerophon Formation incorporates two different members: at the base gypsum alternates with black dolostone (thickness of about 60 m) whereas the upper member consists of dolostones and black limestones (200 m). The plastic behavior causes strong deformations of the evaporite member that for this reason seldom outcrops. Conversely, the upper Member widely outcrops (Carulli, 2006; Venturini et al., 2006).

In the northern part of the study area, the Bellerophon Formation is capped by Werfen Formation (Early Triassic). Werfen Formation incorporates six different members. The lower member consists of an oolitic limestone with an average thickness less than 7 m. Middle members are made up by limestones and marls, dolomitic limestones, marls and pelites, alternating calcareous sand and mud with a thickness of approximately 200 m. The Formation ends up with a member of fine-grained violet sandstones and pelites, reaching a thickness of about 200 m (Carulli, 2006; Venturini et al., 2006).

Moraine deposits partially mantle the valley bottoms and the flat parts of the highlands.

![Figure 2. Aerial (A) and oblique (B) views of a sinkhole located near Sauris Village.](image-url)
The above-mentioned activities allowed us to identify the surface morphologies and later to recognize the sinkholes. All the desk data were validated by field surveys (Figure 2). Particular attention was devoted to the selection of the sinkhole classification. In the present paper, we used the methodology developed by Gutiérrez et al. (2008), which was the most suitable for the encountered evaporitic phenomena.

We recognized and classified 73 sinkholes (Figure 3); the caprock sagging sinkholes are dominant (49) and are common in the northern part of the study area.

Methods and Results
To recognize sinkhole phenomena, the preliminary steps have included the analysis of historical documents collected in archives and the study of scientific papers and technical reports. These investigations have permitted us to define the geological setting of the Sauris area and to outline the possible locations of sinkholes associated with the carbonate and evaporite rocks.

These preliminary stages were integrated with the interpretation of aerial photos and ALS surveys acquired recently by Regional Civil Protection.

Figure 3. Spatial distribution of sinkholes affecting Sauris Municipality. The different colors indicate the types of sinkholes. 28= Quaternary deposits; 23= Moraine deposits – Late Pleistocene; 12a= Carbonates alternated with marls – Late Trias; 11= Carbonates alternated with marls – Late Trias; 10b= Vulcanites – Trias Middle; 10a= Sandstones and shales (flysch) – Middle Late Trias; 9= Massive carbonates – Middle Late Trias; 8a= Massive carbonates – Middle Trias; 8b= Sandstones and shales (flysch) – Middle Trias; 7= Sandstones and shales (flysch) Werfen Fm. – Early Trias; 6c= Carbonates alternated with marls (Bellerophon Fm.) - Late Permian; 6b= Evaporites (Bellerophon Fm.) – Late Permian. In red the main thrusts (after Carulli, 2006); SF= Sauris fault.
are associated with heavy rainfall, which often exceeds 2,000 mm per year in northern Friuli, and mainly by the presence of a torrent, which accelerates the dissolution processes and the gypsum erosion.

The cover suffosion sinkholes (5) and cover collapse sinkholes (2) are mainly located near Sauris di Sotto and involve the Quaternary glacial deposits.

Conclusions
The results of the integrated analysis has produced an inventory of sinkholes associated with the evaporite rocks in the Sauris Municipality. Here, different types of karst phenomena occurred and involve mainly the Werfen Formation, which overlies the evaporites of the Bellerophon Formation.

The sinkholes located in the surroundings of the top of mountains situated in the northern part of Sauris Municipality, affect a 200 m thick of poorly–karstifiable formation.

This is due to the different mechanical properties of the Werfen and Bellerophon Formations. Conversely,
bedrock collapse sinkholes are scarce but are characterized by huge sizes and depths. The maximum depth exceeds 35 m.

The spatial distribution of sinkholes does not coincide with any particular structural alignment even if the investigated area is crossed by several regional faults, which are oriented approximately E-W.

Thanks to the financial support of Regional Geological Survey, the research activities here presented are not limited to Sauris area but include all the municipalities of Friuli Venezia Giulia Region. At present, we have inventoried approximately 200 sinkholes, spread among Forni di Sopra and Tolmezzo villages (Figure 1). Type, size, depth, and other major morphometric characteristics of each phenomena populate a Geodatabase. The latter is crucial to assist local authorities to recognize areas affected by geo-hazard associated to evaporite rocks and can be used for land-planning purposes.

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Abstract
Cover-collapse sinkholes are forming with increasing frequency under buildings. Analyses of sinkhole distribution in Beacon Woods, Florida, preliminarily indicate their occurrence is an order of magnitude greater in urban versus undeveloped areas, suggesting the structures themselves are enhancing the collapse process. The most likely causes are induced recharge via at least one of two sources. First, runoff and drainage from roads, structures, and impoundments that is not adequately dispersed will promote sinkhole development. Second, leaking water, sewer, and septic systems beneath or adjacent to a structure will also promote collapse. The process of cover-collapse from induced recharge is well understood. However, building codes generally do not require drainage and structural engineering practices that would reduce induced recharge and thus reduce the risk of collapse. This paper proposes engineering practices that measurably restrict the accidental discharge of municipal water through leaking subgrade drainage systems or the deliberate discharge of stormwater runoff, induced shallow groundwater recharge from retention ponds and septic drainfields, or heavily-irrigated land use. We recommend these practices be incorporated into building codes and ordinances to reduce induced sinkhole development in areas prone to cover-collapse.

Introduction
Tragedy struck. Mr. Jeff Bush was sleeping in his home the night of 27 February 2014 in Seffner, Florida, when a sinkhole opened under his bedroom and swallowed him. His body was never recovered.

Sinkhole collapses are often viewed with fascination and a certain excitement about their potential danger. The tragedy in Seffner stands as proof that their hazard should not be underestimated. But what was the cause of that collapse? Media interviews mostly pointed to groundwater pumping, which could have lowered the water table, piping soil and sediment downward, resulting in the collapse. This is certainly plausible, as are natural processes unrelated to human activities. However, the opening of the sinkhole directly beneath the Bush home may indicate another origin.

Cover collapse sinkholes from induced recharge are perhaps the most widespread type of collapse. Collapse from groundwater withdrawal is more widely publicized due to the dramatic collapses in Florida, but induced recharge collapses occur frequently in Florida and far beyond. The location of the Seffner collapse, under the Bush home, suggests a recharge-induced origin potentially from leaking water or sewer pipes, roof runoff, and/or an adjacent shallow retention pond. Unfortunately, evidence of the mode of this and other collapses is often lost in the collapse, non-forensic excavation, and/or from filling or other form of remediation.

Many books and papers describe geotechnical measures to remediate sinkholes and prevent them once subsidence is detected (e.g. Sitar, 1988; Sowers, 1996; and the 13 volumes of Sinkhole Conference proceedings, 1984-2013). This paper promotes a proactive approach that we believe will prevent some collapses from ever occurring. We begin by describing the general causes of recharge-induced collapses. Next we use a case study to demonstrate how and why they occur with greater frequency under and around buildings, making them potentially deadlier than other types of sinkholes. While the case study area is in Florida, this paper is not...
focused on Florida so details of some geologic processes are kept general to be relevant to most regions where cover collapses occur. We end by proposing building codes to minimize such collapses.

**Origin of Recharge-Induced Cover Collapses**

Cover collapse sinkholes form where thick mantles of soil, regolith, and/or sediment overlay cavernous bedrock. With time, these unconsolidated sediments move down into conduits in the bedrock. If the sediments move slowly and have little structural strength, the land surface above gradually subsides. Where the sediments are more structurally competent and may also move more rapidly, a cavity develops within until it becomes unstable and abruptly collapses.

The sediment movement results from gravity and the flow of water and occurs in two ways, mechanisms that have been well described and understood for many years (e.g. Newton, 1987; Galloway et al. 1999). First, if the water table of a karst aquifer extends into overlying sediments, declines in groundwater levels below the bedrock will wash some sediment into bedrock conduits. Other sediment will slump and fall into the cavities due to its increased water-saturated weight. Repeated rises and falls in the water table will carry more sediment into bedrock conduits where they are flushed away and through the aquifer. Eventually, subsidence or collapse may be seen at the surface.

This paper focuses on the second mechanism of sediment movement: induced recharge. In this case, the karst water table generally remains within the bedrock, below the unconsolidated sediments. The sediments move into the bedrock conduits when water sinks consistently into the ground at a particular location, saturating the sediment and carrying it downward. As long as water flows in that location, whether constantly or intermittently, underlying sediment will be lost into the aquifer and a subsidence or collapse sinkhole will eventually form (Figure 1).

Many induced sinkholes form along highways, where road runoff sinks into adjacent drainage channels (i.e. swales). Most sinkholes in non-karst areas are induced, often by leaking water and sewer pipes that can supply water to saturate sediments and carry sediment away to create cavities, the most famous and fatal of which were up to 30 m in diameter by 60 m deep, formed in Guatemala City in 2007 and 2010 and killed six people (Hermosilla, 2012).

Less recognized are sinkholes resulting from recharge induced by homes and other buildings. Recharge from roof runoff, if not directed away from a building, infiltrates soil along the building’s foundation, putting the building at risk from subsidence or collapse. Unseen induced recharge also occurs beneath buildings from leaking water and sewer pipes. While leaking water pipes might be indicated by water usage on monthly water bills, sewage outflow is not measured and leaks are not detected without direct testing (Figure 1B). With the Seffner and other collapses occurring directly beneath buildings, the possibility that these buildings and their infrastructure caused some of the collapses must be considered, especially as global increases in population in sinkhole-prone areas may be putting more people at risk.

**Case Study: Beacon Woods, Bayonet Point, Pasco County, Florida**

To test the hypothesis that roads and other urban infrastructure may result in a greater frequency of cover collapse...
collapses than undeveloped land, we selected Beacon Woods as our study area. Located approximately 50 km northwest of Tampa, Florida, the district of Beacon Woods is at the north end of the community of Bayonet Point, about 6 km east of the Gulf of Mexico. Beacon Woods study area covers roughly 8 km² of residential and light commercial developments planned in the 1970s and mostly built out by 1980. With the exception of a golf course, the land use is relatively homogenous across the study area, which is defined by US Highway 19 to the west, Hudson Avenue to the north, Fivay Road and Little Road to the east, and State Road 52 to the south. This study area was selected because of the age of the residential developments, the detailed property/permit records readily available online, the karst topography, and the mapped cave and sinkhole system running its length. The blue lines overlain on Figures 2 through 5 are simplified footprints of the documented sections of this cave system.

Historical aerial photography of the vicinity was downloaded and reviewed, the oldest dating from 1941 when very few buildings had been constructed in the area (Figure 2) (Aerial Photography: Florida Collection [Internet]; Florida Department of Transportation’s Aerial Photo Look-Up System [Internet]). According to the pre-development topography, the ground surface ranged in elevation from about 6 m (20 feet) along Fivay Road down to 3 m (10 feet) to the west; below a depth of 1.5 m (5 feet) the closed drainage features were water-filled (all elevations relative to the National Geodetic Vertical Datum).

In the western half of the study area, the shallow geologic unit is the Suwannee Limestone (Arthur, 1993). While a few limestone outcrops occur in this part of the study area, most observed limestone is in the form of saprolitic boulders, many excavated and placed as decorations during development. The cover soils are fine sands, with minor amounts of sandy clay collected in occasional isolated pockets. Moving east, an overlay of sand dunes (Arthur, 1993) provides topographic relief, guiding surface drainage. The multitude of closed depressions, some water-filled and some dry, provide an indication of the high degree of karstification of the underlying limestone.

While groundwater levels have fluctuated and refinements in surveying and mapping have improved in accuracy, there is an undeniable gradual increase in the areal extent of surface waters evident across the three major map revisions published by the US Geological Survey since 1954 (Figure 3). Most of the wetlands and water-filled features in Figure 3 are now developed

Figure 2. Simplified Beacon Woods cave system (blue) and study area (yellow) overlay on pre-development ground surface (United States Department of Agriculture Natural Resource Conservation Service., 2015. [Cave outline adapted with permission from map by Southeast Exploration Team, personal communication]).

Figure 3. Simplified cave system (blue) and study area (yellow) overlay on pre-development topography (U.S. Geological Survey, 1954).
as drainage canals and stormwater retention ponds or greenspace in the most current topographic map (Figure 4).

The large Beacon Woods Cave System is present beneath the center of the study area, trending, and its groundwater flowing, south to north. Multiple sinkholes provide entrance into the system for both exploration and a considerable volume of surface water from Bear Creek. The mapped portions of the cave system have an average water depth of 46 m, although some sections are considerably shallower. The lower sections of the cave contain brackish water, with diffuse haloclines at depths consistent with the Ghyben-Herzberg principle. A resurgence of the cave system is approximately 4.3 km north-northwest of Smokehouse Pond (Figure 3), at the former Hudson Springs, while dye tracing in the 1960s linked a further upstream yet impassable sinkhole with a spring in the Gulf of Mexico (Wetterhall, 1965).

Compaction grouting of residences affected by collapse or subsidence in the vicinity of the underlying cave system has a high potential, by design, of intersecting some zone of increased permeability such as a fracture, joint, or eroded bedding plane, as well as a somewhat lesser possibility of directly impacting a cave passage. While the impact to endangered cave biota cannot be understated, the more pressing economic concern may be the potential to trigger a collapse of a large section of cave and drastically reducing the capacity for drainage of the basin.

To correlate the occurrence of cover-collapse sinkholes, the predominant type in the region, with urban infrastructure, we compiled “potential ground settlement investigation” and “ground settlement repair” permits filed for the properties within the study area filed in the last 25 years in the Pasco County Florida Public Access to Permit Applications database (Pasco County Florida Public Access to Permit Applications [Internet]). Our analysis found about 750 investigations and 650 repairs documented with the Pasco County Building Department, which accounted for over 6% of all building permits recorded for this 25-year time period. The building coordinates were filed with each of the repair permits and their locations are plotted as red squares on the aerial photograph in Figure 5.

Since 2008, the Pasco County Building Department has recorded the specific details from the repair reports. After extensive data mining we determined that a staggering $33,801,000 dollars (US) have been spent from March 2008 to March 2015 in compaction or slurry grouting, chemical grouting, and/or underpinning the structures within the 8 km² study area. Also, the subsidence incident database compiled by the Florida Geological Society as of October 2014 was sorted for the study area and the limited descriptions were evaluated for applicability to our analysis (Subsidence incidents...
Proposed Building Code to Minimize Recharge-Induced Collapses

Considering the above results, we believe that building codes for sinkhole-prone areas should address the likely dramatic increase of induced collapses around roads and urban infrastructure with measures that will reduce or prevent their occurrence. For the study area, we reviewed the historic aerial photography and found the shallow native soils were reworked or sandy fill soils imported to raise grades below foundations and roadways in order to install a system of curb and drainage pipe to channel surface runoff to isolated retention ponds within nested residential streets or to long swales bordering major roadways. We determined that the typical sources of water infiltration that could be mitigated include:

- roof runoff
- street drainage from curb to culvert
- automatic lawn irrigation systems (operational without a rain gauge or if broken and leaking)
- effluent from septic drainfields
- leaking plumbing below or beside buildings
- obsolete or un repaired shallow irrigation wells
- unlined stormwater ponds
- leaking swimming pools, and
- wastewater spray fields.

The following code modification suggestion is generated from a review of the current 2010 Florida Building Code - Building, specifically Chapter 18 – Soils and Foundations (International Code Council, Inc., 2010). Though shown here as specific to this statute, it is written for general adaptation to any building or roadway construction code in any karst area prone to cover collapse sinkholes. Under “1803.5, Investigated conditions,” many deleterious soil and groundwater conditions (even seismic) are detailed or reserved, but cover-collapse isn’t among them. There may be future avenues available with an addition of a hypothetical Section 1803.5.13 Cover-Collapse Risk Zones:

Where historical photographic evidence of cover-collapse sinkholes exist or where geophysical surveys and subsurface explorations at the project site indicate substantial risk of sinkhole incidence, the building official shall be permitted to deem the site a no-build zone or require a registered design professional to demonstrate that the intended construction will measurably restrict the accidental discharge of municipal water through leaking subgrade drainage systems, the deliberate discharge of stormwater runoff from buildings, roads, parking lots, or other constructed impervious cover, induced shallow groundwater recharge from retention ponds and septic drainfields, heavily-irrigated land use such as plant nurseries, golf courses and water reclamation sprayfields, and/or swimming or decorative pools.

Some terms in the above proposed code would need to be defined, often in accordance with local or state statutes. The proposed code inherently encourages design professionals working in these high risk zones to develop stormwater runoff management options that emphasize detention and evaporation instead of the typically highly-desired rapid infiltration in concentrated areas. This could be accomplished with lined shallow ponds, which would create the added benefit of wildlife habitat between neighborhoods. Also, specifying leak-proof double-walled piping and reinforced fittings as well as the removal or restriction of septic drainfields by requiring municipal sewer lines and transfer stations outside of the high risk zone would likely reduce induced recharge-related subsidence as well as improve water quality. For the homeowner currently living in a high risk zone, the addition of a rain barrel or other catchment system at the discharge point of each gutter would delay the infiltration of roof runoff and reduce the risk of soil settlement or erosion on their property.

Conclusions

Sinkhole collapse and subsidence are often treated as sensational fascinating events of mysterious origin by the media. That perception has often masked the fact that the sinkhole process is well understood and action can be taken to minimize its occurrence.

Our findings in the Beacon Woods study area show that close to $6 million/year have been spent in that 8 km² study area in sinkhole and subsidence remediation and associated repairs. The February 2014 collapse under
the Bush home in Seffner, Florida, proved the cost can be much greater.

Further, our research indicates a potentially greater than order of magnitude increased frequency of sinkhole and subsidence development in association with urban infrastructure over undeveloped land, strongly indicating a causal relationship likely due to induced recharge. While additional research will better quantify this relationship, we believe our results thus far should prompt government agencies charged with public protection through roadway and building codes to adopt codes that would minimize induced cover collapse and subsidence, such as through the example code we have provided.

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Pasco County Florida Public Access to Permit Applications [Internet]. [Pasco County Building Department]; [cited March 8, 2015]. Available from: https://secure.pascocountyfl.net/bccpapa/Default.aspx


CARS AND KARST: INVESTIGATING THE NATIONAL CORVETTE MUSEUM SINKHOLE

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Abstract

On February 12th, 2014, a sinkhole occurred at the National Corvette Museum in Bowling Green, Kentucky. The collapse happened inside part of the building known as the Skydome and eight Corvettes on display were lost into the void that opened in the concrete floor. In this region of Kentucky, known as the Pennyroyal sinkhole plain, subsidence and cover collapse sinkholes are commonly found throughout the landscape. This iconic karst region in the United States is also home to Mammoth Cave, the longest cave in the world, and thousands of other caves and karst features. Investigation of the sinkhole collapse began immediately while the Corvettes were extracted from the debris cone inside the void. Techniques used for investigation included water jet drilling, downhole cameras and drone footage, a microgravity surface survey, and mapping of the void and accompanying cave. After exploration of the sinkhole by karst researchers and compilation of the data, the cause of the sinkhole was determined to be a cave roof collapse in a breakout dome. The cave underlying the collapse is about 220 ft. (67 m) long and 39 ft. (12 m) wide on average with an average depth of 65-85 ft. (20-25 m). The structural integrity of the bedrock (thinly interbedded limestone and chert located at a contact between two major limestone units) is lacking in the area. Talus and breakdown are abundant in the cave in which the sinkhole formed. The progression of the roof failure likely occurred over a long span of time, eventually giving way due to a variety of conditions, including speleogenetic and climatic factors. Current remediation is underway and involves filling the sinkhole with gravel and sand, then installing a micropile supported concrete slab floor under the building. Future changes to the structure will be monitored to detect any activity.

Introduction

On February 12th, 2014, a sinkhole occurred at the National Corvette Museum in Bowling Green, Warren County, Kentucky. The collapse happened inside part of the building known as the Skydome, which is a large circular structure connected to the main building where rare Corvettes are displayed (Figure 1). On the day of
the collapse, eight Corvettes on display were lost into the void that opened in the concrete floor.

In this region of Kentucky, known as the Pennyroyal sinkhole plain, subsidence and cover collapse sinkholes are commonly found throughout the landscape (Ford and Williams 2007, Palmer 2007, North et al. 2014). This is an iconic karst region in the United States and home to Mammoth Cave, the longest cave in the world, and hundreds of other caves and karst features.

In Bowling Green, sinkhole collapses are not uncommon, with many having been mapped by the Kentucky Geological Survey and others that occur on a regular basis, though usually on a smaller scale and often unreported (KGS 2015). There are also over 200 documented caves in Warren County, Kentucky (KSS 2014), lending merit to the potential for sinkholes and cave collapses given the highly karstified nature of the bedrock. Despite this potential, sinkholes do not usually pose a high threat to the community.

**Sinkhole Types**

In many classic karst settings, most sinkhole collapses form from regolith arch failure, where cohesive soils, usually dense clays, create an arch that eventually gives way from below due to undermining caused by spalling of materials into voids during water movement (i.e. storm events large enough to cause infiltration). These types of sinkhole often form rapidly and can cause immediate threat to life and property if they occur in developed areas. Other common sinkholes in south-central Kentucky occur as subsidence landforms, or closed depressions, where surface soils slowly spall into cavities below over long periods of time. These features create a gentle depression in the landscape that sometimes has a “throat” in the center, but usually pose little risk to life or property. A third type of sinkhole, though less common, results from cave roof collapse. Though not as well documented, these types of sinkhole collapses are found throughout the study area, including past examples like the Dishman Lane sinkhole (Kambesis et al. 2003) and several collapsed cave entrances, such as

![Figure 1. Initial sinkhole collapse: The original collapse on the day it occurred at the National Corvette Museum in Bowling Green, Kentucky.](image-url)
Lost River Cave’s main entrance and Crumps Cave in Smith’s Grove, Kentucky (Crawford et al. 1989; Polk et al. 2013). As structural integrity of the overlying bedrock weakens through geologic time, the cave roof forms a breakout dome, or cantilevered dome, which eventually fails in a similar manner as a cohesive soil arch, with a sudden, catastrophic collapse of the roof and overlying materials into the cave below (Loucks 2007). These types of sinkholes are hard to detect and predict, but examples can be found throughout the world of both small and large feature of this type.

**Study Area**

The National Corvette Museum is located in the Pennyroyal Plateau, which encompasses the Pennyroyal sinkhole plain of south-central Kentucky. The Skydome structure sits atop the contact between the Ste. Genevieve and St. Louis limestones (Figure 2), both of which are Mississippian aged formations that comprise a large majority of the cave-forming bedrock in the region, including hosting many of Mammoth Cave’s passages (Palmer 1981, Palmer 2007). These formations are often thinly bedded at the contact and dip gently toward the north. The Corydon ball chert layer is present and located at the contact between the Ste. Genevieve and St. Louis limestones. Sinkholes, both shallow and deep, are prominent features throughout the landscape and vary in depth based on the rock unit in which they form (Howard 1968). Adjacent to the Skydome to its south and northwest are large subsidence sinkholes. The southern sinkhole holds water throughout the year due to compacted clays or an impermeable chert layer at its bottom, or a combination of both.

Surface water flow is absent in much of the area due to the highly karstified landscape. Water is rapidly internally drained to the subsurface through various conduits until reaching the water table below. The regional aquifer level is often located up to 160 to 200 ft. (48 to 62 m) below the surface. Precipitation averages 51 inches (1,300 mm) per year and the climate is humid-subtropical, though the average annual temperature is near 55°F (13°C).

**Investigation Methods**

The sinkhole investigation began immediately after the collapse using a drone fitted with a camera to determine if there were additional voids or passages beneath the sinkhole, but the data were inconclusive. Micropile drilling to support the Skydome structure and the “spire” in its center, which provides structural support commenced and provided an opportunity for additional investigation. Downhole cameras were used in the holes drilled for the micropiles in an attempt to identify any additional cavities or passages extending from the main sinkhole collapse. Data recorded during water jet drilling of the micropiles allowed for generalized fence mapping of the drill logs, which provided basic information on depth to competent bedrock and additional voids surrounding the main collapse. Subsequently, a microgravity survey on a 10-foot grid (3-m) was conducted in the area surrounding the sinkhole and a buffer area outside the Skydome. Compilation of these data indicated the need for further exploration due to the possibility of additional voids extending from the main sinkhole opening away from the debris cone.

Prior to exploration and mapping, construction company Scott, Murphy & Daniel, LLC worked with karst scientists from Western Kentucky University (WKU) to develop a plan for removing the eight Corvettes and stabilize the sidewalls of the sinkhole, which had weakened and fractured concrete slabs cantilevered beyond the edge of the hole and over loose sediments. Construction workers were trained on vertical caving techniques and the Corvettes were removed using a combination of techniques to strap them to 36-ton (36,000 kg) crane...
booms and lift them from the sinkhole. All eight were recovered with varying amounts of damage, with some completely destroyed from the large breakdown that fell into the hole on top of them (Figure 3).

After retrieval of the Corvettes and stabilization of the hole, a team from WKU and EnSafe explored the sinkhole and discovered additional cave passages (Figure 4). A grade 5 cave survey was undertaken to map the sinkhole and cave passages. This was tied in to the surface engineering survey in order to better determine possible areas of concern and to develop a remediation plan for repairing the Skydome floor and any other structures outside the building as needed.

**Discussion**

The oval-shaped opening to the sinkhole measured approximately 36 ft. (11 m) at its longest axis, which is oriented just east of due north. Microgravity and drill log investigations, combined with visual clues, revealed the possibility of additional cave passage extending from the sinkhole to the north and south. Exploration and survey of the sinkhole concluded that it was a collapsed portion of a cave roof that failed and formed the sinkhole opening. The cave measures 220 ft. in length (67 m) and averages 40 ft. (12 m) in width. The deepest location is approximately 80 ft. (25 m) below the floor of the Skydome. The passages extend beyond the Skydome structure and trend toward existing sinkholes to the north and south, likely already collapsed portions of the same relic cave system. This verified the data from the microgravity survey, which indicated the potential for voids in these areas.

Inside the cave, the debris cone consists mainly of large breakdown and weathered limestone, with some overlying soils. The limestone comprising the cave walls was thinly bedded. The sinkhole is located at the contact between the St. Louis and Ste. Genevieve, where the Corydon ball chert layer (present in the ceiling, Figures 5 and 6) created conditions for instability over the span of the breakdown dome that failed. As the layers of rock collapsed over time, the breakdown dome continued to migrate upward toward the floor of the Skydome and likely was already fairly thin prior to its construction, as this would have been a slow geologic process. Additional wedging from calcite mineral formation between

**Figure 3.** Remains of a Corvette: After extraction from the sinkhole, some Corvettes were repaired, while others were damaged beyond repair. Three were fixed by General Motors and private funds, and the others are on display at the Museum.

**Figure 4.** Entrance to southern cave passage: Brian Ham of EnSafe, Inc. stands at the entrance of the southern cave passage. The cantilevered breakout dome can be seen as the sinkhole walls migrate upward.

**Figure 5.** Chert layer: Remnants of a layer of Corydon ball chert, which appears to be cleaved off from where the surrounding limestone has broken away, revealing the impurities that amplified the rock’s weakness.
the limestone beds likely created further weakness in the structure of the rock. There are many stylolites found within the limestone beds as well, along the planes of which additional points of weakness exist. These lateral stylolitic seams can enhance fluid flow and dissolution along their path, where delamination and cracking can occur (Heap et al. 2013), as was observed in the cave.

Over time, a breakout dome formed and migrated upward, thinning the bedrock support and creating a situation wherein structural failure and the sinkhole collapse was imminent.

Hayward-Baker, Inc. (Nashville, Tennessee) was consulted and a plan was formulated to repair the sinkhole and Skydome using micropiles to support a concrete slab floor (Figure 7). The debris cone was first smoothed flat and a one-foot (0.3 m) thick concrete slab was poured, then a double layer of metal sheet pilings were laid horizontally over it and cut to fit the shape of the sinkhole walls. The sinkhole walls were covered in specific areas with shotcrete to stabilize them further. The sheet pilings served as a base for filling the remainder of the hole to the floor with 4000 tons (3.5 x 106 kg) of manufactured sand to support the concrete floor. The primary purpose of the sheet pilings is to block the manufactured sand from filling the north and south passages. Its function is to provide support to micropile drilling equipment and temporary support for structural concrete slab construction. Holes were cut in the sheet pilings to drill the micropiles down to the bedrock. The micropiles are 7 inches (17 cm) in diameter (Figure 7) and will be placed at an average depth of 141 ft. (43 m), with a range of 130 to 200 ft. (39-55 m) below the floor to reach competent bedrock and will be tied into the concrete slab floor to provide support. The 46 micropiles are installed on a 20 x 25 foot (roughly 6 m) grid under the Skydome; there are 23 existing micropiles in place under the footprint of the building and the spire in the center.

Once the construction is complete, the Skydome structure and floor will be independently supported by the micropiles and thus protected from any further collapse or subsidence should either occur in the future. A 4-foot diameter (1.2 m) manhole was left in the floor to access the southern cave passage for additional scientific monitoring and also to integrate into an exhibit. The Corvette Museum is working on an interactive exhibit to provide visitors an educational experience about the sinkhole’s development and repair, as well as information about karst landscapes.

**Conclusion**

The sinkhole at the National Corvette Museum on February 12th, 2014 caused damage to the eight rare Corvettes that fell in the hole. Fortunately, the cars were recovered and study of the sinkhole was possible during this process. Investigation of the sinkhole using multiple methods revealed it was part of a larger cave system and aggregated to increase the confidence that the feature is limited to the vicinity of the Skydome. Compilation of the various data, cave map, and observations provided the explanation of the sinkhole resulting from the failure of the cave’s roof, which was made up of thinly-bedded and impure limestone. While not as common, these types
of sinkholes do happen throughout south-central Kentucky and elsewhere in the world. This case study provides one of the more well documented examples of a cave roof failure inducing a sinkhole collapse. Additionally, the use of geophysical techniques combined with ground-truthed survey of the cave further validates the use of microgravity as a reliable and robust technique for detecting subsurface voids in certain karst regions, like Kentucky. It also provides an excellent example of the type of research and level of investigative detail that can be achieved through interdisciplinary collaboration, as the combined work of engineers, construction crews, geologists, karst scientists, and the Museum staff, among others, provided an open and transparent forum for devising a safe, effective strategy for the sinkhole’s investigation and remediation.

References


LEPT, A SIMPLIFIED APPROACH FOR ASSESSING KARST VULNERABILITY IN REGIONS BY SPARSE DATA: A CASE IN KERMANSHAH PROVINCE, IRAN

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Abstract
There are a variety of widely used methods for porous aquifer protection to assess the vulnerability of groundwater resources, such as DRASTIC; Depth to water, net Recharge, Aquifer media, Soil media, Topography, Impact of vadose zone, and hydraulic Conductivity, SINTACS; depth to ground water (S), effective infiltration (I), unsaturated zone attenuation capacity (N), soil attenuation capacity (T), hydrogeologic aquifer characteristics (A), hydraulic conductivity range (C) and hydrological role of the topographic slope (S). And GOD; Groundwater occurrence, Overlying lithology, and Depth of groundwater. However, some more limited methods (including EPIK; Epikarst development, Protective cover, Infiltration conditions and Karst network development, PaPRIKa; Protection of karst Aquifers based on their Protection, Reservoir, Infiltration and Karstification type and COP; Concentration of flow, Overlying layers, and Precipitation regime) are also suggested for karstic aquifer vulnerability analysis. The latter methods are applied using different parameters such as karst network development, depth of karstification, and protective cover. Due to the nature of the data, these methods are highly affected by local and regional climate conditions. Data gathering for these methods is difficult, time consuming and needs a full understanding of karst systems. Data shortages, especially those related to karst formations in some parts of the world including the west part of Iran, and crucial demands for utilizing water resources demonstrate a great appeal to find a representative method for evaluation of these regions. Conventional methods of karst aquifer evaluation cannot be properly applied in the absence of a required karst data base; therefore, there is a need for a method that could be applied with the least amount of available data. The LEPT method introduced in this paper is a simple approach which provides rough evaluation of the general information gathered from karst areas of the west of Iran combined with field experiments. This method, which utilizes four parameters to assess the vulnerability of karst aquifers, was applied to the karst areas of Kermanshah (a province in the west of Iran) for the first time. Results of this approach categorize karst plains into four zones with very high, high, low and very low sensitivity in terms of their vulnerability to environmental impact. These classes are positively correlated with field information.

Introduction
Despite its undeniable role in drinking water supply for both rural and urban areas, karst aquifers are highly vulnerable to contamination. In some cases, presence of thin or no soil cover, shallow depth/thickness of karst aquifer overburden (epikarst zones) and direct point recharge via swallow holes make these water resources more susceptible to contamination by a variety of anthropogenic pollutants. On the other hand, because of the high groundwater velocity, short residence time of pollutants in karst aquifers affects the processes of contaminant attenuation in karst systems (Goldscheider, 2005). This is especially true in bare or thinly covered karst terrains. Comprehension of the level of sensitivity of karst aquifers to contamination and provision of a thorough karst management strategy can establish an effective framework for planning and scheduling protection programs.

Several researchers have shown keen interest in groundwater protection since Margat (1968) and Albinet
and Margat (1970) first introduced the concept of the vulnerability of groundwater to contamination (Foster and Hirata, 1988; Adams and Foster, 1992; Drew and Hotzl, 1999; Zwahlen, 2004). As a result of these investigations, some methods have been introduced for mapping karst aquifer vulnerability, including DRASTIC (Aller et al., 1987); GOD (Foster, 1987); AVI (Van Stempvoort et al., 1993) and SINTACS (Civita, 1994).

Although there are specifically designed methods to evaluate the vulnerability of karst systems, in some cases, these methods have been modified due to highly heterogeneous and anisotropic characteristics of karst aquifers. These modified methods could not be conveniently and broadly used in every karst system because of their vast input data requirement. The absence of climatic and/or hydrogeological data as well as difficult and expensive ways of data gathering in these fields stimulate a demand for establishing new methods that could be applied with the least available data and still lead to acceptable interpretations.

Several methods were specifically developed for the assessment of vulnerability in karstic areas. These include: COST action 620 or the European approach to vulnerability and risk mapping for the protection of karst aquifers (Zwahlen, 2004; European Commission, 2000; Daly et al., 2002; Goldscheider and Popescu, 2004); EPIK (Doerfliger and Zwahlen, 1998; Doerfliger et al., 1999), PI; Protective function of the layers above the saturated zone and the Infiltration conditions (Goldscheider et al., 2000), COP (Vias et al., 2006); SINTACS PRO KARST is an adaption of the original SINTACS in which the score values for each parameters are changed on the basis of different types of karst found in each specific area (Cucchi et al., 2004); RISKE; Rock, Infiltration, Soil, Karst and Epikarst (Petelet-Giraud et al., 2000); RISKE2 (Plagnes et al., 2005); KARSTIC (Davis et al., 2002); REKS; Rocks, Epikarst, Karstification and Soil cover (Malik and Svasta, 1998); PaPRIKa (Kavouri et al., 2011); COP + K; Concentration of flow, Overlying layers, Precipitation regime and Karst saturated zone. (Vias et al., 2006; Andreo et al., 2009); The Slovene Approach (Ravbar, 2007; Ravbar and Goldscheider, 2007); and, integrative vulnerability assessment in karst areas (Butscher and Huggenberger, 2009).

Residents of the western territories of Iran, which are mainly covered with carbonate rocks, are utilizing karst aquifers to supply their rural and urban water demands. In this region, karst water is a vital source for drinking, agricultural and industrial usage. However, to date, no practical strategies have been established for either karst water protection or vulnerability mapping.

Kermanshah, a province in western Iran, in which 35% of the area is underlain by carbonate formations, is a remarkable example of an area that is strongly dependent on karst water resources. Despite the fundamental impact of these resources on socio-economic and cultural development of the region, no systematic evaluation has been carried out to classify and manage the karstic formations throughout the region. Some research has been conducted on hydrogeology of the aquifers within a number of academic masters theses, doctoral dissertations, and local reports; however, there are still no reliable base maps of karst hydrogeology. Lack of proper data on karst formations, geomorphology, epikarst thickness, karst network development and other infrastructural information of this type has made it more difficult to establish an inclusive pattern for water budget estimation in this region.

This study attempts to introduce a new method for vulnerability mapping of karst aquifers using the limited data available for Kermanshah. Table 1 presents the main

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parameters used in several groundwater vulnerability mapping methods. In the case of Kermanshah’s karst area, data for the ordinary vulnerability mapping are either unavailable or unreliable (with low resolution).

**Study Area**

Kermanshah is located in the west of Iran with an area of 25,009 Km$^2$ and a population of more than 2 million. This region is greatly dependent on karst water resources for drinking and other major demands, such as agriculture, industry and ecotourism. The study area is located in the Zagros zone. This zone is divided into three sub-zones including High Zagros (HZ), Folded Zagros (FZ) and Sanandaj-Sirjan (SS). The major parts of the karst covered areas in the south-western, eastern and south-eastern parts of Kermanshah, are laid in the FZ zone while western and north-western part of the region falls into the HZ zone. The other part of the study area, with no remarkable karst development, is located in the SS zone. Three types of aquifers have been detected in Kermanshah: alluvial or porous, karst and hard rock. Hard rock aquifers have been poorly studied so far and there is a considerable lack of data for recharge estimation through the study area. Alluvial and karst aquifers have covered 3,613 and 6,575 km$^2$, respectively.

The main source of drinking water is pumping from deep-water wells in karst/porous aquifers or from karst springs. Karst aquifers are Mesozoic to Oligo-Miocene aged carbonate rocks such as Ilam, Sarvak, Asmari, Talezang and Bisetoun formations (Figure 1). Regardless of the geologic formations and their ages, karst terrains of the region could be classified into two main categories (based on their surficial soil cover thickness): buried karst and bare karst. In the buried karst, the carbonate/evaporate formations are underlain by a reasonable thickness (10-150m) of Quaternary sediments while, in the bare karst, there is no soil cover overlying the karst formations. From the karst protection point of view, buried karst are significantly more protected against

![Figure 1. Simplified geological map of Kermanshah; 1: alluvium, 2: volcanic rocks, 3: limestone, 4: dolomitic limestone, 5: marly limestone, 6: metamorphic limestone, 7: volcano-metamorphics, 8: marl, 9: sandstone, and 10: gypsum.](image-url)
environmental and anthropogenic pollutant sources than the bare karst. This is because of the ability of soil covers to remove/reduce the pollutants from the downward sinking waters. However, there are still some concerns related to overexploitation activities threatening these kinds of aquifers.

The southwestern part of Kermanshah is located in a semi-arid climatic zone, while the other areas (i.e., northwest, east and north) fall into the cold climate category. The mean annual precipitation of these regions is about 500 mm (up to 800 mm in higher altitudes). In the Parau and Shahu Mountains, there are several swallow holes, vertical shafts and sinkholes located in altitudes higher than 2,000 m above sea level (a.s.l.). Snow melting in the high karst plateau is the main source feeding the lowland springs. These springs mainly emerge at the contact areas between the karst formations and non-karst rocks. In some cases, these springs have a considerably high discharge, i.e., Bel spring in the north-western region of Kermanshah with a mean annual discharge of about 5m³/s. In this province, relatively impervious radiolarites underlaying the carbonate formations have impeded the rate of downward flow of water, forcing it to continue along the contact surface between the two rock units. This process consequently creates a large number of contact springs throughout the region. These springs, which sometimes have large sizes and discharge rates, are called Saraw or Sarab in Kurdo-Persian and local dialect.

The results of a limited number of dye tracing methods showed that there is most often a hydraulic connectivity between highland karsts with lowland discharge points. On the other hand, annual precipitation has a significant effect on the discharge of large springs like Ravansar spring. The quick response of these karstic springs to the precipitation changes can sometimes increase the muddiness of their output flows. Accordingly, conduit-diffuse flow systems are present in mid to large size springs, while the smaller ones often follow a diffuse flow regime. Despite a great number of manmade dams of different sizes throughout the province, karst springs are still serving as a major source of drinking water, especially in rural areas. The main concern about these valuable yet vulnerable sources of water in Kermanshah is the absence of comprehensive studies on karst vulnerability mapping and protection programs.

Methodology

LEPT Method

The acronym LEPT stands for a methodology based on four parameters that could be utilized for karst vulnerability mapping in regions with sparse data. The LEPT method is comprised of four initial data layers: Lithology (karstic rocks); Elevation (sinkhole distribution based on the high karst plateau elevation); Protective cover; and, Topographical slope maps. LEPT as well as EPIK (Doerfliger and Zwahlen, 1998) and DRASTIC (Aller et al., 1987) are all multi-attribute weight-rating approaches (overlay and index method). Sinkholes distribution is mainly controlled by elevation and lithology. The protective cover and topographical slope affect groundwater movement into and through karst aquifers. Primary factors of the LEPT model are elevation classification related to sinkhole distribution and the intensity of karstification in carbonate formations. The final karst vulnerability map is computed using the Equation 1;

\[ V = 4L + 3E + 2P + T \]  

(Eq. 1)

Where V: karst vulnerability; L: Lithology; E: Elevation; P: Protective cover; and T: Topographical slope.

Lithology

The digitized lithology of the 1:250,000 geology map of Kermanshah (Braud, 1978) was used to provide one of the layers for the model. By using this map, Kermanshah’s karst formations were divided into three classes based on their intensity of karstification, density of karst springs (number/Km²) and limestone purity. In this classification scheme, the highly developed karst area receives the highest value of 3, while medium and low/non karst developed formations are assigned the values of 2 and 1, respectively (Figure 2). The higher the value, the higher the vulnerability to karst development would be and vice versa. Mean microscopic porosity of the carbonate formations (obtained from thin section analysis), results of in-situ permeability tests, karst spring density and discharge rate and log observations from drilled boreholes through karstic formations are the main lithological sub-factors utilized in the LEPT evaluation method. The subcategories of these parameters are summarized in Table 2.
Therefore, for further evaluations, geological and soil distribution maps were utilized. Considering the data extracted from the geological maps of the study area, the thickness of the soil cover (wherever there is a soil cover) is always greater than 1 m. This, then, gives all the soil-covered areas a value of 2 in the final classification processes.

**Table 2. Attribute classes for the lithology (L), elevation (E), protective cover (P) and topographical slope (T).**

<table>
<thead>
<tr>
<th>Acronym /Characterization</th>
<th>Weight</th>
<th>Relative weight</th>
</tr>
</thead>
<tbody>
<tr>
<td>L1 Pure limestones and dolomite</td>
<td>3</td>
<td>4</td>
</tr>
<tr>
<td>L2 Marly limestone, gypsum</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>L3 Sandstone, Marl, Crystalized metamorphic rocks, volcanic and Quaternary old deposits</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>E1 High Sinkhole density in karst plateau</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>E2 Sinkhole and dry caves are present but low density</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>E3 Sinkhole is very area or absent</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>P1 Karst lands without protective cover or present a thin layer of soils</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>P2 Lands which covered by thick layer of soil, alluvium and scree.</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>T1 Gentle dip or flat lands</td>
<td>3</td>
<td>1</td>
</tr>
<tr>
<td>T2 Dip between 10 degree to 30 degree.</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>T3 Highly slope lands with some karst and fractures</td>
<td>1</td>
<td></td>
</tr>
</tbody>
</table>

**Protective Cover**

The classification for the protective cover is simpler than the other factors. In this category there are just two main classes: class 1, in which no protective soil cover overlies the carbonate formations (receives a value of 2) and class 2, with thin to moderate thickness of sedimentary layers covering the carbonate formations (receives a value of 1) (Figure 4).

It should be noted that, in susceptibility zoning of the karst areas pursued in this study, epikarst and protective (soil) cover layers were considered as a single category.

**Elevation (Sinkhole Distribution Based on High Plateau Karst Elevation)**

Sinkholes are amongst the most well-known features of karst terrains. These landforms have a great variety of sizes and distribution patterns in the high altitudes of Zagors Mountains. The swallow holes, sinkholes and vertical shafts are main paths of flow concentration within the karst systems. Based on the study conducted by Ghorbani and Mahmoudi (2010), the snow lines of the Kermanshah Mountains have been uplifted from 1,800 m (a.s.l.) in the Quaternary to 2,500-3,500 m (a.s.l.) at the present time. As a result of this study, Kermanshah karst lands were divided into three classes based on their elevation. Higher elevations have more potential for karst development. Therefore, karst lands with elevations higher than 3,000 m fall into class 1 (with a value of 3), those with elevations between 2,000 to 3,000 m fall into class 2 with a score of 2 and the others with elevations lower than 2,000 m are categorized as class 3, with a score of 1 (Table 2 and Figure 3).
map was then subdivided into three categories in accordance with the degree of vulnerability of each factor/layer based on Natural break criterion in the GIS environment. The natural break classification method has commonly been used in landslide susceptibility mapping to categorize the susceptibility classes (Falaschi et al., 2009; Bednarik et al., 2010; Pourghasemi et al., 2013) and sinkhole susceptibility mapping (Taheri et al., 2015).

**Sensitivity Analysis**

A “Map removal” and “Single parameter” sensitivity analyses are two common sensitivity tests for some of the parametric methodologies such as DRASTIC, EPIK, PaPRIKA and so forth. The Map removal sensitivity was performed by Lodwick et al. (1990) and the single parameter was introduced by Napolitano and Fabbri (1996). Sensitivity of removing one or more maps can be expressed as (Lodwik et al., 1990; Gogu and Dassargues 2000):

$$S = 100 \left( \frac{V}{N} - \frac{v_x}{n} \right)/V$$  
(Eq. 2)

Where $S$ is the sensitivity associated with the removal of one map, $V$ and $v_x$ are the vulnerability degrees computed by using Eq. 1 without or with considering the parameter $X$, respectively; $N$ and $n$ are the number of data layers used to calculate $V$ and $v$.

The single parameter sensitivity test was performed to assess the influence of each of the four parameters of the model on the vulnerability measure. With this approach, the real or effective weight of each parameter could be compared with its allocated or theoretical weight.
(Napolitano and Fabbri, 1996). The real or the effective weight is calculated as follows:

\[ W = 100 \times \frac{Pr \times Pw}{V} \]  
(Eq. 3)

Where \( W \) refers to the “effective” weight of each parameter, \( Pr \) and \( Pw \) are the value and weight for each parameter, and \( V \) is the overall vulnerability index calculated using Eq. 1.

**Result and Discussions**

The karst vulnerability map of Kermanshah (Figure 6) was made by overlaying the four parameters of the LEPT method through raster analysis in GIS. The results show an area of very high vulnerability that covers 2,094 km\(^2\) of Kermanshah (8.4 % of the study area). This area is distinguished by its remarkable sinkholes, shafts, caves and other active karst landscapes. Snow melting during spring is the main source for large karst springs feeding into lowlands. Due to the preferential drainage in sinkholes and other open karst landscapes in it, this zone is very sensitive to contamination. Parau, a famous cave in the region, and Shahu, a karst plateau, are both located in this zone.

High and very high vulnerable areas cover 6,400 km\(^2\) of the study area, equal to 25.6 % of Kermanshah’s total area. The dry caves and karren fields which were developed during the Quaternary by fluvial karstification are some of the main features of these areas. Many caves and large springs are located in this zone.

A Low vulnerable area comprises 6,540 km\(^2\) or 26% of the study area. This zone is characterized by karstified formations developed at the contact of a non-karstified area and Quaternary deposits.

A Very low or none vulnerable area covers approximately 40 % of the entire province, with an area of 9,974 km\(^2\). The results of the map removal sensitivity analysis are shown in Table 3.

Results show that the relative influence on the final LEPT map is \( E > L > P > T \). On the final map, the statistical

![Figure 6. Final map of Kermanshah karst vulnerability.](image-url)
On the other hand, the real weights of parameters P and T with respectively 23.52% and 10.92% are greater than their corresponding theoretical weights of 20% and 10%. Therefore, it is important to have these data compared in order to produce a reliable final output map.

References


<table>
<thead>
<tr>
<th>P</th>
<th>Av. (%)</th>
<th>Std(%)</th>
<th>Med (%)</th>
<th>Min (%)</th>
<th>Max (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>L</td>
<td>4.54</td>
<td>4.33</td>
<td>9</td>
<td>0</td>
<td>16</td>
</tr>
<tr>
<td>E</td>
<td>22.4</td>
<td>1.93</td>
<td>22</td>
<td>18</td>
<td>25</td>
</tr>
<tr>
<td>P</td>
<td>1.6</td>
<td>2.99</td>
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<tr>
<td>T</td>
<td>4.98</td>
<td>3.74</td>
<td>6</td>
<td>0</td>
<td>10</td>
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Table 3. Statistics of map removal sensitivity analysis.

<table>
<thead>
<tr>
<th>P</th>
<th>TW (%)</th>
<th>TW (Av. %):Theoretical Weight in percentage (%)</th>
<th>Std(%)</th>
<th>Med ( %):standard deviation</th>
<th>Min: Minimum (%)</th>
<th>Max: Maximum (%)</th>
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<tbody>
<tr>
<td>L</td>
<td>40</td>
<td>41.23</td>
<td>8.56</td>
<td>41</td>
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<tr>
<td>E</td>
<td>30</td>
<td>23.44</td>
<td>6.39</td>
<td>30</td>
<td>12</td>
<td>60</td>
</tr>
<tr>
<td>P</td>
<td>20</td>
<td>21.52</td>
<td>5.31</td>
<td>18</td>
<td>9</td>
<td>33</td>
</tr>
<tr>
<td>T</td>
<td>10</td>
<td>16.95</td>
<td>3.84</td>
<td>13</td>
<td>3</td>
<td>30</td>
</tr>
</tbody>
</table>

Table 4. Statistics of single parameter sensitivity analysis.

parameters (Table 3) show that elevation (E) is the parameter with the highest sensitivity. Accordingly, the L parameter has the second highest value due to its high rating and weighting. The LEPT method is also sensible to remove the (P) parameter because this presents a vast spatial distribution. The (T) parameter is similar to the P parameter. These results are logically acceptable because, based on the distribution of sinkholes on the high karst plateau, the parameter E is playing an important role in dispersing the contaminants.

The single parameter sensitivity analysis indicates that the lithology parameter (L) dominates the vulnerability index with an average weight of 41.23 % versus the theoretical weight of 40 %. Due to its very influential effect on final output of the model, the high sensitivity of the L parameter was expected (Table 4). This dominance has been clearly seen throughout the study area. On the other hand, the real weight of parameter E (23.44 %) is notably smaller than its theoretical weight (30 %). It means that the actual influence of this parameter within the study area is lower than what was estimated based on Eq. 1.


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Numerical Simulation of Karst Soil Cave Evolution

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Abstract
This study is focused on numerical simulation of the formation and development of karst soil caves related to cover-collapse sinkholes. The so-called ‘karst soil cave’ refers to the caves formed in the soil layers above bedrock of sinkhole regions. Because the soil caves are formed and developed under groundwater seepage, studying groundwater level changes can help understand soil cave development and collapse. Based on the improved Terzaghi loosening pressure theory and using excess pore water pressure, two kinds of critical groundwater level decline are discussed. The first, denoted as $\Delta H_0$, is the critical groundwater level decline related to soil cave formation and evolution; the other one, denoted as $\Delta H_T$, is the critical groundwater decline related to cave roof collapse. After a soil cave is formed, its evolution can cause uneven displacement and stress redistribution in the overlying soil layer. The process of soil cave expansion can be understood by investigating the change in displacement and stress. Numerical simulation of the vertical displacement using FLAC3D shows that the maximum vertical displacement occurs at the arch roof of the soil cave and that the displacement can cause tensile failure of the arch roof. The simulated soil layer displacement is used to determine the soil depth disturbed by the cave by delineating the planes of equal settlement. Analyzing the simulated shear stress shows that the maximum shear stress occurs at the arch toes and causes shear failure. On the other hand, the zones of low shear stress can be used to evaluate existence of arching effect in the overlying soil layer. By analyzing the plastic zone of the soil layer, it was found that, in rigid clay, arch roof collapse and tensile failure are the major events that lead eventually to the barrel-shaped or bottle-shaped forms of collapsed pits. In loose soil, shear failure of the arch toe is the major event that eventually leads to the taper-shaped or bowl-shaped form of collapsed pits. Generally speaking, stability and size of soil caves can be determined using the three variables of low shear stress area, equal settlement plane, and plastic zone discussed above. The numerical simulation of this study is valuable to the monitoring and assessment of sinkhole occurrence.

Introduction
A cover-collapse sinkhole, a roughly circular hole in the ground that is variable in depth, is one of the most adverse geological phenomena in an area underlain by limestone bedrock. The collapse feature is a result of the movement of soil or other related materials carried by water down into voids either in the limestone bedrock or within the soil profile. This study aims to understand the formation and development of voids in the soil profile (the karst soil cave) related to the occurrence of cover-collapse sinkholes, which has been a long-lasting challenge for various reasons, including the hidden development of the soil caves and complexity in karst hydrogeology and geological conditions. The main approaches of studying sinkhole soil caves are as follows:
Geophysical Methods

Ground-penetrating radar (GPR) is a relatively popular geophysical method for investigating sinkholes. The GPR relies on the reflection of a high frequency (25-1,000 MHz) electromagnetic (EM) pulse from layer contacts and other “anomalies” (boulders, cavities, utilities, tanks, etc.). GPR has become an important non-invasive technique for rapid geophysical mapping of the shallow subsurface. (Benson, 1987; Witten and Calvert, 1999; Martin-Crespo and Gomez-Ortiz, 2007). However, because of the signal attenuation, depth of penetration is limited. For example, subsurface clay always attenuates GPR signals. This occurs often because most cover-collapse sinkholes are related to clayey overburden on cavity-laden limestone bedrock. Other geophysical methods are available, such as Direct current (DC) resistivity techniques (Panno et al., 1994; Batayneh and Al-Zoubi, 2000), micro-gravity (Butler, 1984) and electromagnetic method (Kaspar and Pecen, 1975). Although these techniques have been widely used in karst regions, they cannot be used for effective monitoring and early warning of sinkhole collapse.

Remote Sensing Methods

Remote sensing methods, such as InSAR (Interferometric Synthetic Aperture Radar), provide satellite images of the Earth’s surface that can be combined to show subtle movements, i.e., deformation, of the ground surface. The ground deformation can be used to monitor and provide early warning of sinkhole collapse. However, remote sensing methods are unsuitable to real-time monitoring (Paine et al., 2009). Similarly, LiDAR (Light Detection and Ranging) technology is unsuitable to real-time monitoring, although it provides high density data which can be processed to produce a high resolution topographic map (Doctor and Young, 2013; Shaw-Faulkner et al., 2013).

Photoelectric Monitoring Methods

These methods include optical fiber sensing (BOTDR and OTDR) and coaxial cable sensing (TDR). By burying fiber or coaxial cable in soil, soil deformation and damage creates strain on the cables, and the strain can be measured by special instruments. This makes it possible for real-time monitoring. Photoelectric sensors have played the role of precise monitoring and early warning for small-scale engineering sites. However, photoelectric sensors have not been applied to a large-scale site (Dowding 2003, Guan et al. 2013).

Trigger Factor Method

In China, groundwater level decline is the major triggering factor of sinkhole collapse, and the decline plays a key role in soil cave formation and evolution (Figure 1). Therefore, in areas prone to sinkhole collapse, monitoring changes in groundwater levels can be used for early warning of sinkhole collapse. When water level change exceeds a critical value, sinkhole formation is expected. To implement this requires determining the critical conditions of karst collapse (Lei 2010, Yan 2014), and this is one of the motivations for this study.

Another motivation for this study is to understand sinkhole evolution caused by uneven displacement and stress redistribution of the soil overlying the sinkhole. The displacement and stress variation is simulated numerically in this study using FLAC3D, a three-dimensional finite difference program. Figure 2 is an example computational grid of FLAC3D used to simulate sinkhole evolution.
Soil Cave Development Process and Numerical Model

The soil cave formation and development is closely related to the properties and structures of soil, water, and rock. Figure 1 illustrates the process of sinkhole formation and development. First, the soil cave formation requires a karst cave opening upward and having sufficient space to accommodate falling soil. Under gravity and groundwater seepage forces, the soil layer breaks and enters the karst cave, and this is the formation of soil cave. Afterward, the soil cave can quickly expand upward. When shear stress is less than cohesive strength of the soil, the soil cave stops developing and remains in a temporarily stable state. When the dynamic conditions change, the soil sinkhole will continue to expand until the collapse of the soil cave roof, as shown in Figure 1.

Groundwater, as the most active factor of sinkhole collapse, plays a key role in the formation and evolution process of karst soil caves. Based on soil sinkhole experiments, Jiang (1998) proposed that, when the change of karst cavity pressure reaches a certain value, soil failure occurs. Since karst cavity pressure and karst aquifer water pressure are subject to the same changes, formation of soil cave and collapse can be studied by investigating karst aquifer water level changes. Based on the relations of excessive water pressure, \( \sigma_w = \rho_w g \Delta H \), and \( \sigma_z \geq \sigma_t = c/3 \), Wan (2003) developed the critical water level decline for soil cave formation and continuous development

\[
\Delta H_0 = \frac{c}{3g \rho_w}
\]

where \( c \) is soil cohesion, \( g \) is the gravitational acceleration, \( \rho_w \) is the water density, and \( \sigma_t \) is soil tensile strength.

For the collapse of a soil cave, when the summation of excessive pore water pressure and vertical pressure is positive, i.e., \( \sigma_w + \sigma_z > 0 \), the soil cave roof collapses, and the critical water level drop for cave roof collapse is

\[
\Delta H_T = -\frac{\sigma_z}{\rho_w g}
\]

The soil sinkhole vertical pressure \( \sigma_z \) without overlying load can be derived using the Terzaghi loose pressure theory improved by Kezdi (1975) for a circular cavern. It is expressed as

\[
\sigma_z = \frac{\gamma H}{4K \tan \phi} \left( 1 - e^{-4K \tan \phi \frac{D}{H}} \right)
\]

where \( D \) is hole diameter, \( K \) is the lateral earth pressure coefficient, \( Z \) is the depth of soil and the roof panel, and \( \phi \) is the angle of internal friction. Equation (3) shows that, when the soil hole diameter is small, the \( \sigma_z \) is negative. Under this condition, there exists a collapse resistance, and the soil cave is temporarily stable.

Analysis of the above equations reveals the following:

1. When the overlying layer of the soil cave is sand and the cohesion \( c \) is very small, \( \Delta H_0 \) is small. As a result, small fluctuation of groundwater level may cause formation of soil cave. In addition, \( \sigma_z \) in a small soil cave is already positive, and it leads to emergence of roof collapse due to raveling of soil into the void space in the underlying bedrock.

2. When the overlying layer is clay and cohesion \( c \) is big, soil cave can be in the temporarily stable state. When groundwater level declines more than \( \Delta H_0 \), soil cave starts to grow upward to the ground surface until a new stable state is formed or collapse occurs.

3. When groundwater level dropped more than \( \Delta H_T \), soil cave roof is unstable, and collapse occurs. When the soil hole diameter is small, \( \Delta H_T > \Delta H_0 \) when the soil hole become larger, \( \Delta H_T \) becomes smaller.

For all the cases of sinkhole formation and collapse, the overlying layer is subject to displacement and stress changes. It is therefore necessary to understand the changes. FLAC3D is used in this study to simulate soil displacement and stress in the stable stage of sinkhole development. The FLAC3D-based model uses a three-dimensional domain with the dimension of \( a \times b \times c \), and the dimension can be adjusted to meet specific project needs. As shown in Figure 2, the model grid consists of tetrahedral blocks. The bottom boundary (the soil rock interface) of the domain is set as a fixed constraint boundary, and the other boundaries are set as the one-way boundary. The soil cave roof is always an arch, and generalized in this study as a hemisphere with the height of \( H \) and the roof span of \( D \). The size parameters can be adjusted for different projects by using the FISH language of FLAC3D programming. The FLAC3D modeling of displacement and stress uses the Mohr-Coulomb failure criterion, and the associated parameters are listed in Table 1.
The simulation procedure is as follows:

1. Conduct initial stress analysis for the entire modeling domain before the soil cave is formed.

2. Generate the cave (with the height of H and span of D) using the nill command of the FLAC3D FISH language.

3. Conduct stress analysis to simulate stress and strain of the new model domain with the cave.

4. Use the nill command to expand the cave size and simulate the sinkhole evolution and development.

### Analysis of Vertical Displacement

In order to analyze the displacement changes during the soil cave expansion and evolution, the vertical displacement is simulated using FLAC3D and the contour lines are shown in Figure 3. The simulation results indicate that the soil settlement is uneven due to the emergence of the soil cave; the maximum vertical displacement always occurs in vault. The vertical displacement can cause tensile failure, extending from the vault to both sides. Correspondingly, ground subsidence also occurs, and the largest subsidence generally appears above the vault.

The vertical displacement of the overlying layer can be used to determine the position of the “equal settlement plane”. The plane is defined as the position where uneven settlement begins. In China, the value of allowable settlement is 30mm, and is used in the community of geological engineering. This value is used to determine the “equal settlement plane”, i.e., lines m-n marked in Figure 3. With the soil cave expanding upward, the

---

### Table 1. Model parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>density $\text{kg/m}^3$</th>
<th>Volume modulus $\text{Pa}$</th>
<th>Shear modulus $\text{Pa}$</th>
<th>$\phi$ $\circ$</th>
<th>$c$ $\text{Pa}$</th>
<th>Tensile strength $\text{Pa}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>clay bed</td>
<td>1.620</td>
<td>3.3xE7</td>
<td>7.1xE6</td>
<td>24</td>
<td>2.5xE4</td>
<td>6xE3</td>
</tr>
<tr>
<td>gravel bed</td>
<td>1.790</td>
<td>4.6xE7</td>
<td>4.7xE5</td>
<td>25</td>
<td>1.1xE4</td>
<td>3xE3</td>
</tr>
</tbody>
</table>

---

**Figure 3.** Diagram of FLAC3D-simulated vertical displacement with the cave dimension of (a) 5m, (b) 5m, and (c) 10m. D is the cave span. The overlying soil is clay. The lines of “m-n” denote the “equal settlement plane”.
position of “equal settlement plane” rises accordingly. When the plane reaches the ground, the whole overlying layer subsides, and the soil cave is in the critically unstable stage for sinkhole collapse to occur.

Analysis of Shear Stress
Using the FLAC3D FISH language, the contours of maximum shear stress are plotted in Figure 4. The maximum shear stress is half of the difference of the maximum principal stress and the minimum principal stress, i.e., \( \tau_{\text{max}} = \sigma_1 - \sigma_3 / 2 \). Figure 4 shows the phenomenon of stress concentration at the arch toes. If the maximum shear stress near the toes reaches the soil strength, then the soil cave is subject to shear failure, and the damage spreads from the toes to the arch vault.

During the simulation process, it is found that a “low value zone” of shear stress appears at the top of the soil cave, where the shear stress contours are depressed downward on top of arch (Figure 4a, b). This phenomenon can be explained by the soil arch effect (Terzaghi, 1943; Handy, 1985). Due to the uneven displacement of the overlying layer, the stress of the entire domain is redistributed, and the stress applied to the arch is spread out to the toes and the surrounding media.

As a result, the maximum principal stress, \( \sigma_1 \), decreases, and the minimum principal stress, \( \sigma_3 \), increases, causing diminution of the maximum shear stress, \( \tau_{\text{max}} \). Therefore, the shear stress “low value zone” can be used to determine the stabilization of the soil cave. The “low value zone” also evolves with the cave development. As shown in Figure 4, as the soil cave extends upward, the “low value zone” becomes gradually smoother (Figures 4a and 4b), and eventually disappears when the soil cave develops to a certain extent (Figure 4c). At this moment, the soil arch effect disappears, and the soil cave is in the unstable state and sinkhole collapse can occur.

Figure 4. Distribution of FLAC3D-simulated shear stress with the cave dimension of (a) 5m, (b) 5m, and (c) 10m. D is the cave span. The overlying soil is clay.
Simulation of Plastic Zone

FLAC3D can be used to simulate the plastic zone for examining the area of potential failure regions around the soil cave. In the rigid clay layer, the plastic zone is mainly above the arch of the soil cave, and becomes bigger with the cave development (Figure 5, subplots a1, b1 and c1). Under this circumstance, the overlying soil layer is subject to tensile failure, and the vault collapses to form a barrel-shaped or bottle-shaped pit. At the same time, due to stress release and the existence of the free face, the soil near the upper arch toes inclines and slides to the pit center. Concentric tension cracks also form on the ground toward the collapse pit.

In the loose sand layer, the plastic zone appears near the arch toes, and develops along the tangential direction of the arch as the soil cave further develops (subplots a2, b2, and c2 of Figure 5). Therefore, the cave collapse starts from the shear failure at the arch toes. Under this circumstance, the entire overlying layer is broken, and taper-shaped or plate-shaped collapse pit is formed. In the early stage of this process, the plastic zone is formed only in the periphery area of the soil cave. With the development of the soil cave, the plastic zone occurs throughout the entire overlying layer. When the layer falls within the plastic zone, ground collapse occurs. Therefore, the stage of soil cave formation can be determined by examining the changes of the plastic zone.

Figure 5. Distribution of FLAC3D-simulated plastic zones with the cave dimension of (a1-a2) 5m, (b1-b2) 5m, and (c1-c2) 10m. D is the cave span. The plastic zone is in the violet area.
Summary
The FLAC3D-based numerical simulation leads to the following conclusions:

1. Soil caves develop gradually, and there exist periods when the sinkholes are stable. The stable periods are altered by groundwater flow, and the soil caves continue developing under the impacts of groundwater seepage force.

2. Based on the improved Terzaghi loosening pressure theory and using excess pore water pressure, two kinds of critical groundwater level decline are discussed. The first, denoted as $\Delta H_0$, is the critical groundwater level decline related to soil cave formation and evolution; the other one, denoted as $\Delta H_T$, is the critical groundwater decline related to soil cave roof collapse.

3. Based on the FLAC3D simulation results, three variables are defined, and they are “low shear stress zone”, “equal settlement plane” and “the plastic zone”. They can be used together to evaluate stability of the soil caves.

Acknowledgments
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EXPERIMENTAL AND NUMERICAL INVESTIGATION OF SINKHOLE DEVELOPMENT AND COLLAPSE IN CENTRAL FLORIDA

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Abstract

The mechanisms of sinkhole formation, development, and collapse are investigated in this study using experimental and numerical methods. Sandbox experiments are conducted to understand how excessive groundwater pumping triggers sinkholes formation. The experimental results indicate that the change of hydrologic conditions is critical to sinkhole development. When seepage force increases due to increase of hydraulic gradient, clay and sand particles start moving downward to form a cavity. The confining unit is of particular importance because the cavity is first formed in this layer. Based on the conceptual model developed from the sandbox experiments, the Fast Lagrangian Analysis of Continua (FLAC) code and Particle Flow Code (PFC) are coupled to simulate the sandbox experiments. PFC was used to simulate particle movement in the sinkhole area, and FLAC is used for other areas. While the current numerical simulation can simulate the experiment results such as the sizes of the cavity and the sinkhole, the simulation capability is limited by the computing cost of PFC. More effort of model development is necessary in the future study.

Introduction

Sinkholes are a common geological feature of karst landscape in Florida, southeastern United States, and worldwide. In particular, cover-collapse sinkholes occur abruptly and can cause catastrophic damages such as death, injury, and property damage. In Florida, a Tampa resident vanished into a sinkhole that opened under his bedroom on a night in March, 2013. In the last several years, sinkholes have become Florida’s insurance disaster due to sinkhole collapse in urban areas. Cover-collapse sinkholes also do severely damage buildings, drain farm ponds, damage roads, and wreck farming equipment, and lead to engineering and environmental problems (Beck, 1988). There is an urgent need to understand the mechanisms of sinkhole development and catastrophic collapse.

Cover-collapse sinkholes occur in the soil or other loose material overlying soluble bedrock. The thickness and cohesiveness of the soil cover determine the size of a cover-collapse sinkhole. Figure 1 shows a typical process of cover-collapse sinkholes formation caused by excessive groundwater pumping. A karst aquifer is the
pre-requisite, and sinkhole development always starts from dissolution of soluble rocks or fractures and conduits to create an opening, which provides a passage for soil transport downward. Groundwater is one of the primary triggering mechanisms for sinkhole development and collapse, because seepage force due to groundwater flow drags clay and sand particles downward. Figure 1(A) shows two layers with cohesive and non-cohesive soil; the cohesive soil layer overlies the karst bedrocks and is beneath the non-cohesive soil layer, which is common in central Florida. In the initial stage of sinkhole development, an opening forms in the rock at the interface between the bed rock and cohesive soil. While the piezometric surface is higher than the water table in the initial stage, after excessive pumping starts, the piezometric surface decreases (Figure 1(B)) and downward groundwater flow is created. Subsequently, sediment in the cohesive soil layer starts falling down due to gravitational force and seepage force applied on the particles (Figure 1(B)). This forms a cavity in the cohesive layer. The cavity gradually increases, and it in turn increases hydraulic gradient and thus seepage force. Because of these, the cavity expansion accelerates (Figure 1(C)). Once the cavity expands to the non-cohesive soil layer, sand movement becomes dramatic. When the non-cohesive soil layer cannot support the overlying material, collapse occurs and a sinkhole propagates to land surface (Figure 1(D)).

While the process of cover-collapse sinkhole formation triggered by groundwater pumping has been understood, mathematical models and numerical modeling tools have not been available for predictive understanding. A coupled model based on FLAC and PFC was used to simulate the soil-structure interactions during a sinkhole event (Caudron et al., 2006). Ahmed (2013) used finite element analyses to detect three-dimensional (3-D) deformations due to submerged cavities that lead to sinkhole. Tharp (2003) employed an elastic-plastic model to demonstrate the development of a sinkhole above a karst cavity. Shalev (2012) adopted a two-dimensional (2-D) visco-elastic model to simulate the sinkhole formation to take into account of the brittle and ductile aspects of sinkhole collapse. Baryakh et al. (2009) established a numerical model that uses the discrete element method to simulate the evolution of the stress-strain state of a rock mass containing a karst cavity. Baryakh and Fedoseev (2011) also set up a finite element model of a growing cavern to describe possible scenarios of sinkholes development in the karstic areas, to determine formation criteria for ground surface sinkholes and underground caverns, and to estimate sinkhole and cavern sizes. Shalev et al. (2006) simulated the dissolution of salt layer and the creation of cavities using the finite element methods. The numerical simulation showed the growth of cavities from the bottom to the top of the salt layer, and suggested that sinkhole collapses shortly after the cavities reach the top of the salt layer. These modeling effort suggests that, while the continuum theory can estimate the stress-strain state of sinkhole events, it is difficult to take into account the change of cavity geometry such as enlargement of the cavity in the cohesive soil layer. While discontinuum theories can be used to resolve this problem, the dis-continuum theories are computationally intensive and not always practically affordable.

**Sandbox Experiment**

Sandbox experiments are conducted to better understand the process of sinkhole development and collapse. Figure 2 shows the schematic view of experiments. A sandbox of 150 cm × 120 cm × 20 cm was constructed with plastic material. There are four tanks to control the water level of unconfined and unconfined aquifer. As shown in Figure 3, the sandbox was filled with three different hydrogeological materials in three layers. The bottom one (in black) represented a karst aquifer with void space. A clay layer (in yellow) overlaid the karst layer to represent a confining layer. Between the two layers, three opening were designed, but only the one of 1 cm in the middle was used in this study to create a sinkhole in the middle of the sandbox. Above the clay layer was a sand layer (in grey) to represent an unconfined aquifer.
Hydraulic head of the unconfined aquifer was controlled by the inner reservoirs at the both sides of the sandbox. The outer reservoirs were used to control hydraulic head of the confined aquifer.

The sandbox experiments are designed to understand impacts of groundwater pumping on sinkhole development and collapse. The impacts are believed to be the major reasons for the sinkhole events in the Dover/Plant City area during the winter of 2010, when more than 100 sinkhole collapses were triggered by excessive groundwater pumping for irrigation to prevent crops from being frozen. The sandbox experiments start by lowering hydraulic head in the confined layer to mimic a pumping scenario. The water level in the unconfined aquifer remains constant. After the drop of hydraulic head in the confined aquifer, a small amount of clay particles moves downward through the opening due to the seepage force caused by hydraulic gradient between the unconfined and confined layers. A cavity starts to form in the clay layer, and slowly expands upward. Once the cavity reaches the sand layer, sinkhole development is accelerated, and sinkhole collapse occurs shortly because of small cohesion of the wet sand. Figure 4 shows the picture after sinkhole collapse.

**FLAC/PFC coupling approach**

In this study, we use the continuum and dis-continuum theories together by coupling the finite difference code, FLAC, based on the continuum theories with the discrete element code, PFC, based on the dis-continuum theories (both FLAC and PFC are developed by the Itasca Consulting Group, Inc.). Since PFC is computationally demanding, it is only used for the small area of excessive displacement above the opening. FLAC is less computationally demanding, and thus used to simulate the larger area of small deformation away from the opening. Using the coupled FLAC/PFC approach minimizes the computational requirement for simulating the process of sinkhole development and collapse.

**Coupling FLAC and PFC**

The coupling of FLAC and PFC is realized by exchanging displacements, velocities, and forces at each modeling step. The data exchange is made possible by the I/O socket connection ability to pass data rapidly between

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**Figure 2. Schematic of sandbox experiments.**

**Figure 3. Photo of sandbox experiments.**

**Figure 4. Sinkhole collapse in a sandbox experiment.**
the two codes running on the same machine or on separate machines with a network connection. As shown in Figure 5, the data exchange is two-directional between FLAC and PFC. In each step of numerical simulation, the velocity at the interface between FLAC and PFC model domains (Figure 6) are first obtained from the FLAC run and then sent to PFC via the I/O socket. After receiving the data from FLAC, PFC starts to update the forces at the interface and then send the results of forces back to FLAC via the I/O socket. Afterward, the simulation moves to the next time step, and the iteration continues until the end of simulation time.

**Numerical simulation**

Table 1 lists the values of the parameters used for the numerical simulation. While the clay/sand particle movements change hydraulic conductivity, to simplify the numerical simulation, it is assumed that the deformation and the particle transport have negligible effect on the hydraulic conductivity and that hydraulic conductivity is constant during the process of sinkhole development. The groundwater flow of the entire domain is simulated using the Darcy’s Law and heat equation.

The FLAC/PFC simulation is set up as shown in the sketch map of Figure 6. Zero horizontal displacements are assumed at the side boundaries, and zero vertical displacements at the bottom boundary in the FLAC modeling area. For the PFC modeling area, the bottom boundaries are the two walls (Figure 6) with the distance of 1cm. At the initial time, the model is in the steady state with the hydraulic head of 0.45m (the datum is at the bottom of clay layer) for the confined aquifer and 0.4m for the unconfined aquifer (sand layer).

The simulation starts by dropping the piezometric surface 0.1m rapidly to create unsteady flow and particle movement. During each time step of the flow modeling, hydraulic pressure and pressure gradient is calculated and then passed to the FLAC-PFC-based mechanical

<table>
<thead>
<tr>
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<th>Sand</th>
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<tr>
<td>Density (Kg/m³)</td>
<td>2200</td>
<td>2600</td>
</tr>
<tr>
<td>Bulk modulus (Pa)</td>
<td>7.00E+05</td>
<td>1.30E+07</td>
</tr>
<tr>
<td>Shear modulus (Pa)</td>
<td>4.00E+05</td>
<td>8.00E+06</td>
</tr>
<tr>
<td>Cohesion (Pa)</td>
<td>8.00E+05</td>
<td>0</td>
</tr>
<tr>
<td>Friction angle (°)</td>
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<td>35</td>
</tr>
<tr>
<td>Hydraulic conductivity (m/s)</td>
<td>1.00E-08</td>
<td>5.00E-05</td>
</tr>
</tbody>
</table>

**Figure 5.** Coupling of FLAC, PFC, and groundwater flow modeling.

**Figure 6.** Illustration of modeling domain. PFC is used for the black area, and FLAC is used for rest of the area.
modeling (Figure 5). The time step used in the mechanical modeling is smaller than the time step used in the flow modeling. The mechanical modeling yields new cavity geometry due to particle movements. The new cavity geometry (i.e., the cavity boundary) is passed to the flow simulation for seepage calculation (Figure 5).

Results of Numerical Simulation
For the flow simulation, the time step of 1 second is selected, and the time step of the mechanical modeling is significantly smaller but determined by FLAC and PFC. A total of 20,000 particles are used for simulating the clay layer and 8,000 particle for the sand layer. The simulation results at the time of 1s, 3s, 5s, 8s, and 15s are selected for analyzing the cavity expansion and the hydraulic head distribution.

Cavity expansion
Figure 7 shows the cavity in the clay layer obtained in a sandbox experiment and at t = 3s, 8s, and 15s of the numerical simulation. The numerical modeling is able to simulate expansion of the cavity. The shape of the simulated cavity is similar to that of the experimental cavity. For example, the angles between the experimental cavity side boundary and the horizontal direction are about 49°~50° (Figure 7), and the corresponding angles of the simulated cavity are about 48°~54°.

Hydraulic pressure and head
Figure 8 shows the vertical profile of hydraulic pressure along the vertical line perpendicular to the opening at the interface between the confining layer and the confined layer. While the pressure profile in the unconfined layer of sand does not change with time, the pressure profile in the confining layer changes dramatically over time, in particular in the early time when the sinkhole starts forming. The pressure change has substantial impacts on cavity geometry and expansion. At the beginning of cavity formation the largest hydraulic gradient occurs at the point of the opening, and it induces a large seepage force on the particles and causes downward movement of clay particles. As a result, the cavity will expand upward to the sand layer until sinkhole collapse. The pressure change in the confining layer happens not only above the opening but at its vicinity, as shown in Figure 9 that plots the spatial distribution of hydraulic head in the entire modeling domain of confining and unconfined layers.

These results indicate that, during the process of sinkhole formation, monitoring hydraulic pressure and hydraulic head in the unconfined layer is not useful, because the two quantities do not change with time. The reason is that the clay layer isolates the pressure propagation to the unconfined layer. When the cavity is expanded to the sand layer pressure change in the layer may be reflected.

Figure 7. Cavity in the clay layer obtained in a sandbox experiment (top) and numerical simulation at t = 3s, 8s, and 15s.

Figure 8. Vertical profiles of hydraulic pressure at different simulation times.
in monitoring wells of the unconfined layer. However, at this stage, the sinkhole has been largely formed, and the monitoring data of the unconfined layer is less useful than the monitoring data of the confining layer. The challenge is to determine where to install monitoring well in the confining layer.

Conclusions
This paper presents a laboratory study for understanding the processes of sinkhole formation, development, and collapse. The experiments are useful not only to illustrate the sinkhole processes but also to develop a conceptual model for mathematical and numerical modeling. The experiments indicate that the confining layer of clay is important to sinkhole formation and development. This paper also demonstrates an approach that uses FLAC and PFC to simulate the laboratory experiments. The results of the numerical simulation are consistent with the phenomena observed in the experiments, in particular the expansion of the cavity due to particle movement caused by the increase of seepage force after hydraulic head in the confined layer drops. The confining layer is of particular importance, because the cavity is first formed in this layer. It is important to monitor

pressure change in the confining layer for detection of sinkhole in its early formation. More effort is warranted to develop more robust numerical models for simulating sinkhole events in the future research.

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References
ACCOUNTING FOR ANOMALOUS HYDRAULIC RESPONSES DURING CONSTANT-RATE PUMPING TESTS IN THE PRAIRIE DU CHIEN-JORDAN AQUIFER SYSTEM – TOWARDS A MORE ACCURATE ASSESSMENT OF LEAKAGE

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Abstract
The Prairie du Chien-Jordan Aquifer system is an important source of drinking water for residents of southeastern Minnesota. Assessment of the hydraulic properties of this aquifer continues to be of interest for wellhead protection and resource evaluation efforts. When performing constant-rate pumping tests on wells constructed in the karsted Prairie du Chien Aquifer, anomalous hydraulic responses resulting from cavernous flow are frequently observed. Hydraulic response in the adjacent Jordan Sandstone Aquifer is also commonly distorted because of bedding-plane fractures and well development techniques such as blasting and bailing. Resolution of these anomalous responses is important for accurate estimates of leakage through adjacent resistive layers. Examples are presented with a rationale for the analysis process.

Introduction
The hydraulic properties of fractured and karsted rocks are notoriously variable. (Teutsch & Sauter, 1991) Standard aquifer tests characterize only a small volume of rock relative to the size of regional aquifer systems. Nevertheless, a sufficiently large number of tests may begin to describe various hydraulic characteristics and types or patterns of hydraulic response may be distinguished. This dataset informs the analysis to reduce uncertainty in estimated hydraulic properties.

Typically, hydraulic conductivity and storativity are derived from a constant-rate pumping test. In a layered system if observation well data were collected and the test period was sufficiently long, the hydraulic resistance of adjacent aquitards may also be estimated. With other information, the hydraulic resistance is used to calculate the rate of pumping-induced leakage to the aquifer. A realistic estimate of the rate of leakage is an important parameter for construction and calibration of regional groundwater flow models of layered systems.

The Minnesota Department of Health (MDH) has compiled a dataset of approximately seventy-five aquifer tests performed on wells constructed in the Prairie du Chien-Jordan Aquifer (PDC-J) system (Figure 1). The tests were conducted for a variety of purposes by drilling contractors, environmental consultants, and state and federal agencies since 1960. The majority of tests were performed on high-capacity wells for wellhead protection planning or to satisfy other regulatory permit requirements. Acquisition and analysis of test data continues.

The extent of the aquifer system in Minnesota is greater than 23,300 sq. kilometers (9000 sq. miles) (Figure 1). This represents a test density of about one test per 310 sq. km. (120 sq. mi). In view of the expense of conducting an aquifer test, the scarcity of appropriately constructed wells from which to obtain water levels and
the large aquifer extent; it is important to maximize the information gained from each test.

Prairie du Chien-Jordan Aquifer Regional Hydrogeologic Setting

Very nearly all of southeastern Minnesota is underlain by a sequence of Paleozoic bedrock formations that comprise the Prairie du Chien-Jordan Aquifer system (Figure 1; Table 1; Olcott 1992; Runkel et al. 2003; Figure 2).

Prairie du Chien Aquifer

The issue of fracture and solution enhanced (cavernous) flow in the Prairie du Chien Aquifer is well documented (Tipping et al., 2007). Karst landforms, sinkholes, caves, springs, and sinking streams are found where the Prairie du Chien Group is the uppermost bedrock unit. Not all void space and preferred zones of cavernous porosity were formed as a result of recent sub-aerial exposure; some were co-depositional. Therefore, zones of conduit flow typical of karst exist throughout the aquifer extent regardless of the type of geologic cover (Tipping et al. 2006). The subaerial exposure of the Prairie du Chien Group has reactivated old features and contributed to the development of new ones. The basal part Prairie du Chien Aquifer, the Oneota Formation, is massive dolostone with low porosity and few fractures. On a regional basis any head differences that are observed between the Prairie du Chien Group and Jordan Sandstone are attributed to the basal Oneota functioning as an aquitard.

Coon Valley Member

The contact of the Prairie du Chien Group with the older Jordan Sandstone is erosional (Tipping et al., 2006; Mossler, 2008). The Coon Valley Member of the Prairie du Chien Group is a dolostone-cemented silty sandstone that formed in topographic lows on the erosional surface on the top of the sandstone. The lateral extent of the Coon Valley Member is limited to the dendritic drainage pattern in which it formed, therefore the occurrence is both limited and quite unpredictable.

Jordan Sandstone

The Jordan Sandstone is a well-sorted medium-grained sandstone which appears to be fairly uniform over its extent. However, locally it contains lithological variations which lower the bulk hydraulic conductivity and bedding-plane fractures which increase the effective conductivity.

Description of Aquifer Test Data

Earlier compilations of this type of data have been limited to the Twin Cities Metropolitan area (Runkel et al. 1999) or selected communities (Runkel & Mossler 2001). The scope of this compilation is the whole extent of the aquifer system in Minnesota. MDH staff collected the test data for approximately half of the tests; the source data of those tests are on file. The raw data of tests conducted by others is also obtained whenever possible. Source data from all but eight tests are available. This is critically important because access to the source data has permitted well-documented tests to be reinterpreted without having to repeat the test.

<table>
<thead>
<tr>
<th>Aquifer Name</th>
<th>Primary Rock Type</th>
<th>Nominal Thickness (meter)</th>
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</thead>
<tbody>
<tr>
<td>Decorah-Platteville-Glenwood Aquitard</td>
<td>Shale and Dolostone</td>
<td>--</td>
</tr>
<tr>
<td>St. Peter Sandstone (OSTP)</td>
<td>Sandstone</td>
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<tr>
<td>Basal St. Peter Sandstone (Twin Cities Basin)</td>
<td>Siltstone and Sandstone</td>
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<tr>
<td>Prairie du Chien Aquifer (OPDC)</td>
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<tr>
<td>Shakopee Fm.</td>
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<tr>
<td>Coon Valley Member</td>
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<tr>
<td>Jordan Sandstone (CJDN)</td>
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</tr>
<tr>
<td>St. Lawrence Aquitard</td>
<td>Dolostone and Shale</td>
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</tr>
</tbody>
</table>

Table 1. Hydrogeologic Setting of the Prairie du Chien-Jordan Aquifer System - Confining Layers are Shaded.
The numbers of tests by aquifer layer are shown in Table 2. The geographic distribution of tests is shown in Figure 1. The greatest numbers of tests have been performed in the Twin Cities metropolitan area and the Bluffland area of southeastern Minnesota where the Prairie du Chien Aquifer is the uppermost bedrock. Comparatively few tests have been performed in the area with a significant thickness of younger bedrock cover.

Descriptive statistics of aquifer properties are shown in Figures 2 and 3. Most tests were conducted in confined to leaky-confined hydraulic conditions. Three tests were performed in a hydraulically unconfined setting. The box and whisker plots, Figure 3, are standard Tukey quartiles; showing interquartile ranges (IQR), the median value and outliers as greater than or less than 1.5 * IQR. Figure 4 shows a cross-plot of leakage factor vs. Storativity. There is an ill-defined unit-slope for these data with slight differences in the populations by aquifer layer.

A detailed account of the analysis methodology for each test is beyond the scope of this paper as most tests were analyzed with multiple techniques. The intent here is to outline the general approach to analysis that was developed for these data, demonstrate that the opposing assumptions of standard transient and steady-state methods are useful to constrain parameter estimates, and how an assessment of well efficiency is required to reduce uncertainty of aquifer properties.

**Well Hydraulics**

The first issue considered in the analysis process is the relative efficiency of the pumped well. The drawdown in the pumping well as compared to that in the aquifer defines the hydraulic efficiency of the well. (Jacob 1947) Two prominent naturally occurring features of the aquifer system, sub-horizontal bedding-plane fractures and solution enhanced conduits, cause wells to produce water very efficiently relative to wells that do not intersect such features. In addition, well development practices affect the hydraulic performance of the pumping well.

As scientists, we have a well-founded bias that efficiency cannot exceed 100%. This is not true for Jordan Sandstone wells that have been developed by the “blast and bail” technique. Dynamite is placed in the well and detonated to break up the rock. Broken rock is then bailed from the well to create a large cavern in the formation. In some cases hundreds of cubic meters of sand are removed from the well. The actual diameter of this cavern is never known exactly but it is certainly much larger than the diameter of the original borehole from the driller’s record.

The combination of natural features and an enlarged borehole causes an increase in specific capacity of the well; a smaller drawdown in the well for the work exerted to remove a given volume of water. Thereby, from the perspective of the aquifer properties - the well appears to be more than 100% efficient. Conversely, a lower transmissivity zone may be encountered by the pumping well, such as the Coon Valley Member. This would result in an under-efficient pumping well because of the local difference in bulk hydraulic conductivity. Once the efficiency of the pumping well is accounted for in the analysis certainty in the local aquifer properties is increased.

No tests, to-date, show the distinctive hydraulic characteristics of linear (√(t)) vs. radial (log [t]) flow in this aquifer system. (Gringarten & Witherspoon, 1971) This is the case even in instances where boreholes show nearly simultaneous response to pumping at different wells. The assumption of porous media flow equivalency in the Prairie du Chien Aquifer is a good starting point for interpretation of test data. However, more subtle effects of radial conduit flow have become evident with the accumulation of test data.

Groundwater flow theory states that the impact of a high permeability fracture intersecting the wellbore will show a characteristic ½ unit-slope on a log-s vs. log-t plot. Similarly, because of increased wellbore storage, a cavern developed in the formation should cause a unit-slope

<table>
<thead>
<tr>
<th>Aquifer Name</th>
<th>Prairie du Chien-Jordan</th>
<th>Prairie du Chien</th>
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<td>1</td>
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</table>

*Table 2. Numbers of tests by layer.*
on the same plot. (Gringarten, 2008) In the existing dataset, the only early-time deviations from the Theis curve have been of the unit slope, characteristic of well-bore storage. This indicates that the hydraulic effect of both the sub-horizontal bedding-plane fractures and zones of cavernous secondary porosity is to create an enlarged borehole.

The dataset of aquifer tests is used to support the development of regional groundwater flow models that are concerned with protecting drinking water quality. As the aquifer system is, by and large, hydraulically confined (Figure 2), the source of water to the system is vertical flow (leakage) through adjacent confining units. Therefore, characterization of leakage is a focus of these analyses.

**Analysis Techniques**

The simple inverse models used to analyze aquifer tests are of two basic types, transient and steady-state. The transient conceptual model (Theis, 1935; Cooper & Jacob 1946) assumes that all water flows to the well from infinity and the aquifer is perfectly confined (no leakage). On this basis, all analyses that exclusively use the Theis conceptual model have limited utility for model development in layered systems.

The steady-state conceptual model (de Glee, 1930; Hantush & Jacob, 1955) assumes that there is no change in storage in the aquifer and that the source of water is from leakage through adjacent aquitard(s). Familiar modifications to the transient conceptual model to include the effects of leakage were introduced by Hantush (1960), Walton (1960), and Neuman and Witherspoon (1972).
Blended opposing conceptual models have been used for many years, however few practitioners think to use the steady-state analysis techniques unless leakage is clearly indicated during a test.

Recharge of some sort must be considered to be present in all tests. All saturated confined-layered systems are replenished by leakage; leakage must be quantified for the aquifer properties to be adequately defined. This approach is not new. For instance, aquifer tests were used to examine the leaky characteristics of ‘tight’ formations for natural gas storage (Witherspoon et al., 1967). However, the reservoir engineering perspective is not often applied to groundwater problems.

**Figure 4. Leakage Factor vs. Storativity by Aquifer.**

**Linked Parameters: T, S, and L**

In the mathematics of the Theis well function, transmissivity (T) and storativity (S) are linked parameters. Basic groundwater flow theory associates ranges of values of storativity with unconfined, leaky-confined, and confined flow regimes. Steady-state flow, as applied to leaky-confined and confined settings, also links transmissivity to the characteristic leakage factor (L). For leaky settings, L is small - on the order of hundreds of meters; conversely for settings with little leakage, L is large – on the order of thousands of meters.

The premise of this analysis procedure is that all three parameters are linked. Therefore for a given trans-
missivity, both the storativity and leakage must 'make
sense' with respect to the hydrogeologic setting. Quan-
titatively, storativity and characteristic leakage factor are
inversely related. A small storativity should be associ-
ated with very small volume of leakage, a large leakage
factor. Conversely, large storativity should be associated
with a large volume of leakage, a small leakage factor.
Given this linkage between the parameters T, S, and L,
an approach has been developed to deal with the effects
of flow through cavernous porosity, bedding-plane frac-
tures, and an enlarged borehole as shown by the follow-
ing examples.

**Examples of Anomalous Hydraulic Response**

When observation wells were used, the most effective
way to detect and then account for the anomalous pat-
terns is to plot all data from a test on one graph with the
Theis distance-drawdown plot, log-s vs. log-t/r².

**Under-Efficient Pumping Well**

A test was performed on a high-capacity well for an etha-
nol plant to satisfy state permitting requirements (Cham-
pion 2008; Blum 2013). In this example, the wells were
completed open-hole over both the Prairie du Chien and
Jordan portions of the aquifer system. The hydrogeo-
logic setting was deep confined with a significant thick-
ness of younger bedrock and clay-rich till cover. There-
fore, the setting should produce a small storativity, in
the range of 10⁻³. Two wells were monitored; pumping
and an observation well, located 479 meters (1570 feet)
apart. Figure 5 shows the effect of an under-efficient
pumping well.

Drawdowns in the pumping well are greater than what
would be predicted by the Theis curve match to the ob-
servation well. The storativity calculated from this data
is large for the setting, at 10⁻⁴.

Figure 6 shows the data from observation well from this
test in more detail with a Theis analysis. It can be seen
that the effect of start of pumping/recovery traveled 480
meters in about 30 seconds. This rate of propagation,
about 16 meters per second, clearly indicates fracture/
conduit flow. The trend of the early-time data (< 10 min-
utes) approximates the unit-slope that is characteristic of
wellbore storage.

The unit-slope departure of early-time data from the
Theis curve in nearby observation wells is frequently ob-
erved in this aquifer system. For tests with several ob-
servation wells the anomalous response is damped and
becomes more ‘Theisian’ in more distant wells. Note

**Figure 5. Example of an Under-Efficient Well.**
that the half-unit slope trend that indicates a high permeability fracture may only be observed in very early time (< 30 seconds) which is rarely captured in sufficient detail.

Interestingly, the unit-slope trend is not often observed in the pumping well itself, such as the case with this test—even those wells equipped with transducers which rapidly collect data in early-time. It appears that transient mechanical and other well hydraulic issues obscure the unit-slope response in the pumping well. For this test there were an insufficient number of observation wells to be able to reliably estimate the characteristic leakage factor. The hydrogeologic setting would cause the leakage factor to be large, probably greater than 6,000 meters (~20,000 feet).

**Over-Efficient Pumping Well**

A recently completed production test of Rochester 41 provides a good example of the effect of an over-efficient well. (Blum, 2014)

Well 41 is located on the east side of the Rochester basin in an area where the uppermost bedrock is the Prairie du Chien Group with scattered patchy remnants of the St. Peter Sandstone. Well 41 was completed in the Jordan Sandstone and had been developed by the blast and bail method. Approximately 115 m$^3$ (150 cubic yards) of sand were removed from the well. Based on the volume of sand removed and the length of open hole, the enhanced borehole radius is at least 1.1 m (3.6 feet).

The production test was conducted at a constant rate of 4580 liter/minute (1210 gallons per minute) for 24-hours. The driller measured water levels manually at the pumped well. Observation wells were four other nearby Rochester Public Utilities (RPU) water supply wells which were monitored by the utility’s data acquisition and control system on a five-minute interval (Figure 7).

Other known interfering wells were the production wells in the RPU system. The manual data collection at the pumped well precluded any early-time data but those data collected appear to be of high quality. Problems in data collection/reporting were that some transducers in the wells report to nearest 1 to 2 feet rather than the normal 0.01 foot resolution. This is a common problem when these data are used for scientific purposes rather than the design purpose.
For this reason, plots of all water level readings, collected on a five-minute interval, appear ‘noisy.’ The analysis plots use water level readings from the observation wells on a half-hour interval for a more clear view of hydraulic response.

Well interference was a significant factor as three of the four observation wells were affected by cycling of their pumps at some time during the test. In particular, data from well 27 (224212) was not usable as this well was in recovery over the time span of the test and no discernable drawdown caused by Well 41 was found. Wells 21 and 23 cycled after 23-hours into the test and therefore only the pumping-phase data were usable.

Limited recovery data were collected from the pumped well only. This test setup was not ideal; however, serviceable data were obtained for analysis. The Theis $t/r^2$ analysis of these data is shown on Figure 8.

There is significant scatter in the drawdowns at the observation wells. The response of Well 21 may be influenced by multi-aquifer well construction. The response of Well 23 may be influenced by proximity to surface

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**Figure 7.** Wells Monitored During W41 Production Test.

**Figure 8.** Initial Analysis, Theis $t/r^2$ plot.
water. Though, it is common for drawdown at a distant observation well to be smaller than expected, with respect to the Theis curve. Therefore, the analysis weights the response of the closest Jordan Sandstone well, Well 33, most heavily.

A strong indication that anomalous hydraulic responses are present in the data is that there is significant variability in aquifer properties between wells and analysis techniques (Table 3).

The effective radius of the pumped well may be estimated by different means depending on the quality of the dataset. If a sufficient number of observation wells at a range of radial distances were used a semi-log regression on the distance-drawdown data using the observation wells only provides an initial value of $r$. Otherwise, trial and error adjustment of the radius of the pumped well to match the Theis curve is the only recourse. The resulting estimate of effective radius must then be tested with other analysis plots to verify consistency with the hydrogeologic setting.

In this case, the distances to the observation wells are not well-distributed; all observation wells are greater than 1000 m from the pumped well. A semi-log distance-drawdown regression is not informative. Trial and error adjustment of the effective radius of the pumped well resulted in an optimal value of $r$, 7.6 m (25 feet) and a value of transmissivity of 287 m$^2$/d (Table 4). With the revised wellbore radius visual matches to the type-curves are consistent between analysis techniques (Figures 11 and 12).

Without the revised wellbore radius of the pumping well, there is significant uncertainty in the match to the steady-state well curve; particularly with respect to the horizontal match, L (Figure 12). In addition, if a borehole radius of 0.3 m (1 ft) is used for the Agarwal recovery analysis (Agarwal, 1980), this results in a non-credible storativity of 0.5 (Figure 9). Whereas, using a radius of 7.6 m in the calculation, the resulting storativity is large, $8.2 \times 10^{-4}$, but is physically possible.

<table>
<thead>
<tr>
<th>Figure</th>
<th>Transmissivity m$^2$/d</th>
<th>Storativity</th>
<th>Well - Analysis Method</th>
</tr>
</thead>
<tbody>
<tr>
<td>8</td>
<td>445</td>
<td>8.4e-5</td>
<td>Theis $1/r^2$</td>
</tr>
<tr>
<td>9</td>
<td>180</td>
<td>5.3e-1</td>
<td>W41-Agarwal</td>
</tr>
<tr>
<td>10</td>
<td>350</td>
<td>8.7e-5</td>
<td>W33-Theis</td>
</tr>
</tbody>
</table>

Table 3. Initial Assessment of Aquifer Properties.

Figure 9. Agarwal Recovery Analysis.
The estimate of \( L \) from the Hantush-Jacob analysis, Figure 13, is not as reliable as that from the de Glee analysis, Figure 12. This results from the sensitivity of this technique to an error introduced by using wells located at distances greater than 0.2 \( L \). In this case, no wells within 0.2 \( L \) were monitored and the Hantush-Jacob analysis is poorly constrained.

**Conclusion**

Application of this method depends on the number of observation wells monitored during a test. Given a sufficient number of observation wells, distance-drawdown analysis techniques, \([\log(s) \text{ vs. } \log(t/r^2)\text{ and } \log(s) \text{ vs. } \log(r)\])\], produce the most consistent results. Uncertainty in transmissivity values is significantly reduced by accounting for well efficiency, avoiding data influenced by wellbore storage, and solving for consistent values of linked parameters; storativity and leakage factor.

As described, this procedure accounts for peculiarities of the hydraulic response in wells in this aquifer system to arrive at a consistent transmissivity based on the different methods.

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**Table 4. Aquifer Properties after Accounting for Large Effective Borehole Radius.**

<table>
<thead>
<tr>
<th>Figure</th>
<th>Transmissivity ([\text{m}^2/\text{d}])</th>
<th>Storativity</th>
<th>Characteristic Leakage Factor ([\text{m}])</th>
<th>Analysis Method</th>
</tr>
</thead>
<tbody>
<tr>
<td>9</td>
<td>180</td>
<td>8.2E-04</td>
<td>--</td>
<td>Agarwal</td>
</tr>
<tr>
<td>11</td>
<td>287</td>
<td>8.0E-05</td>
<td>--</td>
<td>Theis ( t/r^2 )</td>
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<tr>
<td>12</td>
<td>287</td>
<td>--</td>
<td>1830</td>
<td>de Glee</td>
</tr>
<tr>
<td>13</td>
<td>288</td>
<td>9.9E-05</td>
<td>2380</td>
<td>Hantush-Jacob/Cooper-Jacob</td>
</tr>
</tbody>
</table>
Figure 11. Theis $t/r^2$, Projected to $r = 7.6$ m.

Figure 12. Log-log Distance-Drawdown.
Figure 13. Semi-log Distance-Drawdown.

Different assumptions of the source of water to the well. If the large effective borehole radius is not recognized and accounted for, the apparent variability of all hydraulic properties; transmissivity, storativity and leakage factor increase. And more significantly, without the enlarged borehole radius, the correlation between the linked parameters is weakened.

Uncertainty in storativity is low because it is based on a time of propagation of a pressure change in the aquifer that is relatively insensitive to differences in transmissivity. The least certain parameter is the characteristic leakage factor. This approach to aquifer test analysis has significantly reduced the uncertainty in local aquifer properties and enhanced the quality of the overall dataset.

References
Agarwal RG. 1980. A new method to account for producing time effects when drawdown type curves are used to analyze pressure buildup and other test data. In Proceedings of SPE. 55th Sociey of Petroleum Engineers Annual Technical Conference and Exhibition. Dallas, Texas.


NUMERICAL SIMULATION OF SPRING HYDROGRAPH RECESSION CURVES FOR EVALUATING BEHAVIOR OF THE EAST YORKSHIRE CHALK AQUIFER

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Abstract
The Cretaceous Chalk aquifer is the most important in the UK for the provision of water to public supply and agriculture. The Chalk has both matrix and fracture porosity and is thus best considered as a dual porosity aquifer system. Although the matrix porosity is large, typically around 0.35 in the study area of East Yorkshire, UK (ESI, 2010), pore diameters are typically very small, and the water contained in them is virtually immobile. The high permeability fracture network is responsible for the ability of water to drain; spatial variations in fracture network properties mean conventional approaches to aquifer characterization such as borehole pumping tests are of limited utility. Hence this study attempts to better understand the flow system and characterise aquifer properties from the recession response seen at springs during the spring/summer period when recharge is minimal. This approach has the advantage that spring hydrographs represent the sum of the response from entire catchments.

This paper reports numerical modeling for simulating aquifer and spring responses during hydrological recession. Firstly, available geological and hydrogeological information for the study area was used to develop hydrogeological conceptual models. Three different numerical models have been constructed representing three possible scenarios that could represent the aquifer in the selected area. These are: single reservoir aquifer, double reservoir aquifer, and single reservoir aquifer containing tunnel shaped highly permeable zone at the spring elevation respectively. The sensitivity of spring recession response to various external and internal parameter values was investigated, to understand relations between spring recession, hydrological inputs (recharge) and aquifer structure. Spring hydrographs from the real aquifer were compared with the hydrographs generated from models, in order to estimate aquifer properties. The work aims to identify the utility of spring hydrographs in eliciting aquifer permeability structure, as well as identifying the conceptual scenario which best represents the Chalk Aquifer in East Yorkshire, UK.

Introduction
The Chalk is the most significant aquifer in Britain; it underlies much of eastern and southern England. Groundwater from the Chalk aquifer of Yorkshire is an important resource for public supply, agriculture and industry.

Two types of porosity systems have been recognized in the chalk rocks: primary and secondary porosity. The primary porosity is pore spaces formed between rock grains during rock formation processes, simply termed “matrix porosity”. Secondary porosity exists in the form of fractures which were produced by dissolution and tectonic activity (Singhal and Gupta, 2010). This characteristic of dual porosity in the Chalk aquifer was confirmed by many studies (Foster and Crease, 1974; Wellings and Bell, 1980; Price, 1987; Price et al., 1993; Downing, et al., 2005; Mathias et al., 2005).

The role of the porosity systems within the Chalk aquifer are as follows: the fracture system has very low porosity but high permeability which makes it dominate the flow system, while the matrix has very high porosity but low permeability so seldom contributes (Allen et al., 1997; Gale and Rutter, 2006). The storage co-efficient (specific yield) is also likely to derive from drainage of fracture space, rather than matrix porosity (MacDonald and Allen, 2001).
Increasing overburden with depth gave the Chalk a significant feature which is developing permeability toward the top remarkably. Overburden affects the permeability in two ways, first reducing the fracture and aperture size. Second, because of lack of groundwater circulation it prevents processes of fracture enhancement due to dissolution (Foster and Milton, 1974; Foster and Robertson, 1977; Price et al., 1977).

Hydrographs are graphical representations of the time series flow rate, generally consisting of three segments, rising limb, peak and falling limb, respectively. The falling limb, which is also known as a recession curve, is that part of a hydrograph that comes after peak flow. Studying hydrograph recession curves of springs may provide hydrogeological information especially where fracture or conduit flows are significant. This approach is preferred over other geological and geophysical methods (Dreiss, 1982; Bakalowicz, 2005) because the spring drains water from large areas of aquifer, so the discharge is governed by accumulative effect from the flow systems that exist in the aquifer. This contrasts with other geological and geophysical methods that only represent the aquifer locally at the investigation points.

Factors affecting hydrograph shape essentially grouped into two groups, external and internal factors. External factors include physiography, climate and vegetation which control recharge, while internal factors are the hydrogeological properties of the aquifer rocks, such as transmissivity (product of aquifer thickness and hydraulic conductivity). Precipitation intensity, duration and distribution over the catchment influence shape of the hydrograph; intensity and duration of rainfall strongly affect the peak flow. Temperature and humidity influence evapotranspiration and effective rainfall. Catchment size, shape, slope and morphology (surface depressions can act as natural water storage ponds) are important external factors.

It has been reported from comparison between the spring hydrograph recession curve of different springs, that the recession curves steepness and shape (i.e., recession coefficients) are mainly governed by the intensity and geometry of fracture system (Kovács et al., 2005). Based on the analytical curve fitting method based on the Maillet exponential model, it has been suggested that the recession of spring hydrographs from fractured rock aquifers decomposes into several segments, each segment reflecting different flow system in the aquifer (Kovács and Perrochet, 2008; Liu and Li, 2012). However, the analysis of spring recession curves simulated by numerical modeling revealed that multiple segments do not necessarily reflect the presence of multiple flow systems (Baedke and Krothe, 2001; Kovács, 2003). In our study, we investigate the extent to which recession curve shape can provide information about the permeability structure and characteristics, using numerical simulations of flow in conceptual permeability scenarios based on those potentially found within the case-study aquifer.

Site Location and Characterization

The field study area is located at northern part of Yorkshire Wolds of East Yorkshire, it occupies an area about 250 km² (Figure 1A). Two gauging stations exist in the study area, one located at Kirby Grindalythe village in the NW of the study area and second one in Driffield town in the SE of the study area. This paper focuses on the Kirby Grindalythe catchment as this is closer to the topographic divide (Figure 1B), so the catchment boundary conditions are easier to constrain.

The Cretaecous Chalk crops out across the study area and is overlain by glacial sediments to the East. Chalk rocks rest unconformably on Jurassic rocks of the Penarth group (largely argillaceous) and Lias Group (mudstones and thin silty limestone). A schematic diagram of the Geological cross section in the area is illustrated in Figure 1B.

The Gypsey Race is the most significant surface water course in the area, it rises through a series of springs just upstream of Kirkby Grindalythe village and runs eastwards to Bridlington. The Kirby Grindalythe gauging station measures the discharge in the upper reaches of the Gypsey Race, just downstream from these springs.

The unconfined Chalk aquifer is covered by a shallow lime-rich sandy soil on the interfluves and by a lime-rich loamy soil along the water drainages and dry valleys. Both soil types allow the water to freely drain. Figure 1A illustrates location of the study area.

Methodology

To investigate factors that govern groundwater flow in the aquifer, we analyze that part of recession curve representing water discharge in the absence of recharge,
Recession curves show variation in the peak flow at starting recession period, starting date and length of recession period between different water years. To understand relation between this variation in recession curves from same sources and rainfall the total annual effective rainfall has been calculated from climate data (from UK MORECS data) for the years between 2010 to 2014 and then plotted simultaneously with hydrograph for same years. Figure 3 is graphically showing relation between annual total effective rainfall and spring hydrograph.

To overcome the problem of variation which exists between recession curves from different years a master recession curve MRC technique was used for constructing a mean recession curve. Several approaches can be used for constructing a master recession curve: e.g., matching strip, correlation and tabulation method (Brownlee, 1960; Toebes and Strang, 1994; Hall, 1968; Toebes, 1969; Brutsaert and Nieber, 1977; Sugiyama, 1996). In this study the tabulation method was used as it is the most appropriate technique for constructing a MRC for a range of years. In the tabulation method the recession data at regular intervals of time are tabulated in columns, each recession in separate column. The columns are adjusted vertically until the discharge values approximately agree horizontally (Figure 4).

i.e., the recession curve. Actual evapotranspiration (AE) and soil moisture deficit (SMD) information from the UK Meteorological Office Rainfall and Evapotranspiration Calculation System (MORECS) database have been used to identify date of the cessation of recharge and hence the start of flow recession. Figure 2 shows hydrograph recession curves from the Kirby Grindalythe gauging station for selected hydrological years between 1998 and 2014.

As it appears in the figure starting time and length of the recession period was different from year to year, but generally began between February 15 to April 15 (except 2000 when recession started middle of June), and recession ended on early to late September.
The average discharges are calculated, representing the master recession curve. Figure 5 shows construction of a master recession curve for the Kirby Grindalythe station.

The analytical model suggested by Maillet (1905) (Toebes and Strang, 1994; Tallaksen, 1995; Stella, 2013; Eslamian, 2014; Hingray, et al., 2015) was used for initial interpretation of recession curves. This method is the most widely used approach for describing the flow depletion during recession period. The model is expressed by the equation:

\[ Q_t = Q_0 \exp(-\alpha t) \]

Where \( Q_t \) and \( Q_0 \) are flow \([\text{L}^3/\text{T}]\) at time \( t \) \([\text{T}]\) and the start of recession, and \( \alpha \) is the recession coefficient \([\text{1/T}]\).

**Figure 3.** A is annual total effective rainfall for years between 2010 to 2014 over Kirby Grindalythe and Driffield catchments. B. Hydrograph for Kirby Grindalythe station. C. Hydrograph for Driffield station

<table>
<thead>
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</table>

**Figure 4.** Calculation of MRC using the tabulation method.

**Figure 5.** MRC and recession curves from 2000 – 2014 at Kirby Grindalythe gauging station.
The MRC was fitted with the Maillet recession equation by plotting the recession hydrograph on semi-log graph, discharge plotted on the log axes and time on the ordinary axes. It shows a good fit with a single segment, with recession coefficient \(0.017 \text{ day}^{-1}\) (Figure 6).

This paper next examines how recession curves relate to the aquifer permeability structure. Numerical modeling was used to investigate the response of the recession curve to different aquifer permeability scenarios. The models aimed to simulate the spring drainage for the real Chalk aquifer catchments in the area. Both Kirby Grindalythe and Driffield catchments were simulated, but only the former are presented here. Saturated thicknesses of the aquifer, boundary conditions given by catchment water divides, and geological information from previous studies were used in formulation of the conceptual model for each catchment.

Figure 7 shows a conceptual model for the Kirkby Grindalythe catchment. Catchment boundaries were based on topography. The conceptual model was then translated into a numerical simulation grid. Figure 8 illustrates a schematic diagram of the 3D model grid.

Figure 6. Analysis of MRC depending on Maillet model. (A) semi-log graph, the \(R^2\) between the MRC and fitted recession line is 0.99 and recession coefficient \(0.017 \text{ day}^{-1}\). (B) Black curve is MRC from observed discharge; red represents fitted curve to MRC which was calculated based on Maillet equation.

**Groundwater Flow Model**

A transient three dimensional numerical model was developed using Groundwater Vistas to simulate water drainage via a spring (Figure 8). The model was discretized into a uniform grid of finite-difference cells consisting of 70 rows by 45 columns of 100 m x 100m cells and vertically with 15 layers of cells of 2 m thickness. To represent aquifer drainage via a spring during the recession period, no rainfall recharge was added; instead the model was run from an initial head representing that at the start of the recession period.

The aquifer was modeled as unconfined; water depletes from the aquifer through a spring freely under the influence of gravity. The spring was simulated using a drain cell located at the level of the base of the model with very high hydraulic conductivity so as not to mask the conductivity in the aquifer. The modeled catchment was surrounded by no-flow boundaries representing the catchment divide. The soil zone was not explicitly represented in the model, because soil permeability is high enough to allow rainfall infiltration at all times.

Four targets (representing monitoring wells) were placed along the mid-plane of the model containing the drain cell. One of the targets was located at the drain cell for the purpose of recording the flow during recession while the other three targets were located at different distances upstream from the drain cell (100 m, 1,200 m, and 2,500 m) for monitoring hydraulic head.
To investigate the effect of hydraulic conductivity heterogeneity on spring recession three scenarios were tested. All simulations had the same boundary and initial conditions. Figure 9 schematically shows the scenarios tested.

First Scenario (Figure 9A): homogenous and isotropic aquifer.

Second Scenario (Figure 9B): heterogeneous aquifer, consisting of two parallel horizontal reservoirs, with different hydraulic properties. The lower reservoir represents a high permeability zone, corresponding to zone just below the level of water table fluctuation, where the maximum flow occurs. This zone is recognized to have very high hydraulic conductivity in chalk aquifers because of fracture enhancement due to calcite dissolution. The upper reservoir represents cumulative effect of the matrix, small fractures with lower permeability; this zone has been subjected to less water flow so fracture solution enhancement is less well developed.

The low permeability zone which is symbolized by K1 occupied 22m of the total model thickness and the high permeability zone symbolized by K2 occupied the 8 m thickness of the model.

\[ K_1 < K_2 \]

Third Scenario (Figure 9C): A relatively low permeability aquifer contains a longitudinal-tunnel shaped high permeability zone at the drain cell level. This geometry represents a high permeability major fracture zone or solution conduit. The highly permeable zone works as the transporting medium and the less permeable surrounding rock as a storage reservoir.

Figure 8. Schematic illustration of model grid for simulating spring recession in Chalk catchments in the study area (Note: cells are shown larger than actual size relative to catchment dimensions for clarity).

Figure 9. (A) Single reservoir aquifer. (B) Double reservoir aquifer, parallel reservoirs model. (C) Double reservoir, tunnel model.
The high permeability zone is an elongated cuboid with the plan dimensions of 2,000 m x 100 m, and thickness of 8 m, located at the base of the model and at the level of drain cell.

**Hydraulic Conductivity Sensitivity Test**

Sensitivity tests for hydraulic conductivity (K) have been accomplished for all models. All the other conditions and parameters stayed unchanged. The models were run with zero recharge and initial head of 30 m above model base. This thickness is based on the water table map of the area provided by the British Environment Agency. Storage coefficient and specific yield were set to fixed values of 0.0001 and 0.01 respectively (Allen et al., 1997; Gale and Rutter, 2006; ESI, 2010).

Table 1 summarizes input values used for testing sensitivity to hydraulic conductivity; K represents the hydraulic conductivity in homogenous single reservoir aquifer model, K1 and K2 are hydraulic conductivity of low permeability and high permeability reservoirs respectively in the double reservoir aquifer models.

Note that the hydraulic conductivity of the low permeability zones (K1) remained constant while conductivity value of high permeable zones (K2) were changed; this is because the high permeability zones have more significant impact on the recession curve.

The last stage of development was calibration of the models against recession data from field measurements. Calibration was accomplished by using the trial-and-error method (Anderson and Woessner, 1992). For the Kirby Grindalythe catchment model, both single reservoir and double reservoirs simulations were calibrated against field data. Figure 10 demonstrates results of calibration between observed MRC and the recession curves obtained from the numerical models.

**Results**

The recession curves from tunnel and double reservoir models reveal that at the early stage of the recession period the flow rate falls rapidly then flattens off (Figure 10). This pattern of recession for the tunnel model appears more clearly when the contrast between hydraulic conductivity of the block and tunnel zone is larger. The steep initial recession curve arises from rapid hydraulic head depletion within the high permeability zone; the slower recession later reflects drainage behavior from the low permeability zone in the model.

Recession curves from single and parallel horizontal reservoir models are shown in Figure 11; both models behave similarly where the thickness of the high permeability zone within the parallel horizontal reservoir model was about 25% or more of the total aquifer thickness (black and green curves in Figure 11).

The high permeability zone clearly has a dominant impact when its size is sufficient such as to force the

| Table 1. Values of hydraulic conductivity in m/day used for sensitivity test. |
|-----------------------------|-------|-------|-------|-------|-------|     |
| test # | 1 | 2 | 3 | 4 | 5 | 6 |
| single reservoir | K | K | K | K | K | K |
| test # | 1 | 2 | 3 | 4 | 5 |
| 2 parallel reservoirs & Tunnel model | K1 | K2 | K1 | K2 | K1 | K2 |

**Figure 10. Result of calibration between MRC and recession curve deduced from the tested numerical models (s – single porosity model; p – parallel reservoir model; t – tunnel model; numbers are K2; K1 = 1 m/day in all the models shown).**
The coefficient of regression (R-squared) between recession curve from the best fitting single reservoir model and MRC for the period between 2000 to 2014 was 0.79 – the fit is not perfect because the model curve falls rather more steeply than the MRC initially and later flattens off more. Nevertheless, the model curve does fall within the range of behavior seen in the recession curves for individual years.

The results from calibration tests for Kirby Grindalythe catchment suggest that the recession curve which was produced from the single reservoir model with a calibrated K value of 125 m/day show best agreement to the field recession curve (Figure 10). Given the initial model saturated thickness of 30 m this indicates a maximum model transmissivity value of 3,750 m$^2$/day.

The mean transmissivity value in the Yorkshire Chalk is about 1,250 m$^2$/day obtained from borehole measurements (Gale and Rutter, 2006). A pumping test in a Low Mothrope borehole close to the catchment area shows a transmissivity value of 450 m$^2$/day (Figure 12), whereas a pumping test at Etton south of the study area shows transmissivity values of 1,000-2,200 m$^2$/day (Gale and Rutter, 2006). The above data suggest that the spring recession-derived T values may be higher than those likely to be observed in pumping tests. The recession derived K value of 125 m/day agrees better with K found by calibrating numerical simulations (e.g., 4 to 170 m/day, University of Birmingham, 1978 from Allen et al. 1997; Jones et al. 2000). This result suggests that spring hydrograph analysis can be a better choice for deriving hydraulic properties representative of the catchment scale than pumping tests, where complex fracture system are responsible for the permeability. In these cases, borehole tests may not be as representative, as they offer information only at and near the drilling site.
discharge and model discharge were analysed using the analytical exponential model of Maillet, to identify the aquifer permeability scenario that best matched the field data.

This study confirms that the highly permeable fracture system dominates flow in the Chalk aquifer in the study area. Moreover, it revealed that in such complex fractured aquifers, the hydraulic parameters measured through borehole tests may not be representative; transmissivity values obtained from model calibration are higher than those from the borehole tests in the area.

References


STUDY ON THE CRITICAL VELOCITY OF GROUNDWATER TO FORM SUBSIDENCE SINKHOLES IN A KARST AREA

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Abstract
Subsidence sinkholes in a karst area are a common geological hazard causing disaster accidents. However, the critical hydraulic conditions of forming subsidence sinkholes have not been well understood. Based on the theories of pipe flow, this paper derives expressions of a critical hydraulic condition for assessing whether it leads to subsidence sinkholes. A case study with samples of the cohesive soils taken from the Wuxuan county, Guangxi province, China, was conducted to evaluate the derived critical value. Combining the derived critical value and the monitoring data of hydraulic conditions in the study area our results indicate that subsidence sinkholes are not forming under the current hydraulic conditions.

Introduction
Subsidence sinkholes in karst terrain are a common geological hazard, which can lead to natural disasters in the natural and built environment. One of the main triggering factors of subsidence sinkholes is groundwater. The groundwater alterations may trigger or accelerate sinkhole formation. But it is difficult to obtain the critical condition of the groundwater variation to form subsidence sinkholes. The objective of this study is to discuss the issue.

Hydrodynamic condition
Subsidence sinkholes are mainly caused by water seepage. In a seepage process, water seeps into soil through the pores (Figure 1a), and it takes the shear stress on the soil grains of the pore wall. When the shear stress is enough to break the soil, it leads to soil-caves. It is hard to make the properties of water seeping into the soil pores clear, as the natural pores of soil are irregular. Ideally, the study makes the soil pores to be regular circular pipes with certain diameters (Figure 1a).

Therefore, water seepage into soil pores is equivalent to water flow in pipes (Figure 1b), which is in accordance with the piping flow theories.

Assuming that water seeps into pores its characteristics can be analyzed with the piping flow theories in order to obtain the critical values of suffosion. An arbitrary micro-unit of the soil cross section is selected as the study object, and an irregular pore of soils is treated as the pipe of \( d \) in diameter shown in Figure 1b. Ac-
According to the piping flow theories, the water shear stress to the pore wall can be formulated as:

$$
\tau = \gamma RI
$$

(1)

Where $\gamma$ is water unit weight; $R$ is hydraulic radius; $I$ is hydraulic gradient in the pipe. According to the Newton Inner Friction Law, it can be formulated as follows:

$$
\tau = -\mu \frac{dV_r}{dr}
$$

(2)

Where $\mu$ is the dynamic viscosity coefficient of water, and $V_r$ is the flow velocity at $r$ away from the pipe axis.

According to the boundary conditions, the pipe wall velocity is 0:

$$
V_r \bigg|_{r=r_0} = 0
$$

(3)

After integration, it can be defined as:

$$
V_r = \frac{\gamma}{4\mu} (r_0^2 - r^2)
$$

(4)

The mean velocity is:

$$
\overline{V} = \frac{Q}{A} = \frac{2\pi \tau_0 t}{\pi r_0^2} = \frac{Q}{8\mu r_0^2}
$$

(5)

It can be further calculated as:

$$
I = \frac{8\mu \overline{V}}{\gamma r_0^2}
$$

(6)

For the pipe:

$$
R = \frac{r_0}{2} = \frac{d}{4}
$$

(7)

After the substitution of formula 7:

$$
I = \frac{32\mu \overline{V}}{\gamma d^2}
$$

(8)

Where $d$ is the diameter of the pipe. After the introduction of the critical seepage quantity:

$$
Q_{cr} = \frac{\pi d^2}{4} \overline{V}_{cr}
$$

(9)

After the combination of formulas (1), (8) and (9), the critical shear stress, mean flow velocity and hydraulic gradient are obtained (Jiang 2014):

$$
\tau_{cr} = \frac{32\mu \overline{V}_{cr}}{\pi d^3}
$$

(10)

$$
\overline{V}_{cr} = \frac{\tau_{cr} d}{8\mu}
$$

(11)

In the seepage process, if the critical shear stress—which can cause suffosion—is $F$ it is correlated with the acting force between soil grains (or aggregates). $F$ is an inherent attribute which is unrelated with the fluid and the pipe diameter. When the fluid shear stress reaches the clay critical shear stress $F$ the soil grains separate and move away. Under ideal conditions (the identical properties, the constant acting force between the wall grains, and the uniform distribution) clays have an invariant critical shear stress in the seepage process of pores with different sizes. The following is a formula based on equation 11, with different pipe diameters, i.e. $d_1$ and $d_2$ (Jiang 2014):

$$
\tau_{cr} = \frac{32\mu}{\pi} \frac{Q_{1cr}}{d_1^3} = \frac{32\mu}{\pi} \frac{Q_{2cr}}{d_2^3}
$$

(12)

Formula 12 indicates that the critical shear stress can be obtained by simulating suffosion in large diameter pipes of soil, and then the critical hydraulic conditions of different small diameter pores can be worked out. Based on this principle, the critical shear stress can be tested by means of the seepage of large diameter pores.

**Test methods**

**Test procedure**

The pipe flow test device (Figure 2) can be divided into two main parts: hydrodynamic system and sample pot.
system, simulating seepage hydraulic pressure and pipe flow respectively. The hydrodynamic system is mainly composed of a water pipe and discharge valve. Water is filled in the water pipe, which raises the water level. The height of the drain valve and the water inflow are adjusted to maintain a constant seepage pressure, which can be read through the sighting tube.

The sample pot is modified in the pinhole test device. Small gravels are put into the pot to avoid the water flow rushing right into the sample. There are permeable stones between the sample and the tail cover, preventing the wall effect at the tail. Pores of a certain diameter are pierced through the middle of the sample to simulate pipes.

The procedure of experiment includes the following six steps:

a) Sample preparation: the sample is 90 mm in diameter and 38.1 mm in thickness, and the pipe is 1 mm in diameter.
b) Follow the schematic diagram to install the device.
c) Saturate the sample.
d) Increasing the water head: fix the drain valve at the level of the starting water head, and fill water into the water pipe until the water runs out from the drain valve. Use the same method to increase the head with 10 cm increments until soil failure. Keep heads for 30 minute intervals.
e) Measurement and observation: It is crucial to measure the flow quality and water temperature accurately. Then observe the seeping phenomenon every 10 minutes, such as the permeating capacity of water and water turbidity, bubbles, and flowing particle, and record to the corresponding water head.
f) Test termination: When the soil breaks through to the phenomenon of suffosion, e.g. particle loss, water head drastically decreases suddenly, water turbidity increases, etc., the test is terminated.

Critical seepage quantity
Before the sample becomes eroded, the seepage quantity adjusts itself to the ups and downs of the seepage water head value, with roughly the same increase extent. During the variation, slight turbidity and deposition occur in the flow water. Occasionally, the seepage quantity decreases as the seepage water head increases. This is because the debris, which are formed in the process of sample preparation and pipe piercing, are eroded and washed away by the water flow making the pipe clogged up. The distribution of the sample data can be seen in the unbroken state curve in the semilogarithmic plot in Figure 3.

When the sample is eroded, the seepage quantity generally becomes turbid and has coarse sediment grains, the seepage quantity increases significantly, and the seepage water head changes. It can be viewed as suffosion: a) when the seepage quantity increases, with the difference of the increase between seepage quantity and seepage water head.
head more than 20% (the seepage quantity increases with the water head, with a typical increase extent of about 10%. Accordingly, an increase extent of more than 20% can be recommended to be one of the evidences for suffosion.); b) the seepage water head decreased greatly; c) the flow water becomes turbid; d) the coarse sediment grains appear. In the semilogarithmic plot, the seepage discharge manifests itself as stair-like upward inflection points (as shown in the broken state curve in Figure 3). After the test, it is found that, the pipe diameter of the pipe above the water becomes larger and grains are washed away (Figure 4a, b).

After the suffosion, the value of the critical seepage quantity can be estimated as that of the previous seepage quantity, namely, the low value of the inflection points shown in Figure 3.

Application in Wuxuan
Wuxuan is located at the center of Guangxi province, China (N23°35′53.57″, E109°39′34.43″), which strongly develops karst and is covered by clay. It has the proper geological condition to form sinkholes.

The soil sample was taken undisturbed from a borehole (ZK5), Specimens were extracted at the bottom of layer of a depth 9.0-10.0 m. Three samples were tested to obtain the critical shear stress.

Critical shear stress
Table 1 shows the test result according the test method mentioned above. The critical seepage quantity is from 1.754 ml/s to 1.806 ml/s. Water temperature is also tested to obtain the dynamic viscosity coefficient. Re (Reynolds number) shows water flow in the test is in a state of laminar flow (Re<2000). Therefore, it calculates the critical shear stress effectively by equation 12. The mean value 21.5 Pa is the critical shear stress of clay in the study area.

Pores diameter
Pore diameters were measured by a scanning electron microscope. The circle equivalent diameters of 43 pores were measured in vision field of 120,000 times mag-

<table>
<thead>
<tr>
<th>No.</th>
<th>Critical seepage quantity (ml/s)</th>
<th>Water temperature (°C)</th>
<th>Dynamic viscosity coefficient ($10^{-3}$ Kg/ms)</th>
<th>Reynolds No.</th>
<th>Critical shear stress (Pa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ZK5-1</td>
<td>1.806</td>
<td>12.3</td>
<td>1.225</td>
<td>1877</td>
<td>22.5</td>
</tr>
<tr>
<td>ZK5-2</td>
<td>1.766</td>
<td>14.4</td>
<td>1.157</td>
<td>1943</td>
<td>20.8</td>
</tr>
<tr>
<td>ZK5-3</td>
<td>1.754</td>
<td>13.7</td>
<td>1.179</td>
<td>1894</td>
<td>21.1</td>
</tr>
</tbody>
</table>

Table 1. The critical shear stress of samples.
nification. Measured results presented the minimum at 15.75 nm, the maximum at 72.08 nm, and the average at 31.6 nm. Therefore, 31.6 nm is the representative diameter of the tested sample.

Critical velocity
The critical velocity of suffosion is calculated by equation 11. Water temperature is monitored every 20 minutes by an observation system in 2013 (all the time of the year). The max temperature is 22.6 °C, and the min is 22.3 °C. The average 22.38 °C is the water temperature in 2013; to obtain that μ is 0.948×10⁻³ Kg/ms. τ is 21.5Pa, and d is 31.6nm, therefore Vcr is 0.5375 cm/min by equation 11.

Assessment
The described test accesses quantitatively the risk of subsidence sinkholes via hydraulic condition. If groundwater velocity is more than (has risk) the critical velocity or not. Figure 5 shows the relationship between groundwater velocity and the critical velocity. It indicates that groundwater velocity is less than the critical velocity. As a result, Wuxuan has no risk of subsidence sinkholes caused by suffosion.

Conclusion
Suffosion is a main factor that leads to subsidence sinkholes. It derives the formula of the critical groundwater velocity of suffosion via the piping flow. Three samples from Wuxuan were tested by self-designed equipment to obtain the critical velocity, 0.5375 cm/min. Compared with the monitored hydrodynamic condition, groundwater velocity is less than the critical velocity. As a result, Wuxuan has no risk of subsidence sinkholes for suffosion.

Reference

Figure 5. Velocity of groundwater in 2013 year. That the velocity is less than 0 indicates the moving direction of groundwater is upward. When the velocity is more than 0, the moving direction is downward.
EVALUATION OF FIRST ORDER ERROR INDUCED BY CONSERVATIVE-TRACER TEMPERATURE APPROXIMATION FOR MIXING IN KARSTIC FLOW

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Abstract

Fluid dynamics in karst systems is complex due to the heterogeneity of hydraulic networks that combine the Porous Fractured Matrix (PFM) and the interconnected drains (CS). The complex dynamics often requires to be treated as “black boxes” in which only input and output properties are known. In this work, we propose to assess the first-order error induced by considering the temperature as a conservative tracer for flows mixing in karst (fluvio-karsts). The fluvio-karstic system is treated as an open thermodynamic system (OTS), which exchanges water and heat with its surrounding. We propose to use a cylindrical PFM drained by a water saturated cylindrical CS, connected on one side to a sinkholes zone and, to the other, to a resurgence flowing at the base level of the karstic system. The overall structure of the model is based on the conceptual model of fluviokarst developed by White (2002, 2003). This framework allows us to develop the equations of energy and mass conservation for the different parts of the OTS. Two numerical models have been written to solve these equations: 1) the so-called AW (for Adiabatic Wall) configuration that assumes a conservative tracer behavior for temperature with no conductive heat transfer, neither in the liquid, nor in the PFM or even through the wall separating the CS from the PFM; and, 2), the CW (for Conductive Wall) configuration that takes into account the heat and mass transfers in water and from water to aquifer rocks both in the CS and in the PFM. Looking at the large variability of karstic system morphologic properties, dimensionless forms of the equations have been written for both AW and CW configurations. This approach allows us to gather the physical, hydrological and morphological properties of karstic systems into four dimensionless Peclet, Reynold, Prandtl and dimensionless diffusivity numbers. This formalism has been used to conduct a several orders of magnitude parametric exploration based on the Peclet and the Reynolds numbers. The final errors, between the AW and CW configurations, remain less than 1% over all of the parametric range. The combination of error curves bounds a closed volume in error space that gives a first upper bound of the error made by considering the temperature as a conservative tracer. Applying the method to an illustrative example of karst allows us to reach a first order error within a few degrees °C.

Introduction

Karst aquifers are often considered as potential solutions for meeting water needs required by agriculture, industry and human consumption. A highlight is the ability of these systems to return, during the dry season, the rainfalls of the watershed. However, exploitation by surface, or underground catchment, or pumping of these resources requires cautions to not jeopardize the balance of downstream ecological systems (Weber et al, 2006; Jemcov, 2007). However, exploitation of such resources by surface catchment, underground catchment or pumping requires careful evaluation in order not to jeopardize the balance of ecological systems downstream (Weber et al., 2006; Jemcov, 2007).

Karst environments present broad spectra of hydrodynamic properties which range from a Porous Fractured Matrix (PFM) to thin interconnected Conduit Systems (CS) whose diameters can range from fractions of a centimeter to tens of meters. This variability induces, for the discharge and recharge phenomena, temporal responses ranging from a few minutes to several months. In addition, except for the largest caves, it is generally not possible to enter the heart of the karst system to proceed at direct measurements of temperatures,
In general, external temperatures are subject to more or less rapid fluctuations depending on the sunlight, the daily cycle, the weather trend or season. Conversely, in depth, with the notable exception of phenomena consecutive to run-off or sudden rainstorms, the thermal fluctuations are damped in the CS and the PFM. Thus, most of the studies mentioned above made the assumption of rapid temperature fluctuation damping (Sinokrot and Stefan, 1993; Luetscher and Jeannin, 2004; Dogwiller and Wicks, 2005).

Our work aims to evaluate the influence of conservative temperature approximation based on the theoretical framework provided by the OTS. For this purpose, our theoretical development takes into account the different heat transfer possible between the different elements of the OTS. However, it is impossible for a “black box” like approach, to account for all of the natural karstic system complexity. While early models sometimes regarded karst as continuous media, more recent studies have shown the way for the inclusion of more complex internal structures (e.g., Covington et al., 2009; Luhmann et al., 2011; Covington et al., 2011; 2012). Following these works we have considered a cylindrical water saturated CS, carried by an abscissa axis x, separated from a cylindrical PFM by a permeable wall. This framework allows checking the mass and energy conservations solving water and rock temperatures in the different parts of the OTS with two different configurations: 1), a conservative tracer behavior for temperature in which the mixing between CS and PFM flows occurs in the CS without heat dissipation or dispersion in the CS, in the PFM or through the wall separating the CS from PFM (in the following this configuration is called AW for Adiabatic Wall); 2), the same dynamic of mixing between CS and PFM flows but with conductive heat dissipation within the CS in the PFM and through the wall that separates them (This configuration is called CW for Conductive Wall).

The overall morphological structure used for our study is that the conceptual model developed by White (2002 and 2003) for fluviokarst. This structure corresponds to the Cent-Fonts (Hérault, France) karstic system whose description is given in the Annex part and which serves as an illustrative example for a potential application of the method. In the White’s fluviokarst model (redrawn in Figure 1a) the stream that dug the valley is partially lost in a sinkholes area at the entry of the CS. However,
indicates that the water motion forms a zero divergence velocity field in the PFM and in the CS. In this study, we will also consider that the axial velocity, \( v_x \), is zero in the PFM and that the radial velocity, \( v_r \), is zero in the CS. Therefore, the mass conservation equation can be written in cylindrical coordinates \((x, r)\) as:

\[
div(\vec{v}) = \frac{\partial v_x}{\partial x} + \frac{\partial v_r}{\partial r} + \frac{u_r}{r} = 0
\]

By linearity, it is possible to apply the Ostrogradsky's or Green's theorem over each CS and PFM cylindrical section (of thickness \(dx\)) to convert the volume integral in a flux integral. The resulting Eq. 2 and Eq. 3 ensue for the PFM \((r \geq r_h)\) and for the CS \((r < r_h)\).

Furthermore, to describe the CW configuration, it is necessary to take into account the conservation of thermal energy during the conductive transfers in the CS, in the PFM and between water and the rocks of the aquifer. Following the work developed by Covington et al (2011), we apply a conduction-temperature advection equation in PFM, but adding a new dispersion-conduction term that takes into account the cooling effects of the diffuse infiltration. In the PFM for \((r \geq r_h, x \in [0, L])\), the energy conservation becomes Eq. 4:

\[
v_r \frac{\partial T}{\partial r} = D_M \left[ \frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial r^2} + \frac{2\partial T}{r \partial r} \right]
\]

and, in the CS, for \((r < r_h, x \in [0, L])\):

\[
v_r \frac{\partial T}{\partial x} = D_M \frac{\partial^2 T}{\partial x^2} + \frac{2\partial T}{r \partial r}(x, r_h)
\]

where \(v_r\), \(v_x\), and \(T\) are the velocity and temperature fields depending on the cylindrical coordinates \(x\) and \(r\). However, to simplify the presentation of Eq. 5, these dependencies are not explicitly written except if the values are taken at specific points as \(r = 0\) or \(r_h\).

In nature, karst systems exhibit great variability which is mainly due to the diversity of soils, to the extent of the watershed and the degree of karstification. CS present lengths ranging from tens of meters to a few tens of kilometers, while the permeability of the PFM,
or the hydraulic radius of CS cover several orders of magnitude. Only a non-dimensional approach, classical for fluid mechanics studies, can describe such broad ranges of local properties through dimensionless number approach. The morphological aspect ratio of the karstic system, \( r_h / L \), may display broader variations. However, hydrological and thermal properties of the various flows (\( T, Q \) for the intrusion at the sinkholes), but also the far field temperature and the rate of diffuse infiltration in the CS (\( T\infty \) and \( qM \)) also impact directly the temperature of the mixed water in the CS. Similarly, the physical parameters of rocks and water will also be gathered in these dimensionless groups as the viscosity or the ratio of water to rock thermal diffusivities.

Eqs. 1 to 5 have been rewritten using \( L \), the CS length, as length scale; the discharge difference between the resurgence and the intrusive flow at the sinkholes area (divided by \( \pi r_h^2 \), the CS wall surface) defines a natural velocity scale for diffuse infiltration \( V = (Q_s - Q_i) / \pi r_h^2 \). A time scale \( \tau = L / V \) ensues from these two scales and, with these conditions, the relationships between dimensionless parameters of length, speed and time are respectively \( x' = x / L, u' = u / V = L r_h / (Q_s - Q_i) \) and \( t' = t / \tau = t V / L \). The relationship that links the temperature to the dimensionless temperature is \( T' = (T - T_i) / (T_i - T\infty) \). Within this framework, with \( r = r_h / L \) and \( u = u / V \), the mass conservation equations in PFM and in CS become:

\[
\begin{align*}
\frac{\partial}{\partial t'} \left( \frac{r}{r_h} \rho' (x', t') \right) &= \frac{\partial}{\partial x'} \left( \frac{r}{r_h} \rho' (x', t') u' \right) + \frac{\partial}{\partial x'} \left( \frac{r}{r_h} \rho' (x', t') \right), \\
\frac{\partial}{\partial x'} \left( \frac{r}{r_h} \rho' (x', t') \right) &= \frac{\partial}{\partial x'} \left( \frac{r}{r_h} \rho' (x', t') \right) = \frac{\partial}{\partial x'} \left( \frac{r}{r_h} \rho' (x', t') \right).
\end{align*}
\]

and the energy equations in PFM and in CS become:

\[
\begin{align*}
\frac{\partial}{\partial t'} \left( \frac{r}{r_h} \rho' (x', t') \left( C_v + \frac{\eta}{\rho'} T' \right) \right) &= \frac{\partial}{\partial x'} \left( \frac{r}{r_h} \rho' (x', t') \left( C_v + \frac{\eta}{\rho'} T' \right) u' \right) + \frac{\partial}{\partial x'} \left( \frac{r}{r_h} \rho' (x', t') \left( C_v + \frac{\eta}{\rho'} T' \right) \right), \\
\frac{\partial}{\partial x'} \left( \frac{r}{r_h} \rho' (x', t') \left( C_v + \frac{\eta}{\rho'} T' \right) \right) &= \frac{\partial}{\partial x'} \left( \frac{r}{r_h} \rho' (x', t') \left( C_v + \frac{\eta}{\rho'} T' \right) \right) = \frac{\partial}{\partial x'} \left( \frac{r}{r_h} \rho' (x', t') \left( C_v + \frac{\eta}{\rho'} T' \right) \right).
\end{align*}
\]

Four groups of dimensionless numbers appear in Eqs. 6 to 9. They are combinations between the Peclet number, \( Pe = LV / D_w \), the dimensionless thermal diffusivity, \( D = D_m / D_w \), the Prandtl number, \( Pr = \nu / D_w \), and the tubular Reynolds number \( Red = 2 V r_h / \nu \).

The Prandtl number and the dimensionless thermal diffusivity depend only on the physical properties of water and rocks. In the following of this work, we will consider that the uncertainties on these parameters are significantly lower than those that might come from the variability of the hydrological. The latters mainly affect the Reynolds number (through the hydraulic radius of the CS) and the Peclet number (through the CS length). We therefore have built our parametric study on these two parameters which endorse most of the variability.

If the mass conservation equations (Eqs. 6 and 7) are not changed when the temperature is considered as a conservative tracer (AW case), this is not the case for the energy conservation equations (Eq. 8 and Eq. 9). In that case, it is necessary to consider the limit when the conductive heat dissipation disappears, that is to say when \( D_m \to 0 \). In the PFM, for \( (r \geq r_h) \), it comes:

\[
\lim_{D_m \to 0} Eq. 9 \Rightarrow \rho' u' T' \left\{ \frac{\partial}{\partial x'} T' \right\} = \frac{\partial}{\partial x'} \left[ \frac{u' T'}{r} \right] - \frac{2}{r} v' \left[ T' - T_{\infty} \right].
\]

According to the constant thermal boundary condition for the far field temperature (\( T'_{\infty} = -1 \)), the solution is given by Eq (11):

\[
\frac{\partial T'}{\partial r} = 0 \Rightarrow T' = -1
\]

It follows as an expectable result that, in the absence of heat conduction, the PFM temperature is uniform and equal to those of the far field. Now, let’s study the temperature equation in the CS for the AW case. It is then necessary to study the limit of Eq. 9 with zero thermal conductivity in water and rocks. In this case, if \( D_m \to 0, Pe \to \infty \). Factoring and simplifying the term Pe on both sides of Eq. 9 leads to:

\[
\lim_{Pe \to \infty} Eq. 9 \Rightarrow u' T' \left\{ \frac{\partial}{\partial x'} T' \right\} = 0
\]

That is to say:

\[
u' \left[ T' - T_{\infty} \right].
\]

Eq. 13 matches the classical expression of the thermal energy conservation for a mixing without heat dissipation.

**Numerical Modeling**

Two numerical programs were written to solve the conservation of energy in the CS and in the PFM for both AW and CW configurations. They are based on finite-difference second order accurate methods (Douglas and
Rachford, 1956). Both codes strictly comply with the same boundary conditions given by the flows entering the sinkholes, the rate of diffuse flow from PFM to CS and the intrusive and far field temperatures. The dimensionless temperature at the CS input is \( T'_{i} = 0 \), while the CS temperature at the CS resurgence remains free. The thermal boundary condition at the outer edges of the calculation grid (\( r=0.02 \)) is the far field dimensionless temperature, \( T'^{\infty}=-1 \).

For AW cases, a numerical 1-D code, solves Eq. 13 in the CS to calculates the temperature as a function of the abscissa \( x \). On an other hand, for the CW configurations, Alternate Direction Implicit (ADI) finite-difference methods are used to successively solve the temperature equations in the radial and axial directions. The equations were discretized on a grid of 100 points in the radial direction and 500 points in the longitudinal direction. A convergence test stops the iterative process when a steady state solution is achieved (in effect when the relative changes of the temperature are less than \( 10^{-4} \) for all the points of the computation grid). Continuity and coupling of the temperature between the CS and the PFM are ensured by sharing the thermal boundary conditions along the CS wall: the temperature calculated in the CS serves as radial boundary condition for solving the temperature in the PFM. This method allows coupling the resolution in both parts of the solution while taking into account the heat dispersion in the CS and heat spreads within the CS and between CS and PFM.

A first comparison of the AW and CW hypotheses was conducted by introducing the values of the morphological, hydrological, and thermal main data observed on the Cent-Fonts fluviokarst site in summer 2005 (see annex part, and Ladouche et al., 2005 for report). Thus, a CS length \( (L = 5,000 \text{ m}) \) is the distance, as the crow flies, between the sinkholes area of the Buèges stream and the Cent-Fonts resurgence near Hérault River (Figure 5); the hydraulic radius chosen for the CS \( (r_h = 5 \text{ m}) \) corresponds to the speleological situation in the cave near the resurgence (Figure 6); the flow rate at the sinkhole is \( Q_i = 0.055 \text{ m}^3/\text{s} \); the output stream at the resurgence is \( Q_s = 0.4 \text{ m}^3/\text{s} \); temperature at sinkhole is \( T_i = 295.25 ^\circ K \) (22.1 \(^\circ\)C); and far-field temperature is \( T^\infty = 285.35 ^\circ K \) (12.2 \(^\circ\)C) (Ladouche et al., 2005). Further, we assumed a water thermal diffusivity \( D_w = 1.43 \times 10^{-7} \text{ m}^2/\text{s} \) (Pechnig et al., 2007). These values lead to the following dimensionless numbers: \( Pe = 1.50 \times 10^8 \), \( Red = 4.29 \times 10^4 \), \( Pr = 6.99 \) and \( D_m / D_w = 9.93 \).

Figure 2 shows the temperature field obtained for the CW configuration in PFM and CS, after resolutions of Eqs. 8 and 9. The monotonic cooling with abscissa along the CS results from the mixing of the cold diffuse flow with the hot water that enters the CS at the sinkholes area. However, in the PFM around the CS, close to the sinkholes area, the heat transmitted by conduction in the PFM, induces thermal boundary layers that encounters the cooling flow from the far field. At a radial distance of a few tens of meters away from the CS wall, the thermal boundary layers attenuate rapidly and fall to a temperature close those of the far field.

On the other hand, solving Eq.13 provides the temperature of CS water with the conservative tracer hypothesis (AW). In Figure 3, the Comparison of the blue curve (AW configuration) with the red curve (CW configuration) shows that the CS cooling is overestimated by the conservative temperature hypothesis. Due to the

![Figure 2](image-url)
effect of heat conduction in the CS, in the PFM and from the CS to PFM, a significant amount of thermal energy is consumed in the PFM to hold the hot conductive boundary layers against the diffuse cold water advection from the far field. Both curves diverge quickly where the maximum of energy is required to maintain the higher boundary layers (abscissa close to the sinkholes area). For larger abscissa, the decrease of the boundary layer amplitude induces a stabilization of the temperature difference between the two AW and CW assumptions.

The final temperature difference at the CS output corresponds to the thermal energy used to maintain the equilibrium of water and rocks temperature in the PFM in spite of antagonistic effects between conductive and advective heat flows while the thermal diffusivity of the rock is about one order of magnitude higher than those of water. This final temperature difference bears a first-order information on the error due to the conservative tracer assumption in the PFM. Beyond this simple illustrative example, the next section will describe the results of a parametric study realized to reach a more general quantification of this error.

**Parametric Study**

The high variability of morphological, hydrological and thermal properties of karst systems stresses the need to conduct a parametric study to cover the wide ranges of dimensionless parameters that describe them. In this context, we conducted a systematic study of the temperature relative deviation (Eq. 14) obtained in the CS for both CW and AW configurations. It’s dimensionless expression is given by Eq. 15.

\[
\epsilon(x) = \frac{T_{CW}(x) - T_{AW}(x)}{T_{CW}(x)} \tag{14}
\]

\[
\epsilon'(x) = \frac{T_{CW}(x) - T_{AW}(x)}{T_{CW}(x) + T_{AW}(x)} \tag{15}
\]

As mentioned above we have restrained our attention in this section, to the study of the effects of Reynolds numbers and Peclet numbers changes. The Peclet number (\(Pe = LV / D_w\)) measures the ratio of advection to conduction characteristic times. Therefore, the higher \(Pe\), the preeminent effect of conduction compared to advection. Moreover, it directly characterizes the CS length, which is one of the most variable morphological properties of karst systems. The second entry of this study is the Reynolds number (\(Red=2Vr / υ\)), which is characteristic of a second highly variable morphological parameter of the karstic system : the CS hydraulic radius. These two quantities are mathematically linked through the Prandtl number \(Pr = (L Red Pr) / r_h\). However, it should be noted that in the energy equation (Eq. 9), \(Pe\) is present at the numerator of the Eq. 9’s terms while \(Red\) is present at the denominator. According to their increasing or decreasing values, they will induce opposite effects on these terms that characterize the relative importance of conduction, dispersion and advection.

To better understand the cross-influence of these two parameters we have, initially studied the behavior of the \(\epsilon'(x)\) depending on \(Pe\) (ranging from \(10^6\) to \(10^9\)) at constant \(Red(=2Vr / υ)\), the value previously obtained for the Cent-Fonts fluviokarst); then, in a second step, we studied the behavior of \(\epsilon'(x)\) by varying \(Red\) from \(10^3\) to \(10^7\) at constant \(Pe(=1.500 \times 10^8)\). The results of these two steps are shown respectively on the upper and lower parts of Figure 4. The left panels show the evolution \(\epsilon'(x)\) as a function of the abscissa x in the CS, while the right panels give an overview of the final errors at the system output.

In any case, at the sinkholes area location (\(x=0\)), the error is zero. This result is expected since at this particular stage, no energy transfer by conduction has occurred between hot intrusive water at sinkhole and cold intrusive water from PFM diffuse infiltration. Conversely, as soon as the offset x of the sinkhole increases, \(\epsilon'(x)\) increases especially as the boundary layers observed in Figure 2 are stronger. For higher abscissa x, the rate of variation of \(\epsilon'(x)\) decreases. In fact, two distinct types of behavior are displayed for \(\epsilon'(x)\) on the left panels of Figure 4. The firsts are characterized by curves that start from...
zero (for $x = 0$), reach maxima, then decrease more or less slowly for higher abscissa. The seconds shows monotonic growths of $\varepsilon'(x)$. The transitions between the two regimes are progressive and occur with the increases of the test parameters ($Pe$ or $Red$). However, it should be noted that, in all the cases, the error curves $\varepsilon'(x)$ remain well below 1%.

Let’s now examine in more detail how $\varepsilon'(x)$ evolves with $Pe$ (at constant $Red$) (Figure 4, top panels). According to the physical meaning of the Peclet number, its decreasing corresponds to a decrease of the relative importance of conductive transfers face to the advective ones, homogeneously in the whole system. In fact, for the lowest values of $Pe$ ($10^6$ and $5 \times 10^6$; Figure 4, top left panel, light brown and red curves) $\varepsilon'(x)$ increases monotonically with $x$. However, it converges uniformly to zero for further decreases of $Pe$. Conversely, for the highest values of $Pe$ ($5 \times 10^6$ and $10^7$, Figure 4, top left panel, dark blue and purple curves) $\varepsilon'(x)$ reaches maxima for abscissa close to the sinkhole area (low values of $x$). In these cases, the amplitudes of these maxima decrease with increasing $Pe$. $\varepsilon'(x)$ finally also tends to zero for all the value of the abscissa. In any cases, the maximum of the relative errors are achieved for intermediate values of both parameters. They remain less than 1% (the maximum values of 0.0092 and 0.0062 are respectively achieved for $Pe = 710^7$ and $Red = 10^7$). It is also interesting to emphasize that the results obtained for the illustrative example of Cent-Fonts fluviokarstic system naturally fall within this bounded error range.

**Summary and Discussion**

The purpose of this work is to try to assess a first order of the error done by considering temperature as a conservative tracer in fluvio-karstic systems. For that, we developed and solved the energy and mass conservation equations, leaned against the White’s conceptual model for fluviokarst, and within the theoretical background of OTS. We applied theses equations to a cylindrical CS, which receives hot intrusive water from a sinkholes area and, through a PFM, a cold diffuse flow from the far field all along its underground path. This set forms an OTS in which we studied two configurations. The first (AW) assumes that no conductive heat is lost in the CS water, neither between the PFM water and the aquifer embedding rocks; nor between CS and PFM through the permeable separation wall. The second (CW) takes into account the conductive heat dissipation in CS water, in PFM and the dispersion of heat by conduction through the PFM wall. These equations have been rescaled that leads to a new system of equations where four groups of dimensionless numbers measure the relative magnitudes of the various conductive and advective terms. Solving these equations in both configurations with strictly similar thermal and dynamic, boundary conditions allows assessing a first order of the errors induced by the conservative tracer assumption for temperature.

However, it is clear that our results lead only to a first-order information about this error because the method cannot completely eliminate or estimate other sources of error. Indeed, in order to proceed to the numerical solving of the mass and energy equations we need to consider laminar flows, in a saturated CS. Furthermore, we must keep in mind that the method is better applied during the recession periods when the hydraulic regime of the fluviokarst is as close as possible of a steady state. Further studies are needed, to go beyond the first results presented in this paper. However, some of our results seem encouraging since whereas the variability of karst systems has largely been accounted by the range extent of the parametric exploration, errors in the
The temperature assumption for particular karstic systems can be reversed by adjusting the scaling scheme. A simple calculation allows for quick retrieval of the error in the physical space, thanks to the morphological and hydrological properties of a particular fluviokarstic system. Comparing model results with field data provides the possibility for critical analyses and decision-making support for its applicability to local cases. Focusing on the illustrative example of the Cent-Fonts fluviokarstic system, Figure 4 shows that the relative error \( \varepsilon'(x) \) is limited to less than 1% (actually 0.0092 to 0.007). The error volume formed by the error curves, due to the variations of these two parameters, converges to zero for their extremal values. Meanwhile, the maximum of errors is reached for median parameter values that are characteristic of realistic karstic systems in both morphology and hydrological considerations.

From the results obtained in the non-dimensional space, it is possible to evaluate an upper bound of the first order of the error induced by the conservative approach.
reaches, 0.00613 at the exit of the resurgence in the dimensionless field \((Pe = 1.50 \times 10^4\) and \(Re = 4.29 = 10^3\)). When rescaled in the physical domain, the error \(e(x)\) indicates a temperature disparity \(T_{cw} - T_{aw} = 1.77 \text{ °C} \times 288.50\). It is clear that this information can be used to infer potential propagation of uncertainties in calculations or cooling rates. It can also be used to propose a calibration of the effects of the conservative temperature approximation depending on the equations and on the dimensionless properties of the karstic system (Machetel and Yuen, in preparation). In these next works, we will focus on examining that the theoretical evaluation of the error proposed in this work is compatible with the thermal data available on other karst systems. Indeed, at the sight of this study, it seems possible to use the results from a theoretical analysis to coerce information on internal thermal conditions of the karstic system. The results seem open interesting research opportunities that may be applicable to other systems whose workings are often described in terms of “black box”, as geothermal or hydrothermal systems.

**Appendix : Cent-Fonts resurgence (Hérault, France)**

The Cent-Fonts resurgence is the base level outlet of a fluviokarstic system, which watershed covers an area of 40-60 km\(^2\) (Figure 5). This basin is located north of Montpellier, on the right bank of the Hérault River within a thick dolomitic Middle and Late Jurassic limestone sequence.

Several structural, geological, geochemical and hydrological studies were devoted to this karstic system for several decades (Paloc, 1967; Camus, 1997; Petelet et al., 1998; Schoen et al., 1999; Ladouche et al., 2002; Petelet-Giraud, 2003; Aquilina et al., 2005; Ladouche et al., 2005; Aquilina et al., 2006, Marechal et al., 2008; Dörfliger et al., 2009). The watershed is bounded to the north and northeast by the Cevennes fault, the surface course of the intermittent Buèges stream, and southeast by the Hérault River, which drains its base level. The watershed encompasses the upper course of the Buèges stream, which flows on an impermeable Triassic outcrop until it reaches a sinkholes area crossing a batonian dolomitic area a few kilometers downstream from Saint-Jean-de-Buèges (Figure 5). From that point, the Buèges stream surface course forms a valley mostly dried up that joins the Hérault River a few kilometers upstream of the confluence with the Lamalou. On the other hand, the underground pathway from Buèges sinkhole to the Cent Fonts resurgence was established by tracing (Dubois, 1962; Schoen et al., 1999).

**Figure 5.** Cent-Fonts fluviokarst watershed. Redrawn on the Hérault geological map 1/50000 France BRGM (J2: Bajocian; J3-5: Bathonian, Callovian, Oxfordian; J6: Kimmeridgian; J7: Tithonian). The Cent-Fonts fluviokarst watershed covers about 60 km\(^2\) including 10 km\(^2\) for the Buèges resurgence watershed (Schoen et al, 1999).

**Figure 6.** CS map of the Cent-Fonts resurgence. Unrolled 3-D speleological map of the Cent-Fonts CS near the resurgence. Vertical distances (m) are conserved. Depths below the main cave entry are given in m (italic). During the summer 2005 pumping tests, piezometric heads and temperatures have been measured in the so-called « CGE », « Reco » and « F3 » boreholes. Temperature and discharges were also recorded at the output of the pumping device located in the «F3» borehole, at the entry of the Buèges Stream swallow zone and in the Hérault River near the resurgence (Ladouche et al., 2005).
The Cent-Fonts karstic system takes root in a dolomite layer Bathonian of 150 to 300 m thick and probably extends into the layer of Aalénien - Bajocian underlying. It forms a fluviokarst by catching the Buèges Stream loss at the Sint-Jean-de-Buèges sinkhole and by draining the watershed rainfalls which percolate through an upper Jurassic epikarstic layer (Petelet-Giraud et al., 2000). Therefore, the Cent-Fonts karstic system is similar to the conceptual model of White (Figure 1a) with a CS collecting sinkholes losses and a diffuse infiltration flowing from a PFM into the CS. After 5 km of underground path (as the crow flies), the CS poors through the Cent-Fonts resurgence in the Hérault River which drains the base level of the karstic system. The resurgence discharges through a shallow network of springs that flow a few tens of centimeters above the Hérault (Schoen et al., 1999). During the dry season, the resurgence discharge ranges from 0.250 to 0.340 m$^3$/s summer (Maréchal et al., 2008). The detailed structure of the CS near the resurgence output (Figure 6) has been explored by divers (Vasseur, 1993). In the mapped area, the CS cross section ranges between 4 to 16 m$^2$ (Dörfliger et al., 2009). The Cent-Fonts resurgence has undergone numerous field observations since 1997. Several years of flow measurements have been recorded to calibrate the base flow of the Cent-Fonts resurgence, of the Buèges stream and the losses in the sinkhole area. A pumping test campaign was conducted during the summer of 2005 by BRGM under contracting authority of Conseil Général de l'Hérault. This campaign provides to scientists many temperature and flow records from surface and deep holes measurements (Ladouche et al., 2005).

### Acknowledgments

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### Table

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<th>Acronyms</th>
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<td>Adiabatic Wall</td>
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<td>OTS</td>
<td>Open Thermodynamic System</td>
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<td>PFM</td>
<td>Porous Fractured Matrix</td>
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CONCEPTS FOR GEOTECHNICAL INVESTIGATION IN KARST

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Abstract
There seems to be a lack of recognition in the literature that addresses the variety of karst in the United States of America and some of its offshore territories. For example, there are the well-known solutioned carbonates of Florida and the Caribbean, but there are also the somewhat older, harder carbonates of St. Croix, U.S.V.I. Even Florida’s recently deposited karst varies from region to region. There are also the ancient, flat-lying carbonates of the interior craton that often have semi-horizontal cavities resulting from variations in ground water levels affecting bedding and the contorted rocks of the Appalachians with its apparently chaotic variations in solutioning found across-strike and in relation to folds, faults, and fracturing. In addition, there are various salt and gypsum deposits in the south and southwest that pose their own problems to man’s works.

As the geology differs, so does, to some extent, the investigation requirements, investigation techniques and engineering solutions. There is no single set of investigative tools that fit all karst sites. Geophysical investigations are apparently far less suitable for the broken and twisted Appalachian karst than in the flat-lying mid-continent carbonates or the less contorted “karst” of Florida.

Specific procedures developed for geotechnical investigation in true karst have been documented for many years now. However, it appears that many practitioners are not aware of them or choose not to use them because of the possibility of increased costs; or too often, a lack of geotechnical understanding of the work of others in karstic areas outside their sphere of experience.

This paper will attempt to provide a rational geotechnical approach to carbonate rock investigations in the United States while recognizing the inherent variabilities of the targets and the economics of pre-construction investigations; with the understanding that one size does not fit all.

Karst Variability
Carbonate bedrock is found throughout the world. Spectacular examples of true, pinnacled karst are found in China and the namesake plateau in Slovenia and Italy. However, all carbonates are not the same, although this distinction is often overlooked by investigators. They range in strength and character from the soft offshore corals of the Caribbean Islands and Florida to marbles. More confusion is added by having the older, hard, but flat-lying carbonates of the central US, the hard, stressed, folded and fractured limestone, dolomites and marbles of the Appalachian Mountain chain and the soluble gypsum (evaporite) of the southwestern US (which can form sinkholes that dwarf the well-publicized ones of Florida).

As can be deduced from the extensive portion of the US underlain by karst (Figure 1), development growth has likely forced administrators, politicians, the public, as well as engineers and geologists to recognize the concerns of building atop karst. The result has sometimes been environmentally aware regulation and better technical understanding to address the problems posed by this variable and generally disguised environment. We can no longer conscientiously drill three or four test borings to characterize a 500-acre site for construction or use one boring per mile to address the engineering concerns along a roadway or transmission line in karst and assume that we have all the information necessary for evaluation and design of structures.

Geotechnical analyses and recommendations are not the same for all conventional (non-karst) sites, but they are even more varied and complicated for karst sites. In formulating a site study/evaluation, one must appreciate; A) The potential variations in physical properties across and below a site, B) the applicability and appropriateness of the available suite of site investigative tools, and C) the availability or lack of potential planning and/or engineering solutions to cover the uncertainties that will likely exist at the karst site in question.
The intention of this paper is to point out the difficulties of performing geotechnical investigation in karst as a result of the differences in bedrock ages, degree of deformation and perhaps most important, the degree of tectonism experienced by the bedrock in different regions. Not all karst is the same; not all the exploratory tools used or the manners in which geotechnical investigations are performed should be the same for all types of karst. It should be noted that our experience has been in “limestone” (CaCO$_3$), dolomite (CaMg(CO$_3$)$_2$) and/or marble, we will allow others to comment on gypsum (CaSO$_4$$\cdot$2H$_2$O) and other evaporites.

Another aspect that must be considered is the existence of local or State ordinances regulating either the construction or impact allowed at karst sites. These regulations, where they exist, can have different intentions. For example, many municipal “limestone” ordinances in Pennsylvania are primarily directed toward inhibiting development; Virginia’s toward protecting ground water; Michigan’s toward control of feed lot expansion; and New Jersey’s generally toward limiting construction on karst or providing a means of development of the site conscientiously, not necessarily economically.

**Geology**

For simplicity, this paper will attempt to crudely divide this presentation into three groups of karst.

- Old, mid-continent, generally flat-lying carbonates,
- Old, folded and faulted Appalachian carbonates, and
- Recent, coralline limestone.

For those interested in a more precise division, we suggest the United States Geological Survey (USGS) compilation Characterizing Regional Karst Types under the Framework of the New National Karst Map (Weary et al., 2008). Generically, the differences are age and degree of tectonism. The USGS further differentiates US karst types by thickness of the overburden and precipitation. The overburden thickness of concern to the geotechnical community are generally less than
100 feet; although mining, some dams and similar large construction can be exceptions. The recent New National Karst Map (Figure 1) also includes the areas underlain by evaporite karst, which are not covered in this paper.

All of the aforementioned karst types of concern were originally deposited in warm, relatively shallow seas. Deposition or coral growths continue today in the warm waters of the Atlantic Ocean and Caribbean Sea. The older carbonates (Cambro-Ordovician-aged) have been subjected to a variety of stresses and, in the Appalachian Mountains, a series of orogenies resulting in faulting, folding and fracturing (and some metamorphism) along what was the Atlantic coast approximately 300+ million years ago.

The mid-continent carbonates are generally the same age as the Appalachian rocks, but have not been subjected to the same orogenic forces, but did experience some of the stress fields caused by several openings and closing of the proto-Atlantic Ocean.

**Geologic Concerns**

The occurrence of sinkholes (dolines) swallowing buildings, automobiles, farm equipment and people has been well publicized. Less well recognized are the settlement of structures (including dams) and sinkhole occurrences in roadways, in backyards, below swimming pools, farms, manure storage facilities, railroad structures, fuels storage areas and bridge abutments. Less recognized hazards exist such as creating flooding and compromising stormwater detention/retention/infiltration systems, thus allowing contaminants to reach ground water supplies.

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**Figure 2.** Representation of the simplistic divisions of United States karst (Section A from Schmertmann and Henry, 1992).
The irregularity of the Appalachian, as well as some recent, softer Florida-type bedrock surfaces presents additional concerns in its effect upon the ground surface, further complicating geotechnical evaluations (Figure 2). Long and sometimes sinuous conduits are common in the central US’s flat karst (e.g., Mammoth Cave). Both lateral and vertical variations in the overburden materials (thickness and properties) are more common in the Appalachians. After a visit to almost any commercial cave, one can visualize the effects that changing ground water conditions over time have had on the bedrock. These differences in rock type, age of deposition and the range of effect from tectonism have to be considered in geotechnical investigation and for potential remedial solutions in these various environments.

Geotechnical Evaluation
The first step in developing a program of investigation is to expand upon the knowledge of the geology in the area or site of concern. State and Federal agencies generally have a wealth of information concerning subsurface conditions. These data can include:

1. Bedrock types and their expected depth below grade.
2. Driller’s logs and well yields from specific locations or geologic formations. These data can provide clues to the degree of fracturing and or solutioning found within various formations. However, we must be aware of driller’s classifications such as “gray rock”.
3. The existence and density of caves, sinkholes and disappearing streams.
4. The existence of known or suspected faulting, antiforms and synforms (i.e., where the bedrock has been stressed and subjected to deformation) where increased solutioning is likely to be experienced.
5. Bedrock strengths and quality of overburden.
6. Textural classifications that can provide a clue to material solubility.

Be aware that similar rocks and soil types do not necessarily have the same formation names from State to State.

Geotechnical Engineering
Sinkholes are an obvious concern in areas underlain by carbonate bedrock. Pictures of huge holes in the ground swallowing cars and houses make for big news. However, much of the older rocks can be quite hard when protected from weathering. For example Mammoth Cave in Kentucky and the Natural Bridge in Virginia.

Considering the expected lifetime of many construction projects, founding on these materials can still be an appropriate approach. The unconfined compressive strength of Cambro-Ordovician limestones and dolomites can be on the order of 10,000 to 15,000 pounds per square inch. As a result, the roof over a cavity can sometimes support large loads; though if it fails it will likely be well-documented.

However, even the relatively weak, recent corals of Florida and the Caribbean can support significant loads where not compromised by solutioning. Although the Schmertmann and Henry representation of Figure 2 (Section A) may somewhat exaggerate concerns at a Florida site, the concerns remain.

There are a number of engineering solutions to founding structures, roads, utilities, detention/retention basins and tunneling in karst. The basic problem is evaluating the subsurface conditions satisfactorily and to define the solution in a reasonably economic manner. So, it becomes somewhat of a balancing act. There is a need to find suitable materials to carry the proposed loads during the economic life of the structure. The problem is more complicated than it would be at most non-karst sites. One of the problems of founding on Appalachian karst over a more conventional site is compounded by the variability, both laterally and vertically, of the seams and fractures, and the general subsurface conditions. The material properties of these contorted rocks can vary significantly over short distances.
filtration in overburden soils and increasing the speed that contaminant can reach a receptor.

**Preliminary Site Evaluation**

In any “limestone” investigation, the “best bang for the buck” is usually the results of the initial stages of a site study. The first step is a review of any available data from federal and state sources including environmental reports and studies performed for nearby sites.

Aerial photos of the site are, whether from aircraft or satellites, highly valuable and often available from archives. A series of aerial photos taken over time can show changes in vegetation, landforms, farming practices, etc. For example, why is a tree standing alone in the middle of a cultivated field? Karst features can develop over time and then later masked by farming practices. Aerial photos taken in the early spring (before tree cover) and during wet years (e.g., Figure 3) can show changes in moisture that can be quite telling. Persistent linear and circular features are particularly suspicious if noted in photos taken over time or on LiDAR. Even drought-induced “crop lines” have been used as a tool in delineating potential sinkhole locations (Panno et al, 2013). These features should be further investigated by a site reconnaissance to help in identifying any noted features for use in a subsurface model.

In the past, sinkholes had many uses, garbage receptacles, debris pits and even as unrealized flood control aids. Areas that have been mined or quarried can indicate mineralized zones that can have an increased susceptibility to solutioning. In Appalachian karst the ground elevation variations can be more severe than observed from a windshield reconnaissance. Standing on a high point overlooking a site or flying over a large site at low altitude can be very informative, particularly within more flat-lying areas. We have observed sites as pockmarked with sinkholes as a World War I battlefield.

In addition, the construction process itself can create unstable or weakened conditions. Often, there is poor control of surface and ground water during construction of a facility. Excavation at a site can remove a protective layer of low permeability soils over solutioned rock, and pond water in these compromised areas. Ground water can travel along the top of the bedrock surface until it finds an entrance into a cavity, eroding soils from directly atop the bedrock and increasing the area of concern. Changing the hydraulic head and/or flow rate at a construction site, either by cut or fill, might alter otherwise stable conditions. The effects of changing the hydraulic conditions at karst sites are exemplified by the failure or remediation of many dams built atop karst. Also, the potential for ground water contamination is much higher in karstic environment by decreasing

Florida and mid-continent karst concerns can be caused by the lateral and vertical movement of ground water, with the effects being greater over shorter periods in the softer Florida and Caribbean karst.

Can the investigator sample enough locations on the site, either by direct or indirect means, to provide an appropriate support solution for such a variable subsurface? Imagine the difficulties of investigating a site such as the one shown on Figure 3 for the development of a satisfactory model of the subsurface. The many variables related to different karst areas of the US make it virtually impossible to answer all site related questions definitively before the start of construction. Hence, any construction-related planning should consider contingencies for increased costs for both inspection and the possibility of additional or supplemental remediation.

In addition, the construction process itself can create unstable or weakened conditions. Often, there is poor control of surface and ground water during construction of a facility. Excavation at a site can remove a protective layer of low permeability soils over solutioned rock, and pond water in these compromised areas. Ground water can travel along the top of the bedrock surface until it finds an entrance into a cavity, eroding soils from directly atop the bedrock and increasing the area of concern. Changing the hydraulic head and/or flow rate at a construction site, either by cut or fill, might alter otherwise stable conditions. The effects of changing the hydraulic conditions at karst sites are exemplified by the failure or remediation of many dams built atop karst. Also, the potential for ground water contamination is much higher in karstic environment by decreasing
Site/Route Investigation

If the site or selected route seems economically and technically viable, then it is likely that some additional geotechnical study will be required. Obviously, the nature of the project will change the character of the investigation. It is often advisable to phase future investigations.

The information most desired at any karst site is the distribution and dimension of soil voids and bedrock cavities, and whether those cavities or voids are filled with water or soil. Also necessary is some knowledge of the bedrock surface variations. The expected variations in the bedrock surface will differ largely from relatively flat mid-continent rocks to Florida’s variable and generally soft carbonates and the even more erratic, hard Appalachian rocks.

Another consideration is what effect the planned construction would have on local ground water supplies, which can be influenced even from considerable distances in a water-filled cave system. Dye studies performed by knowledgeable and aware professionals seems to be the only way of assessing possible ground water concerns and travel paths that new construction atop karst can effect. Exploring caves (spelunking or diving) by the more adventurous investigators can be very useful, although dangerous.

A host of geophysical procedures have been espoused as an effective investigative tool. These techniques include seismic reflection, refraction and tomography, electrical conductivity and resistivity, self-potential, ground penetrating radar, gravity, and Spectral Analysis of Surface Wave (SASW) methods. Apparently, the best results in the use of geophysics at karst sites have been a combination of geophysical procedures coupled with test borings (Benson, et al, 1998).

The efficacy of investigating with a single geophysical tool, using air-track probes and test borings to calibrate the results in Appalachian karst, is unfortunately exemplified by the following statements by one geophysicist in his report of a resistivity survey: “Generally, resistivity data is very good with good repeatability and trends that correlate well from intersecting and adjacent survey lines.” Followed later by: “The results of the survey show several different subsurface conditions. Detected by the survey are possible sinkholes and possible depth to bedrock variations. Differences in results from the drilling program may be caused by geologic differences affecting the electrical properties of the subsurface materials, modeling parameters, and orientation of the electrode array.” Apparently, even the geophysicist could not correlate the survey results into a coherent model in Appalachian karst. This geophysical investigation did not seem to be a useful tool for characterizing this site’s subsurface conditions in preparation for any kind of development.

Thus, it appears that geophysical procedures alone will not yield the answers to many karst concerns and can, at best, be interpreted with the aid of test borings and probes drilled by experienced, cooperative personnel under the technical direction of experienced field personnel including geophysicists.

Test pits can be performed in a conventional manner and can be very informative if portions of the rock surface can be exposed. Potential signs of solutioning can be deduced from near surface-effects. Is the weathered bedrock relatively uniform and straightforward or is there evidence of leaching or groundwater movement? Are the remains of an old, filled sinkhole obvious in the pit walls or bottom (Figure 4)? Is relict bedding distinguishable in the pit wall?

Test borings should be drilled using rotary-wash techniques without the use of drilling mud whenever possible so that drilling water loss depths and quantities can be monitored. In clean, sandy soils this may not be possible, but drilling with augers and periodically introducing water (say between samples) or the use of a light mud where necessary...
should allow for drilling fluid losses to be monitored while keeping the boring open.

Drilling water lost at the top of the rock usually indicates a down-gradient channel or the gradual erosion of soil into open bedrock solutioned passages. Soft soil zones are often observed atop the bedrock surface or adjacent to pinnacles. Conventional soil sampling techniques are generally adequate. Although providing water to the drilling site can become a logistical problem, it can be mitigated by casing off zones of significant water loss.

Encountering a karstic bedrock surface can be quite eventful. Carbonates have many faces; will it be sound, weathered, broken, a bedrock pinnacle or an erratic boulder, or saprolite below sound rock? Coring most carbonate rock is best done using double- or triple-tube, split core barrels. At least one spare core barrel should be on hand as the variable conditions that can be encountered are often hungry for drilling equipment utilized less than cautiously.

The information that can be observed from cores derived from a split core barrel is far more representative of the actual bedrock conditions and well worth the increased expense of its use. Fracture frequency and orientation is more easily observed and fracture and cavity filling is often captured in the barrel, along with highly weathered zones; all of which could be lost or minimized by hammering the core from a non-split barrel. Again, experienced drillers and competent inspectors are essential.

**Foundation Design Considerations**

Most foundation solutions are available for use once the scope of the subsurface concerns is recognized. The most commonly used sinkhole stabilization solutions are; A) excavate to sound rock and backfill to building grade, B) transfer construction loads to sound rock or bypassing the area of concern, C) densifying overburden materials and D) grouting of cavities with non-shrinking materials. Whatever concept is chosen, the execution should be flexible and hopefully cost-effective.

**A) Excavate to Sound Rock and Backfill**

The simplest is excavation to sound materials and returning the area to grade with compacted fill (sometimes after dental grouting of bedrock openings) or even lean concrete if the excavation is shallow enough. The unfortunate part of such a program is the need for enough quality subsurface information to be able to generate accurate excavation and backfilling costs. A pinnacled bedrock surface makes this very difficult.

For example, the Maryland State Highway Administration (MDSHA) responded to a large sinkhole immediately adjacent to a major interstate highway. The sinkhole was 110 feet long by 30 to 35 feet wide and 35 feet deep at the throat. In an effort to keep the highway embankment stable, the sinkhole was quickly and partially backfilled with some 2,700 cubic yards (cy) of quarry waste. Unfortunately, while drilling to place grout, the rock surface was revealed to be quite variable, necessitating an additional 2,045 cy of grout to fill the subsurface cavity. The maximum bedrock depth encountered was 100 feet (Martin, 2004). Even with the local geologic information available from highway construction and local quarry operators, they could not anticipate the extent of the weakened or missing subsurface materials.

**B) Transfer Construction Loads to Sound Rock**

Transferring loads from weakened subsurface areas to those capable of supporting loads is another founding alternative in karst. Bridging openings in the bedrock surface or soft soils zones with a reinforced concrete pad have been used.

A common foundation solution can be the use of driven piles or caissons, particularly with the present day ability to drill through the pile or caisson shaft in order to evaluate the quality of the founding materials below the pile tip and to possibly introduce grout if conditions warrant.

Pin piles have often been used satisfactorily because the pre-drilling used for their installation allows an increase in knowledge of the subsurface conditions at the pile location. However, grouting the pin piles to bond them to the sides of the hole can require large amounts of grout and there can remain unsupported lengths through cavities and soft soil zones.

Mathematical models have been or can be developed to assess the load-carrying ability of cave or sinkhole roofs in order to provide a requisite number and type of deep foundations to be used. However, it is probably more economical to perform a one-time investigation/foundation solution such as drilling and grouting, or installing a pile foundation to resistance depths, then drilling through the pile...
There are many hundreds of square miles of the United States underlain by karstic soils and bedrock. Unfortunately, all karst is not the same, though this has not been as well recognized as it should. The media has enjoyed reporting on sensational sinkhole occurrences that have swallowed houses, cars and people, but even if an “expert” has been contacted, the geotechnical concerns that may have existed are treated in passing.

Geotechnical practitioners must be more communicative when dealing with property/facility owners, planners and engineers. Clients and designers should be made aware of the possible dangers lurking below as well the impact the karstic subsurface conditions could possibly have on their plans.

It is difficult to understand why more municipalities do not have appropriate “limestone” ordinances as some have been tested in court and proven legal. The same can be said about the geotechnical and structural design professions; few appear to understand the difficulties that can result from the existence of carbonate bedrock below a site. Obviously, experienced consulting is necessary to ensure sound construction in karst and in the development of “limestone” ordinances. These ordinances should be directed toward the varying conditions of an individual karst site, as well as the differences in karst from region to region as discussed herein.

As much data is available to the geotechnical engineer prior to planning a subsurface investigation, this should be utilized to its fullest in considering the target. Determining an appropriate investigation program for a karst site is dependant upon this knowledge, which can only be utilized to its fullest with experience.

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Sinkhole is a ground surface depression that occurs with or without any surface indication. Sinkholes commonly occur in a very distinctive terrain called karst terrain. This terrain mainly has a bedrock of a carbonate rocks such as limestone, dolomite, or gypsum. Sinkholes develop when the carbonate bedrocks are subjected to dissolution with time to form cracks, conduits, and cavities in the underground bedrock. These features allow the overburden soils (on top of the carbonate bedrock) to transport through them to the underground cavities, which results in surface collapse due to the upward progression of the soil cavity toward the ground surface. Sinkholes vary in shapes and sizes. They have different shapes such as inverted cone, shallow bowl, and shaft shapes. Also, they can range from less than a meter to hundreds of acres and from 30 cm to 30 meters in depth (Waltham et al., 2005).

Sinkholes can be formed due to several processes such as bedrock dissolution, soil suffosion, rock collapse, and soil collapse. Based on the formation processes, sinkhole generally can be classified to six types: Solution (Dissolution) sinkholes, Collapse sinkholes, Caprock sinkholes, Dropout (Cover-collapse) sinkholes, Suffosion (Cover-subsidence) sinkholes, and Buried sinkholes (Lowe and Waltham, 2002) (Williams, 2004) (Waltham et al., 2005).

The state of Florida is one of the most susceptible places to sinkholes in the United States due to its geology. Florida’s karst geology is underlain by carbonate deposits, which is continuously subjected to a dissolution
process due to the circulation of the groundwater (Atkinson, 1977) (Quinlan et al., 1993) (Tihansky, 1999). The dissoluble carbonate bedrock is overlain by several layers of sand and clay soils. These clay and sand sediments vary in thicknesses based on their location within Florida (Bottrell et al., 1991). Florida’s sinkholes are mainly classified into three types: dissolution sinkholes, cover-subsidence sinkholes, and cover-collapse sinkholes. All of these types are the results of one or both of the dissolution and suffosion processes. The dissolution process is the chemical process where the carbonate rock dissolves due to the exposure to acidic water forming cracks, fissures, conduits, and cavities in karst. While, suffosion is a physical process of transporting the unconsolidated soil sediments to the bedrock’s underground cavities through the existing cracks and conduits (Sinclair and Stewart, 1985) (Tihansky, 1999).

Florida’s climate has a very distinctive two seasons (dry and raining seasons). The groundwater reaches its highest level in the end of the raining season (September). However, this level decreases until it reaches its lowest level at the end of the dry season (May). This kind of groundwater seasonal variation is one of the most important factor that triggers sinkhole collapses in Florida (Lewelling et al., 1998) (Sinclair, 1986) (Tihansky, 1999).

**Problem Statement**

Sinkhole prediction is a complex task due to the combination of different factors (geological and hydrological factors) involve in forming sinkholes. There is a broad field of the ground investigation techniques that can be used to investigate possible sinkhole locations. These techniques can be direct investigation by using soil probing, poring, drilling and sampling, or indirect investigation by using either geophysical methods or aerial or satellite remote sensing. The problem with the direct methods is that the borehole can easily miss a progressing underground cavity. Besides that, sinkhole history maps, and aerial and satellite remote sensing are not providing assurance that all the surface depressions (subsidence) detected by these methods are actually sinkholes (Waltham et al., 2005). No single method works in all situations, and an integrated approach must be adopted. As a part of this integrated approach, we studied the relationship between groundwater levels and sinkhole collapse.

**Previous Work**

A discussion on previous research on sinkhole soil models is presented in this section. In the past, some models were implemented using different approaches such as centrifuge models, analogical models, and actual soil physical models (Abdulla and Goodings, 1996) (Goodings and Abdulla, 2002) (Chen and Beck, 1989) (Caudron et al., 2006a, 2006b) (Caudron et al., 2008) (Lei et al., 1994) (Lei et al., 2005).

In 1989, Chen and Beck designed a two dimensional soil model to study the mechanisms of sinkholes. They used layers of natural sediments, which were tested in a parallel-plate type tank with a bottom opening. This tank has wooden bottom and Plexiglass sides. Chen and Beck (1989) simulated 23 different trials of homogeneous and stratified soils with initial conditions of dry, partially

**Research Scope**

The motivation behind the present research was to find a ‘sign’ to guide the ground investigation team to the potential hazardous area of sinkholes based on existing information such as groundwater levels. Since groundwater change is one of the main driving forces to cause and accelerate sinkholes in Florida, it is anticipated that the indication of the sinkhole collapses may be noted in the groundwater behavior before the surface collapse occurs. Hence, a small-scale physical model was designed and built to naturally simulate sinkholes. This model is a spatial-temporal model type. It was mainly designed to monitor the groundwater fluctuations around a predetermined sinkhole. The monitoring wells were radially distributed around the sinkhole in the physical model.

The model was initially designed based on a typical profile of Florida’s karst hydrology and geology. An important assumption in this test was that the dissolution process has taken place previously. In this model, the dissolution fracture is represented by a circular hole that transports a certain volume of soil through the limestone to an underground cavity. Moreover, this spatial-temporal model was designed to simulate a period of time at the end of the dry season in Florida (May), where the groundwater drops to its lowest levels. In general, the model is used to study the relationship between the groundwater fluctuation and sinkholes’ formation, location, and time.
saturated, or saturated. This simple model was designed to simulate a cover-collapse sinkhole. The objective of this study was to obtain some data about the sinkhole’s mechanical processes which were not known at that time. In this model, the authors found that type of the sediments, namely sand or clay, controls the time of the collapse. Also, the initial conditions of the sediments, such as dry, saturated, or partially saturated varies the speed of the sinkhole development. The model also proved that in the stratified overburden, the collapse may stop when a cohesive stratum is encountered at the top of the opening. This will cease the internal erosion either permanently or temporarily. While this qualitative two-dimensional soil model is a very simple model, however it can provide some basis for more sophisticated quantitative physical models of sinkhole to be developed (Chen and Beck, 1989).

Finally, a large-scale experimental study of sinkhole physical models was conducted by the Institute of Karst Geology in China (CAGS) in 1997. The model was aimed at studying the factors that control the formation of a sinkhole (Lei et al., 1994, 2002, 2005). CAGS’s physical model consists of three main components that are a base unit, recharge-discharge system, and observational system. It is a large-scale model with dimensions of 3 m in height, and 2 m in both depth and width (Lei et al., 1994, 2005).

Next, Lei and others, in 2005, simulated certain sinkhole formations in Hongshan District by using two conceptual models. This study investigated the effects of the width of limestone cracks, rate of water pumping, and mudstone thickness (the mudstone layer is located on the top of the limestone). It was concluded that groundwater pumping triggers more sinkhole collapses. In addition, the cracks in the limestone have a direct relation to the voids in the soil sediments in terms of size. Finally, it was noticed that the rate of the declination of groundwater is an important factor in the sinkhole collapse (Lei et al., 2005).

**Sinkhole Evaluation Based on Groundwater Recharge**

In 1994, Foshee and Bixler conducted a study of cover-subsidence sinkholes in Florida. The development of sinkholes around State Road 434 and Harbor Isle intersection in Seminole County, Florida, caused minor pavement settlement for that intersection. Seven different sinkholes occurred north and south of State Road 434. These sinkholes also caused settlements to building, roads, and yards. Hence, the Florida Department of Transportation (FDOT) decided to monitor the pavement settlement for State Road 434 to evaluate potential causes. A subsoil explorations program was conducted by using several cone penetrometer tests and the installation of permanent piezometers. The data evaluation of this study showed that there was a layer of very loose soils located at deeper ground strata. This loose soil was subjected to internal soil erosion (raveling). This raveling soil migrates slowly through limestone cracks to underground cavities and conduits in the carbonate bedrock. Eventually, this raveling process ends with a surface depression called cover-subsidence sinkhole. The main driving force of this raveling process is the downward groundwater movement, which is called recharge. This recharge occurs because of the difference in the shallow water table and the confining aquifer water level if recharge points exist which are the bedrock cracks. Recharge was observed in this site by studying the piezometer reading for almost two years. However, in this study, only piezometer readings at a specific time intervals were plotted as contour maps. The piezometer head contours showed a very clear depression indicating the settlement location. Foshee and Bixler (1994) stated that studying sinkholes by the pore-pressure-contouring technique should be further investigated to validate the reliability of this technique in different types of subsurface soil conditions.

**Current Sinkhole Physical Model**

This current study’s main objective is to conduct a spatial-temporal analysis for network of groundwater monitoring wells to try and predict the location of a sinkhole collapse. In reverse analysis, a network of wells were distributed in a radial distances around a predetermined sinkhole location. Sensor devices were chosen and programmed to detect the water level fluctuations with a high degree of accuracy. The water level was monitored at 0.5 mm resolution. The data was also collected at a high sampling rate of 100 Hz. Due to the lack of initial research funds, a simple 55-gallon metal drum to be used (56 cm diameter) for testing. A 5 mm circular hole was drilled at center of the base of the drum. This hole represented a crack or a collection of close cracks in the limestone bedrock. The purpose of this circular hole was to transfer a certain volume of soil sediment out of the model to mimic the loss of soil.
The sinkhole simulator included a network of eight groundwater monitoring wells. These monitoring wells were distributed in a radial manner around the center, which was the predetermined location of the eventual sinkhole. Figure 1 shows the radial distribution of the eight monitoring wells. Each monitoring well was made of a one-inch PVC pipe. These pipes were perforated all around to allow the water to enter. The pipes were then wrapped with a geotextile fabric to allow only the water to pass and filter the soil particles. Eventually, every PVC pipe (well) was equipped with a 12 Inch eTape Liquid Level Sensor (MILONE Technology). The PVC pipes were also used to maintain the sensors in vertical orientation during the test to achieve the highest accuracy of their results. The sensors were used to read the actual water levels at the eight monitoring wells. The locations of the monitoring wells were set to be at the following distances (10 cm, 12 cm, 14 cm, 16 cm, 18 cm, 20 cm, 22 cm, and 26 cm) from the center of the test as shown in Figure 1. These locations were chosen based on a series of tests to make sure that they are far enough from the sinkhole failure zone. This assures that the closest pipes will not influence the formation, spread and collapse of the sinkhole cavity. A cross-section of the sinkhole simulator is also shown in Figure 1.

In this study, a sandy soil with 1% passing the 200 sieve from Orlando, Florida, was chosen for the physical model. This soil was classified as a dark brown fine sand (AASHTO type A-3). The soil had an optimum moisture content of 13 %, a maximum dry unit weight of 104 lb/ft³, and a specific gravity of 2.6. The first step in the test was to seal the opening (limestone crack) using a rubber sheet in the bottom of the metal drum. Then, the pre-cleaned sandy soil with a moisture content 13% was well compacted in soil mold. Prior to adding the soil, the eight PVC pipes (monitoring wells) were installed at the radial locations shown in Figure 1. The thickness of the soil layer was varied between 150 mm and 200mm. The soil layer was fully saturated to a depth of 22.5 mm and 30 mm from the ground surface, respectively, for a period of 24 to 48 hours. These levels represent the shallow water table in the soil sample.
Results and Discussion

In this study, more than 30 model configurations were tested. However, the results of only four different tests are presented in this paper. The first two tests were with soil thickness of 150 mm (representing the overburden soil above the limestone bedrock) and with initial groundwater level at 22.5 mm from the ground surface. While, the other two tests were with a 200 mm soil thickness and an initial groundwater level of 30 mm from the ground surface. This sinkhole physical model is designed to run a sensitive spatial-temporal analysis by using a dense network of water level sensors to read the groundwater fluctuation with high resolution (0.5 mm) high sampling rate (100 Hz). The sinkhole occurred after 16.0, 19.7, 20.0, and 26.6 minutes in TEST 1, TEST 2, TEST 3, and TEST 4, respectively.

The results of TEST 1, 2, 3, and 4 are plotted in Figure 4 to illustrate the groundwater drops with time. These figures also show the effect of the radial locations of the eight monitoring wells prior to the sinkhole collapse. It was observed in all tests that the groundwater drawdown was faster in the wells closer to the predetermined sinkhole location than the wells further away from the center. This natural phenomenon is called the cone of water depression. In all tests, the cone of depression developed well before surface collapse occurred. It is also observed that the cone of depression gets steeper with time as the underground cavity within the sediments gets bigger.

In order to see the development of the groundwater cone of depression, the groundwater drawdown was plotted against the eight radial locations of the monitoring wells (i.e., 10 cm, 12 cm, 14 cm, 16 cm, 18 cm, 20 cm, 22 cm, and 26 cm distances from the sinkhole location). Figure 5 shows these plots for TEST 1 and TEST 2. It can be seen in Figure 5 (a, b, and c) that there is a very distinctive water cone that starts right after the initiation of the sinkhole formation by opening the bottom hole. The top of this inverted cone is pointing toward the sinkhole location and also its slope gets steeper as time gets closer to the sinkhole collapse. It is also observed that some of the water level sensors might not follow the sequence of the drop in the water level, which implies that a closer sensor shows a higher water level than a more distant sensor. This kind of behavior is possibly due to the inability of having a very homogenous soil all around the sample, since compaction level may vary somewhat within the same soil. However, the general trend of the
groundwater drawdown forms a very distinctive cone of depression, which can point to the potential location of a sinkhole that is developing underground.

During all tests, the sensor water readings showed distinctive progressive drops with time. The progressive drops were analyzed to investigate their relationship to the sinkhole collapse location and time. Only the results of TEST 1 were chosen to illustrate this behavior in this paper. As it is seen in Figures 6 and 7, there were progressive and sudden drops in the groundwater table. These drops start after initiating the sinkhole (by opening the hole) and then transferred from the nearest sensor to the sinkhole to the second nearest sensor with a time lag. These drops can be observed to move from the closer sensors to the further sensors with time. This behavior of the sudden drops of the groundwater level was also observed on the experiment display screen during the test, when the soil has a faster rate of sediment loss out of the bottom hole. This means that the progressive drops are representing a certain internal collapse of the cavity within the sediments. Also, the amplitude of the progressive drops is related to the rate of sinkhole formation. Thus, the progressive drops of the groundwater table can serve as an indicator for the potential location of sinkhole.

To avoid the overlap of the sensors data, only some selected sensors are studied in Figure 6 and 7. It can be seen clearly, that the progressive drops are repeatable behavior in different wells’ readings. However, these drops were transferred with a time lag from the near sensor to the furthest sensor from the predetermined sinkhole location. The most likely explanation for this behavior is the internal collapse of the cavities within the sediments, since all other parameters and factors related to sinkhole formation were controlled. One can notice the effect of the sinkhole underground formation in early stage at a groundwater monitoring well located near a progressing sinkhole first. Then this behavior might be transferred to the next monitoring well over a certain time period (time lag). This time duration varies depending on the distance that well is from the progressing sinkhole location. In general, the time lags in the progressive drops could be used to measure the proximity of the sinkhole. This can be achieved in the future by correlating the expected sinkhole time to the progressive drops of the groundwater table.

Finally, it can be noted that the trend of the variation of groundwater levels from all tests showed a good agreement in general. The spatial-temporal model proved that there is a groundwater cone of depression prior to the sinkhole surface collapse. This water cone indicates...
Figure 4. Groundwater level fluctuations with time in the sinkhole physical model test.
Figure 5. Groundwater selected readings in different times versus the wells radial locations.
the future potential location of the sinkhole collapse. Also, repeatable groundwater progressive drops were observed in all models. These progressive drops were transferred from one well to another over a certain time period called time lag. Both the progressive drops and their time lags can provide information relevant to the sinkhole locations and their progression rates.

Conclusions
In this paper, a small-scale sinkhole model used to physically simulate the natural sinkhole collapse and to provide a potential avenue to predict the location of a sinkhole. The sinkhole simulator consisted of two main components: The soil mold and the monitoring system. The monitoring system was used to conduct a spatial-temporal analysis of data collected from a network of groundwater monitoring wells (sensors). These wells were distributed in a radial pattern around a predetermined location of a sinkhole. A different soil levels (overburden soil) and initial groundwater levels were tested in this model. This model has a one circular opening to simulate a crack in the limestone that allows the transfer of a volume of soil through the dissolving bedrock layer. During all tests (more than 30 runs), the fluctuations in the groundwater levels showed a very distinctive trend. The level in the wells nearer to the sinkhole always showed water levels lower than the distant wells. This naturally occurring behavior can be referred to as a cone of depression. It can be concluded, that the current physical model was successful in showing the formation of this groundwater cone of depression that occurs before there are any surface signs of sinkholes. This, in turn, in a reverse manner, can be used in predicting the potential location of sinkholes that are forming underground and show no surface indications.

By studying sensor data, some progressive drops were evident, which are consistently seen at the same location over multiple runs. Also, these progressive drops migrate in time from the closer sensor to the sinkhole to the further sensor. This time lag behavior and the corresponding progressive drops are indicators of the potential location of sinkholes. Thus, both the progressive drops and their time lags can help in investigating the sinkhole locations and the sinkhole progressing rate. This can be achieved by correlating an actual progressing sinkhole to the groundwater table fluctuation and progressive drop measurements. This paper presented a simple physical model and more advanced testing is planned.

References


MONITORING THE THREAT OF SINKHOLE FORMATION UNDER A PORTION OF US 18 IN CERRO GORDO COUNTY, IOWA USING TDR MEASUREMENTS

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Abstract
Sinkhole formation is a common occurrence in northeast Iowa, and US 18 in Cerro Gordo County was constructed over an area where sinkhole formation had only been locally known. It had not been recorded or identified in the Iowa DNR database at the time. Since 2004, sinkholes have developed along the right of way. Geophysical surveys contributed very little in the identifying the cause. However a Soil Survey (drilling program) identified numerous voids within carbonate bedrock. The soil borings indicated that shale overlying the carbonate rock has been removed/eroded, and resulted in the development of a karst subsurface through the dissolution of the carbonate rock. Without removing the structural fill and site soils to expose the rock, it will not be possible to impede the natural processes occurring. An alternative approach was adopted and consisted of: (a) removing the existing pavement, (b) installing coaxial cables in trenches excavated within the subgrade, (c) replacing the pavement as double reinforced pavement (including shoulders), and (d) monitoring the cables using Time Domain Reflectometry (TDR). The cables are interrogated several times a day and data is transmitted via cellular modem to Iowa DOT facilities. Among the data transmitted is a log file of deformation activity along each of the cables which is evaluated and an action plan is initiated based on: (a) information in the activity file, and (b) updated plots for each cable. Unexpected behavior has been observed, with activity occurring annually between the months of September and March. Although several explanations have been proposed, there is no definitive correlation between locations of the activity detected by TDR, sinkhole locations, or geophysical anomaly locations. In spite of this uncertainty, real time remote monitoring for ground movement is continuing.

Background
In 2004, several sinkholes formed along US 18 between STA 248+00 and 250+00 (distance of 200 m or 600 feet). A couple of sinkholes formed on the shoulders of the eastbound lane (Figure 1 and Figure 2). In addition, several small sinkholes and one rather large sinkhole formed in the ditch of the eastbound lane. A geophysical investigation, to determine the cause and extent of the problem (and propose options for remediation) included a Ground Penetrating Radar Survey, a Resistivity Survey, and a Soil Survey (drilling program). No voids were identified beneath or within a couple meters of the pavement surface. It was decided to continue to monitor the area for sinkhole formation and to investigate the integrity of the roadway should additional sinkholes form within the limits of the area of concern (Figure 3 STA 248+00 to 253+00, 500 m or 1500 feet), as determined by a geologic review and analysis of the drilling data. In addition, as a remediation to the problem, a double reinforced inlay or overlay was proposed for the area of concern, as soon as it became practical.

Figure 1. Sinkhole that developed in the north shoulder of the eastbound lane.
In 2008, a large sinkhole formed within the EB ditch at STA 251+50, east of the previous sinkholes and within the area of concern. An investigation of the immediate area below the roadway included only Ground Penetrating Radar. No voids were identified beneath or within a couple meters of the pavement surface.

In 2009, a project was designed for letting that would install a double reinforced inlay, including shoulders, for the designated area of concern. In addition to the inlay, the installation of a Time Domain Reflectometry (TDR) System was chosen as a device for real-time monitoring of the formation of voids/sinkholes under the roadway.

Site Conditions and Regional Setting

Karst refers to geologic, hydrologic, and landscape features associated with the dissolution of soluble rocks, such as carbonates and evaporites. A common feature of karst landscapes are sinkholes, which form when the land surface collapses into subsurface voids formed in the slowly dissolving rock. In Iowa, carbonate rocks are the common bedrock throughout the state, and are

In 2005, a sinkhole developed at STA 249+30 just outside the Right of Way and within the limits of the area of concern (Figure 3). An investigation of the immediate area below the roadway included Ground Penetrating Radar (GPR) and Multi-Channel Analysis of Surface Waves (MASW). No voids were identified immediately beneath the pavement, or within a depth of a couple of meters beneath the pavement surface.

Figure 2. Sinkhole that developed in the south shoulder of the eastbound lane.

Figure 3. Plan view and cross section along project area.
mantled with a variable thickness of glacial and other unconsolidated materials. Where these unconsolidated materials are less than 15 m (50 feet), and particularly less than 7 m thick (25 feet), sinkholes may occur. There are three areas in Iowa where large numbers of sinkholes exist (Figure 4): (1) within the outcrop belt of the Ordovician Galena Group carbonates in Allamakee, Clayton, and Winneshiek counties; (2) in Devonian carbonates in Bremer, Butler, Chickasaw, and particularly Floyd and Mitchell counties; and (3) along the erosional edge of Silurian carbonates in Dubuque and Clayton counties (Halberg and Hoyer, 1982).

The site is along US 18 in Cerro County between C.R. S56 and US 65 south of Mason City near Floyd County (Figure 4). There is structural fill over natural sand and gravel, and rock is at a depth of 10 m (32 ft) (Figure 3). The geological interpretation of the subsurface below the area of concern (a 450 to 500 meter, or 1200 to 1600 feet, stretch of roadway) is that a buried streambed or glacial trough exposed the carbonate bedrock which has since been undergoing karst dissolution. There is shale overlying the carbonate rock east and west of the area (Figure 3).

**Fractures and Voids in Rock**

In response to the development of sinkholes (Figures 1 and 2), borings were performed by Iowa DOT in 2004 which located voids in the limestone rock underlying natural sand and gravel soils (Figure 3). Soil piping, a process of transporting material out of the soil column and into voids within the carbonate bedrock, is the assumed cause for the sinkholes. The voids are assumed to be locations where fractures in the rock have become enlarged.

**Sinkhole Development and Precipitation Events**

Sinkhole locations (Figure 3) correlate with locations of median drain pipes and arched pipe culverts below US 18 (Figure 5). The drain pipes at median dikes direct flow to the south side of US18, and the arched pipe culverts direct flow from south to north along a preexisting swale. Surface runoff is concentrated by the median drain pipes and preexisting swale to locations where sinkholes have developed. Major precipitation events occurred in May 2004 and in May/June 2008 (Buchmiller and Eash, 2010). It was during these periods when sinkholes developed.

**Geophysical Survey and Anomalies**

A geophysical survey was performed in 2005 using GPR and MASW. Anomalies were typically at depths less than 1 m and were mostly attributed to variations within the structural fill. Drilling to provide confirmation did not identify anything to account for the anomalies.

While there has been a variety of geophysical data obtained at this site, the focus of this paper is upon TDR measurements.

**TDR System Design and Installation**

The TDR system was designed to minimize disturbance of the subgrade after removal of the existing pavement, but provide early warning if sinkhole development occurs below the new double-reinforced pavement. Coaxial cables were installed in trenches at a depth of 1 m, and redundancy is provided by two trenches below each lane (Figure 5).

As shown in Figures 6 and 7, the existing pavement was removed and trenches were excavated. Two trenches, 0.5 m left and right of centerline under each lane were excavated to a depth of 1 meter (for a total of eight cables). The 22 mm diameter solid aluminum coaxial cable was laid out and cramped at a spacing of 15 m (50 feet) as shown in Figure 8 and 9. Larger crimps were made at either end of the monitoring zone (STA 252+65 and STA 248+00). After placing the cable in each trench (Figure 10), cement grout was placed over the cable (Figure 10 and 11) so that in the event of sinkhole formation, the grout will fracture and cause deformation of the metallic cable. The grout has a compressive strength of approximately 200 psi (compared with concrete which has a compressive strength of 3000 psi). The trench was then backfilled by compacting excavated soils, and the new pavement was installed.
The data acquisition system (DAS) consists of a programmable datalogger, TDR unit, coaxial multiplexer, and wireless modem (Figure 13 and 14). The datalogger controls the TDR and multiplexer and switches channels so that each cable is interrogated at specified times. The datalogger stores the acquired digital TDR waveform and the data points are then downloaded via wireless modem to Iowa DOT facilities.

The most common approach for monitoring ground movement using TDR technology is to compare current...
unexpected waveforms (discussed below) and the inability to explain their cause created significant skepticism. Furthermore, it created reluctance to implement the automated monitoring and email notification capabilities.

As a consequence, it was necessary to modify the algorithm used to monitor cable deformation, and also the algorithm used to initiate email notification to project personnel. These are now treated as separate functions of the monitoring system.

The TDR system will send out an Initial Notification or “alert” when it has detected cable deformation that meets alert levels based on criteria programmed into
TDR Activity

There have been thirty-three (33) events detected as summarized in Table 1. Thirty (30) events have occurred under the roadway. Fifteen (15) of the events under the roadway are located between STA 249+50 and STA 250+00 in the vicinity of the two arched pipes (Figure 5). The TDR activity in the vicinity of the two pipes has persisted from Sept 2011 to the present.

In November 2011 and January 2012, exploratory pavement core drilling was performed in the westbound lanes at STA 249+72, STA 249+75, and STA 249+78, but did not find evidence of subsurface movement between the pavement and subgrade, and GPR surveys in the westbound lanes from STA 249+00 to STA 250+50, and through the arched pipe culverts (see Figure 5) did not find any voids.

Table 1a. Summary of TDR Activity.

<table>
<thead>
<tr>
<th>Cable Station</th>
<th>Date Initiated</th>
</tr>
</thead>
<tbody>
<tr>
<td>EBTA</td>
<td></td>
</tr>
<tr>
<td>248+08</td>
<td>Nov 2012</td>
</tr>
<tr>
<td>248+75</td>
<td>Jan 2014</td>
</tr>
<tr>
<td>249+29</td>
<td>Jan 2014</td>
</tr>
<tr>
<td>249+68</td>
<td>Jan 2014</td>
</tr>
<tr>
<td>250+37</td>
<td>Nov 2014</td>
</tr>
<tr>
<td>EBTB</td>
<td></td>
</tr>
<tr>
<td>248+00</td>
<td>Dec 2011</td>
</tr>
<tr>
<td>249+57</td>
<td>Nov 2013</td>
</tr>
<tr>
<td>249+69</td>
<td>Jan 2013</td>
</tr>
<tr>
<td>249+80</td>
<td>Mar 2014</td>
</tr>
<tr>
<td>250+50</td>
<td>Mar 2014</td>
</tr>
<tr>
<td>EBPA</td>
<td></td>
</tr>
<tr>
<td>248+00</td>
<td>Dec 2011</td>
</tr>
<tr>
<td>249+30</td>
<td>Oct 2014</td>
</tr>
<tr>
<td>EBPB</td>
<td></td>
</tr>
<tr>
<td>248+00</td>
<td>Dec 2011</td>
</tr>
<tr>
<td>249+51</td>
<td>Jan 2013</td>
</tr>
<tr>
<td>250+09</td>
<td>Nov 2014</td>
</tr>
<tr>
<td>251+56</td>
<td>Nov 2014</td>
</tr>
</tbody>
</table>

EBTA = eastbound travel lane trench A
EBTB = eastbound travel lane trench B
EBPA = eastbound passing lane trench A
EBPB = eastbound passing lane trench B

Table 1b. Summary of TDR Activity.

<table>
<thead>
<tr>
<th>Cable Station</th>
<th>Date Initiated</th>
</tr>
</thead>
<tbody>
<tr>
<td>WBPA</td>
<td></td>
</tr>
<tr>
<td>248+24</td>
<td>Jan 2014</td>
</tr>
<tr>
<td>248+63</td>
<td>Jan 2014</td>
</tr>
<tr>
<td>248+88</td>
<td>Nov 2012</td>
</tr>
<tr>
<td>249+77</td>
<td>Jan 2015</td>
</tr>
<tr>
<td>WBPB</td>
<td></td>
</tr>
<tr>
<td>251+86</td>
<td>Dec 2014</td>
</tr>
<tr>
<td>WHTA</td>
<td></td>
</tr>
<tr>
<td>249+69</td>
<td>Sept 2011</td>
</tr>
<tr>
<td>249+77</td>
<td>Sept 2011</td>
</tr>
<tr>
<td>249+83</td>
<td>Sept 2011</td>
</tr>
<tr>
<td>249+89</td>
<td>Sept 2011</td>
</tr>
<tr>
<td>249+95</td>
<td>Sept 2011</td>
</tr>
<tr>
<td>251+27</td>
<td>Jan 2014</td>
</tr>
<tr>
<td>251+67</td>
<td>Dec 2014</td>
</tr>
<tr>
<td>WHTB</td>
<td></td>
</tr>
<tr>
<td>249+86</td>
<td>Nov 2012</td>
</tr>
<tr>
<td>249+93</td>
<td>Nov 2014</td>
</tr>
<tr>
<td>251+42</td>
<td>Feb 2014</td>
</tr>
<tr>
<td>251+60</td>
<td>Nov 2012</td>
</tr>
<tr>
<td>251+87</td>
<td>Jan 2015</td>
</tr>
</tbody>
</table>

WBPA = westbound passing lane trench A
WBPTB = westbound passing lane trench B
WBTA = westbound travel lane trench A
WBPB = westbound travel lane trench B
Obviously, the TDR activity has been generated by some mechanism other than ground movement and void formation that would occur if sinkholes were developing under the roadway.

**Discussion**

The premise for installation of the TDR system was monitoring of the development of sinkholes beneath US 18 between STA 248+00 and 252+60. Based on experience, it is anticipated that the grouted coaxial cables will be subjected to localized shear as the ground moves.

Beginning in September 2011, waveforms were being acquired that were not consistent with the expected cable deformation. When cables are deformed due to ground movement, distinct negative spikes develop in the TDR waveform due to reflections from these locations of cable deformation (Figure 16). The reflections that have developed are more characteristic of cable abrasion (Figure 17) rather than cable deformation.

When ground movement occurs, the grout is fractured and localized deformation of the cable begins. For example, during installation of the cables, crimps were made at a spacing of 15 m to provide distance references in the TDR waveform. These crimps provide distinct negative spikes in the waveform (see Figure 16). This is the type of reflection that develops when ground movement causes shear deformation of the cable.

The reflections that have developed are not indicative of cable shear or tensile deformation due to ground movement. The reflections which have developed in disparate locations along cables are associated with an inductive type of cable fault (see Figure 17). This type of fault is typically due to abrasion of the outer conductor of the coaxial cable.

The TDR reflections have been persistent, and have occurred in each of the eight cables (see Figure 18 for the cable installed in a trench under the eastbound travel lane). In addition to the unexpected TDR reflections that began developing in September 2011, there have been periods of intermittent reflections along cable. These intermittent reflections are common, and they are not associated with cable deformation. The unexpected reflections and the intermittent reflections were large, and extended over long portions of the TDR records. These created false alarms, and raised questions as to the validity of information being provided.

In Figure 15, locations of sinkholes, geophysical anomalies, voids in rock (based on drilling), and TDR activity are plotted. There are a few locations (EB STA 248+75, EB STA 249+29, EB STA 251+56, and WB STA 248+63) where there is some correlation. However, there are many other locations where persistent TDR activity has been recorded but there is no correlation with evidence of sinkhole locations.

![Figure 15](image-url)

*Figure 15. Overview of sinkhole locations, geophysical anomaly locations, and TDR activity locations.*
abraid the cables. In addition, TDR reflections indicative of cable crimping or shear have developed in the adjacent cables under the westbound travel lane at WBTB STA 249+86 and WBTA STA 249+69 (Table 1b).

**Cable Deformation Due to Shrinkage of Soils or Grout Shrinkage**

There has been an ongoing drought, and soil shrinkage is a possible mechanism. This theory would be supported by the fact that moisture is not entering cables at the abraided areas. This theory would require soil shrinkage sufficient to locally fracture the grout in tension and then continued shrinkage to pull the cable through the grout in order to abraid the cable. Similarly, shrinkage of grout could cause cracking, and then continued shrinkage could cause the cable to be pulled through the grout.

**Cable Deformation Due to Frost Heave**

Similar to soil shrinkage, frost heave could cause tensile fracture of the grout and continued heave could pull cables through the grout. However, the timing when the TDR reflections developed does not entirely coincide with the period when frost would be deepest which would likely rule out frost heave.

**Cable Deformation Due to Gnawing Animals**

The TDR traces are consistent with those obtained from cables where squirrels have gnawed the outer aluminum conductor (Cook, 2013). In addition, the development of sinkholes on either side of the eastbound lanes in 2004 is similar to those which have been created by burrowing animals (Barsness, 2011). Common burrowing mammals in the area include foxes, skunks, groundhogs, and

**Assessment of Anomalous TDR Response**

It has been possible to replicate the inductive type of TDR reflections. Using bench scale tests in the laboratory in which the outer conductor was progressively abraided, it is possible to produce the large positive TDR reflections such as those that have developed. There are several possible explanations as to the type of cable damage that is occurring which would generate these reflections.

**Cable Deformation Due to Ground Movement**

Ground movement is a conceivable mechanism. Ground movement due to sinkhole formation would cause fracture of the grout in tension and continued movement would pull the cable through the grout which could

![Figure 16. TDR waveform for crimp in cable (Tektronix, 1989).](image1)

![Figure 17. TDR waveform for abraided cable (Tektronix, 1989).](image2)

![Figure 18. TDR waveforms for the cable installed in eastbound travel lane trench A. Reflection from crimp at 56 m is circled. Reflection from abrasion at 245 m is indicated with arrow. Left axis is dimensionless reflection coefficient in rho.](image3)
gophers (Lorena. 2010). To assess this scenario, the site was visited by IDOT and IDNR personnel. One burrow entrance was located, but not near locations where TDR reflections have developed along the coaxial cables.

While these TDR reflections are not indicative of cable deformation due to ground movement, they could be precursor indications (e.g., collapse of large burrows).

Summary

Sinkholes have developed along US 18 in Cerro Gordo County where karst dissolution of carbonate rock is occurring in an area where overlying shale is missing. Soil piping of overlying soil material into voids that formed/enlarged along fractures is the assumed process for sinkhole formation.

Coaxial cables were installed in trenches beneath the roadway to monitor ground movement associated with sinkhole formation. TDR activity has developed in 2011, 2012, 2013, and 2014. This activity has been initiated each year during the period from October to January, and predominantly in November. Although activity has been detected in each of the eight coaxial cables, there is no definitive correlation between TDR activity locations, sinkhole locations, or geophysical anomaly locations.

Several explanations have been proposed for the activity, and based on the TDR waveforms, one possibility is that burrowing animals are gnawing on the cable. However, there is very limited evidence of animal burrowing.

The effectiveness of TDR measurements has been clearly demonstrated (O’Connor and Dowding, 1999). However, sometimes we encounter situations in which some of the data and its interpretation are not clear.

In spite of the uncertainty with regard to TDR activity to date, if ground movement impacts the coaxial cable the resultant cable deformation will be detected. The threshold for notifications is high enough that if it is reached the most likely cause is the deformation and breaking of the cable, which would result from the formation of a significant void/sinkhole beneath the pavement. Consequently, the primary objective of monitoring for voids/sinkholes developing beneath the reinforced pavement is being achieved.

References


Cook T. 2013. Personal communication. Director of Global Engineering, Times Fiber Communications, Inc.


Predicting the required quantity of grout needed to remediate a sinkhole-damaged home is a challenging task that involves significant amounts of uncertainty. The difficulty arises from the limited amount of subsurface information that is available to make subsurface predictions particularly in complex karst environments. In typical sinkhole investigations, our understanding of the subsurface is limited by the three to four data points (borings) that provide a small window into actual subsurface conditions. This information is normally obtained from borings and from information inferred by geophysical surveys. In many cases, the information is not sufficient to make accurate predictions of grout quantities. This paper will discuss the uncertainties in analyzing the many factors that influence grout prediction; it will provide a method of calculating grout quantities and discuss how one may moderate the difficulties in prediction of grout quantities. Examples of case studies are given showing pre-grout and post-grout information and the lessons learned from these comparisons.

Introduction
This paper is based on experiences with compaction grout in Florida and in particular in west-central Florida. A large percentage of the remediation investigations and grout monitoring projects have been performed for insurance companies who would like estimated and actual grout quantities to be reasonably close.

A factor in writing this paper was to elicit comments on the method presented here to ultimately provide the grouting community with a method that more accurately predicts grout quantities.

The prediction of grout quantities is an imprecise practice; it is imprecise because we are trying to measure conditions that are irregular in both vertical and horizontal directions. These conditions result from variable chemical and mechanical weathering patterns that serve to complicate subsurface conditions. Compounding the problems with interpreting complex subsurface conditions is that we have direct information about these conditions from only three or four borings. In the absence of more definitive information we are required to extrapolate over large distances to fill in the sizeable amounts of missing information. Furthermore, since the karst subsurface is not uniform, the complexity of the subsurface becomes a controlling factor in the accuracy of grout predictions.

In many cases the equations we use to calculate grout quantities provide weighted averages for assumed conditions located 20 feet (6.1 meters) or 50 feet (15.2 meters) apart.

Extrapolation between distant data points and variability in such factors as porosity, soil composition, induration etc. compromise the accuracy we obtain in our grout models. In this paper we will discuss the causes of subsurface variability and methods we can use to mitigate them.

Cost of Compaction Grouting
When damage in a building is caused by active sinkhole conditions, owners and insurance companies are anxious to determine the cost of repair. A major expense in remediation is the quantity of compaction grout needed for remediation. Typically, average grout quantities range from 200 cubic yards (152.9 cubic meters) to 500 cubic yards (382.3 cubic meters) at a cost of approximately 175 dollars per cubic yard; this can result in a cost for compaction grout alone of from $35,000 to $87,500. Adding to the grout cost is the cost for drilling grout holes, chemical grout and monitoring. These other costs can easily cause the total to reach $100,000 or more.

Subsurface Information
As stated, most grout estimates are based on assumed subsurface conditions defined by essentially three or four borings and a geophysical survey. The completeness of this information is dependent on the complexity of the subsurface. To illustrate this, consider the idealized subsurface conditions shown in Figures 1 and 2. In Figure 1 where the subsurface is relatively uniform...
circulation in all or some of the borings, is the rock surface highly weathered etc. These and other factors are signs of potential problems in predicting grout quantities.

**Method**

Equation 1 is the method used to calculate grout quantities. The equation is rather simple; it calculates the volume of grout cylinder injected into the ground considering the thickness of each sequence of unique soil found in the borings. Factors are added for the assumed porosity, void reduction, continuity and uncertainty of each material found in the borings. The problem with this calculation is that it is greatly dependent on the information obtained in the borings and in particular on the assumed extent and variation of the soil properties occurring between boring locations. The illustrations in Figures 1 and 2 provide examples of potential variations in the limestone surface (and potentially in variations in the soil conditions occurring above the limestone). In most cases, the greater the variation in the surface of the rock, the greater the likelihood for error in estimates of grout quantities.

In Figure 1 the rock conditions are relatively uniform and grout estimates are generally more accurate. In contrast, Figure 2 shows a relatively irregular rock surface where prediction of grout quantities is difficult because of the irregular rock surface and the greater error that occurs in interpreting between borings. In fact, depending where the boring samples the rock surface, one may be misled into assuming the conditions in Figure 2 are the same as in Figure 1.

Equation 1 (Hussin, 2012) is as follows:

\[
G_p = T \times \pi \times r^2 \times n \times V_r \times UF
\]

Where:

- \(T\) = Thickness of soil to be grouted
- \(r\) = radius of effective treatment area (typically 0.9 meters, reductions can be made for larger clayey components found in the soil section)
- \(n\) = Porosity (\(V_v/V_t\)) of dominant soil types. Typically: (SMloose=0.45, SMdense=0.25; CLsoft=0.55, CLstiff=0.37)
- \(V_r\) = Void reduction factor is the amount you expect the porosity to be reduced. 30% reduction is typical.
- \(G_p\) = Amount of grout estimated per grout point
- \(UF\) = Uncertainty factor (10 - 75%). Depends on confidence in soil location and composition, see text.

In reality we only see less than 1% of the soil material under the footprint of the building (Zisman and Clarey, 2013).

**Calculation of Grout Quantities**

**Unknowns**

Up to now it has been discussed that grout is expensive and difficult to predict where and in what quantity it will be needed. Unfortunately, it is not possible to write an equation to quantify what is not known. What can be done is we can minimize the unknowns by being alert to likely excursions in the soil profile. For example: are N-values uniform (is there continuity in lithology from one boring to another), is there a great difference between the depth to rock or depth to soft material from one boring to another, is there loss of...
Examples of Grout Quantity Determinations

Two examples are given of compaction grout determinations. Example 1 is illustrated in Figures 3, 4, and 5, while Example 2 is illustrated in Figures 6, 7, and 8. In Example 1, grout quantity was estimated to an accuracy of 92%, while in Example 2 grout quantity was estimated to an accuracy of 69%. In each of the examples, an explanation is given detailing the factors that help or hinder the determination and how accuracy can be improved.

Figure 3. Subsurface profile in Example 1.
Figure 4. Depth to top of rock from borings in Example 1.

Figure 5. Depth to sound rock from grout holes in Example 1 along with relative amount of grout used.

Figure 6. Subsurface profile in Example 2.
The figures provided in each example show: 1) a profile of the conditions found in the borings, 2) a profile of the depth to rock found in the site borings and 3) a profile of the depth to rock found in the grout holes. On the grout hole depth to rock figure, the relative amount of grout pumped in each hole is shown. This was done to determine if there was a correlation between the depth of the drill hole and the amount of grout used.

**Example 1, Grout Prediction**
In Example 1, the overall soil profile generally consists of: dense to very dense soil from a depth of 10 feet (3.0 meters) to 65 feet (19.8 meters) below land surface (bys). This is followed by intervals of soft/loose material at various depth intervals in three of the four borings. The soft/loose material found in this section (Figure 3) will account for the largest amount of grout, depending on the uniformity of the soft/loose zones. The dense soils, which make up the upper portion of the profile, will take substantially less grout. In some cases where relatively uniform N-values are found, CL and SC material may be combined in calculation of grout quantities. This overall interpretation of the soil profile was used in the calculation shown in Table 1 and resulted in an overall accuracy of 92% in compaction grout prediction.

**Example 1 Discussion**
In Example 1, our estimate of grout volume provided a 92% prediction accuracy. The high accuracy in this example is the result of the relatively flat relief of the rock surface coupled with the limited extent of soft/loose soil conditions. These results are summarized in Table 3 where it is seen that a small difference was found in relief in the limestone surface determined from the borings compared to that determined from the grout holes. This small difference in relief (16 feet—4.9 meters) of the limestone rock surface measured from the results of four data points (borings) versus 35 data point (grout holes) further suggests the surface is relatively flat. In addition, the distribution of soil materials was more uniform and hence resulted in a more accurate estimation of soil variables used in grout prediction.

**Example 2, Grout Prediction**
Subsurface conditions in Example 2 are different than those in Example 1. First, the loose/soft material
in Example 2 is in the upper part of the profile and its distribution appears to be more widespread. This suggests that a large part of the zone above the limestone may have to be grouted. Second, there is greater relief in the rock surface in Example 2. This greater variability in the rock surface translates into greater variability and complexity in the soil overlying the rock compounding errors in the nature and distribution of the soils.

A further indication that Example 2 will take more grout is the variability in thickness of the loose/soft zone and the more widespread distribution of this zone. An example of the grout calculation for Example 2 is shown in Table 2.

**Example 2 Discussion**

Table 3 provides a comparison of the grout results in the two examples and helps to elucidate why more error occurred in the grout prediction of Example 2.

**Concluding Discussion**

The following is a summary of the conclusions reached in this paper:

1. We have small windows into the subsurface from the three or four borings drilled for the investigation. This limited amount of data may not give sufficient information into the composition of the subsurface. Figures 1 and 2 provide an illustration of this problem.

2. To compensate for the absence of detailed subsurface data, we must examine other cues that may be present in the subsurface. This information can be found in: historical air photos, grout information from neighboring homes, topographic relief of the regional ground surface, geologic history of the area etc.

3. Equation 1 can provide reasonably accurate (+/- 35%) predictions of grout quantities if we can rationally estimate the type and variability of the soil and rock conditions between boring locations. This requires use of incidental information from geology, aerial photography, topographic, neighboring homes etc.

4. The factors in Equation 1 should be applied judiciously. They should reflect the uncertainty in the subsurface.

5. We can expect higher grout quantities when the relief in the rock surface increases or if abrupt changes in lithology are found between the borings. Conversely, if the lithology is

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**Limestone Relief** is the difference in elevation between the highest and lowest expression of rock

**Diff. in Relief** is the difference between the limestone relief found in the grout holes minus the limestone relief found in the borings.
continuous from one boring to the next, we can make more accurate grout predictions.

6. An additional factor that can significantly impact the amount of grout placed for remediation is the manner in which the grout is placed. Figure 9 illustrates this problem. During grouting high pore pressures develop that can cause the soil to fail in an undrained state, remolding the soil into a liquefied mass. This causes an increase in the amount grout used and weakening of the subsurface. This condition can be mitigated by limiting the pressure and limiting the maximum amount of grout that can be placed in a 24-hour period in one grout point.

7. In the examples shown here, no strong correlation was found between grout hole depth and grout take. However, many factors were present that obscure this relationship, such as distance to boring, changing soil conditions, borehole conditions, etc.

8. It is hoped that this paper will encourage others to share their methods of predicting grout quantities and share their experience with using this or similar methods to predict grout quantities.

References

Figure 9. Potential problems with improper grouting.
PRE-CONSTRUCTION ROCK TREATMENT AND SOIL MODIFICATION PROGRAM USING LOW MOBILITY GROUT TO MITIGATE FUTURE SINKHOLE DEVELOPMENT IN A 2,787.1 SQUARE METER (30,000 SF) MAINTENANCE FACILITY

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Abstract
The US Army required construction of a 2,787.1 square meters (30,000 sf) maintenance facility supported on shallow foundations at the Fort Campbell Military Installation. During the subsurface investigation a seven foot air-filled void was encountered in the bedrock within the building footprint. Electrical Resistivity Imaging (ERI) was conducted in an attempt to determine the lateral extent of the encountered void and to establish the general prevalence of karst features at the site. Due to uncertainty in the subsurface conditions, a rock treatment and soil modification program was developed which consisted of a series of targeted exploratory grout holes advanced in 126 locations in the structural areas of the building footprint. The intent of the program was not to prevent the development of a soil dropout, but to improve the foundation support of the structure so that the facility would perform acceptably if a future soil dropout were to occur during the design life of the facility. This was achieved by targeting each footing with 3 exploratory grout holes. The intent of each grout injection was 1) to identify the top of rock elevation, 2) determine if a karst feature existed, 3) cap the karst bedrock below the footing and treat defects in the rock, and 4) provide localized improvement of soft soils through the use of low mobility grout columns under each footing. Drilling refusal elevations were obtained for every grout hole and were assumed to represent the top of bedrock. Each exploratory hole was closely monitored for pressure and volume in 0.61 meter (2-foot) stages. Zones where the bedrock had lower elevations or took excessive grout at low pressures were targeted with additional tertiary holes. The tertiary holes were verified with additional SPT sampling. Documented ground improvement was achieved, evident by increased SPT blow counts ranging between 25 to 50+ post treatment. Based on results from this program, lower grouting pressures could have been utilized as part of the refusal criteria to successfully identify and treat karst features.

Geology and Subsurface Exploration
The project site is located in the Mississippian Plateau, an upland region mostly underlain by Upper Mississippian Series limestone and dolomite assigned to the Ste. Genevieve and St. Louis Members of the Slade Formation. Regional geomorphology includes intense karst development including the Mammoth Cave-Flint Ridge cave system to the northeast. As a result of karst development, the plateau has developed a complex pattern of sinkholes and solution features within the bedrock. A chert zone near the soil/bedrock interface was assumed to mark the contact between the Ste. Genevieve and St. Louis Members. Bedrock at the study site exhibited erratic pinnacles extending into the overburden soils with soil-filled slots extending into the rock unit. During excavation of the site several pinnacles were encountered and removed. Proximal sites to the north, west, and south have a documented history of sinkhole development as shown in Figure 1.

During site reconnaissance, a slight closed surface depression was observed in the building footprint along with three confirmed sinkholes along the southern and western perimeter of the site. Final grades for the site required construction of two retaining walls to level the existing hillside to achieve finished grade. Existing grades in the area of the building footprint required between 0.61 and 3.05 meters (2 and 10 feet) of cut to achieve the finish subgrade elevation for the building. Bedrock refusal depths ranged between 5.5 and 12.5 meters (18 and 41 feet) below the ground surface (bgs) based on the geotechnical investigation. Site soils in the northeast and southeast corners of the proposed facility contained 3.1 to 4.6 meters (10 to 15 foot) zones of very soft soils with rod drops of 0.3 to 0.61 meters (1 to 2 feet) during SPT sampling. The average SPT N-value across the site at a depth of 7.6 meters (25 feet) was 7 bpf with several zones recording blow counts of 0 or 1. Drilling fluid was lost in the two boreholes where rock
core was obtained and a 2.1 meters (7 foot) void was encountered in the bedrock at the NE corner of the building as shown in Figure 2.

Karst features encountered at the site coupled with the existing topography of the area limited options for remediation. Relocation of the facility to a different site or moving the building on the current site were not feasible options. Additional investigations would be required if the design was going to proceed for construction at the current site. Discussions with Geophysicists led to the selection of Electrical Resistivity Imaging (ERI) to attempt to further characterize the subsurface conditions at the site in terms of the varying geology and to attempt to detect any additional karst features using an AGI Super Sting R8 IP system. The geophysical program consisted of 11 traverses totaling 1,219 linear meters (4,000 linear feet) utilizing several arrays including the Dipole-Dipole array, Wenner array, and inverse Schlumberger array. The electrode spacing used for the survey was a function of the geometry of the traverses and ranged from 1.83 meters (6 feet) for north-south oriented profiles and 2.44 meters (8 feet) for east-west oriented profiles using 56 electrodes. Results were mixed as additional zones of weathered bedrock and pinnacles were identified but the void encountered during drilling was not detected, likely due to interference from a steel fence on site. The test results were consistent with potential karst geology but did not provide conclusive evidence. An excerpt from the ERI Plots is shown in Figure 3. The Geophysicist recommended additional site drilling or Seismic Refraction Techniques. Although the extent of karst issues on the site was still unclear, schedule and funding prevented additional investigations. However, it was clear the risk for future karst development was unacceptably high. If additional investigations would have been initiated the use of Cone Penetrometer Testing (CPT) would likely have been utilized in an effort to better define the variability in the top of the bedrock surface.

**Low Mobility Grouting Rationale**

The subsurface and geophysical investigations led us to believe that the risk to the building would require remediation and also raised awareness that the western portion of the building was also at a very high risk from potential karst features based on interpretation of the ERI results. After meeting with key stakeholders, it was determined that pretreatment of potential karst features in the bedrock would be attempted through low mobility grouting. Traditional low mobility grouting programs using a 2.43 to 3.66 meter (8 to 12 foot) grid system under the entire building footprint have been used effectively at Fort Campbell in the past. If such an approach was used at the study site this would have required about 375 holes at a cost of approximately $1,000,000. The approach from this case study utilized targeted exploratory grout holes for cap grouting of the bedrock and improvement of the overlying soft soils specifically in the structural areas of the building, and was accomplished with approximately 111 holes (not including tertiary holes).
The intent of the program was not to prevent the development of a sinkhole but to improve the foundation support to better survive a soil dropout and minimize disruption to the military operations if a sinkhole were to occur during the design life of the facility. This was achieved by targeting each spread footing with 3 exploratory grout holes as shown in Figure 5. The intent of each grout injection was 1) to identify the top of rock elevation, 2) determine if a karst feature existed, 3) cap the bedrock in the area and treat the defect in the rock, and 4) provide some localized improvement of soft soils through the use of low mobility grout columns under each footing. This methodology allowed for the individual assessment and comparison of the subsurface conditions at each footing location. Direct comparisons could be made comparing the depth to bedrock and considerations given to the specific subsurface characteristics of each footing to determine whether tertiary holes or significant revisions to the grouting plan would be required.

at a cost of approximately $225,000. This low mobility grouting methodology consisted of a grout program that required the advancement of casing into the ground, initiating cap grouting at the soil/bedrock interface, and then installation of grout columns above the cap grouting zone as the casing was removed from the ground. Grout columns are singular elements which compose typical compaction grouting techniques, except in this case a formal grid pattern for the entire building footprint is not established.

The rationale behind this technique is that preventing future sinkhole development is highly improbable, and therefore there are diminishing returns in treating the entire building footprint in floor slab areas and also treating the pinnacle areas in the bedrock. Therefore a plan was devised to treat the soil specifically within the zone of influence of the spread footings as shown in Figure 4.

The intent of the program was not to prevent the development of a sinkhole but to improve the foundation support to better survive a soil dropout and minimize disruption to the military operations if a sinkhole were to occur during the design life of the facility. This was achieved by targeting each spread footing with 3 exploratory grout holes as shown in Figure 5. The intent of each grout injection was 1) to identify the top of rock elevation, 2) determine if a karst feature existed, 3) cap the bedrock in the area and treat the defect in the rock, and 4) provide some localized improvement of soft soils through the use of low mobility grout columns under each footing. This methodology allowed for the individual assessment and comparison of the subsurface conditions at each footing location. Direct comparisons could be made comparing the depth to bedrock and considerations given to the specific subsurface characteristics of each footing to determine whether tertiary holes or significant revisions to the grouting plan would be required.

Figure 3. ERI Plots Indicating Pinnacles and Valleys in the Bedrock Surface.
Low Mobility Grouting Methodology

The methodology consisted of advancing casing to the top of bedrock, cap grouting through a port in the casing at the top of bedrock, and finally installation of low mobility grout columns as the casing was withdrawn. Cap grouting was utilized to reduce infiltration and piping of groundwater and soil material into openings in the bedrock, and provides a barrier that prevents soil loss into bedrock voids. The grout material was thick enough to bridge and choke off small defects in the bedrock surface and withstand soil and hydrostatic pressures, typically 0.3 to 0.61 meters (1 to 2 feet) thick. Additionally, since cap grouting was only conducted in the structural areas, impacts to the existing groundwater flow regime were minimized.

After cap grouting was completed, low mobility grout columns were used to improve the upper soils by inducing lateral pressures between grout holes at each footing location. The low pressures of the grout injection displaced the native soil which densified the soil overlying bedrock. Typically more compaction effects occur at deeper depths than near the surface of the overburden. This was accomplished by removing the grout pipe in 0.61 meter (2 foot) stages and injection of a low mobility grout pumped at a low pressure to form grout columns above the cap grouting zone. Through monitoring the variations in volume and pressure at a constant flow rate, potential karst features or softer, weaker zones of soil can be identified. The casing was removed as the grouting continued such that a continuous series of grout bulbs was created from the bedrock to within 3.05 meters (10 feet) of the ground surface. Grout columns were terminated at this depth below the bottom of the column footings to provide a soil zone below the footing to allow for some natural settlement under the column footings and to reduce differential settlement between the floors, walls, and columns. Other benefits of this technique included strengthening soft soils associated with karst features, reduction of the anticipated total settlement associated with the structure, and increasing the bearing capacity of the soil.

The grout consisted of a stable, sanded grout with a slump between 5.08 to 12.7 centimeters (2 to 5 inches). The grout mix did not bleed and had a consistency of mortar. Tight control on the water/cement ratio was maintained through grouting operations. The grout strengths achieved were approximately 5,171 to 8,618.4 kPa (750 to 1,250 psi) to be strong enough to bridge across small voids at the top of the rock. Many grout mix designs are proprietary and will vary based on the grouting equipment, grout pump, soil characteristics, and admixtures.

Low Mobility Grouting Refusal Criteria
Grout Injection continued within each zone beneath the injection pipe until one or more of the following occurred:

1. Grout flow ceased at a header pressure reading of 2,757.9 kPa (400 psi)

![Figure 4. TEMF Foundation Plan with Spread Footings and Grout Injection Locations.](image-url)
tion required the designers to be an integral participant in the decision making process during the grouting program. The project specifications required the submital of a detailed work plan and a grouting plan by the contractor within 30 days of initiating work. This allowed for a thorough evaluation of the contractor’s experience, proposed equipment and methods, and understanding of the project objective. Additionally a pre-construction meeting was required between all stakeholders at which time details of the proposed work plan, specifications, and local geology were discussed in depth. Several key points were refined during these discussions and contingency plans were developed. The need for coordination and discussion prior to mobilization on projects such as this is critical to the overall success and efficiency of the grouting program.

**Low Mobility Grouting Summary**

For each grout injection location, a table was produced as shown in Figure 6, which tracked the time, volume, pressure, and refusal criteria. Real time plots of volume and pressure were maintained by the Contractor but were not required for submission in this contract.

![Figure 5. Planned Method of Treating the Subsurface Conditions Under the Structural Elements of the Facility.](image-url)
During and after completion of the planned grouting program, the holes were evaluated to assess if tertiary holes were required. Footing locations requiring tertiary holes were selected based on the following criteria:
1. Did the lower portion of the injection take the maximum allowed volume per stage.
2. The pressure sustained while grouting in these zones was less than 1,034.2 kPa (150 psi).
3. The elevations where bedrock was encountered in the grout holes.

The grout injections were evaluated based on the pressures, volumes, and the top of rock elevations to prioritize high risk footings. The bedrock elevations were plotted into a contour map shown in Figure 7. Areas determined to be at most risk for future sinkhole development were identified and targeted with tertiary holes. It is noted that the variation of the apparent top of bedrock elevations at the project site varied substantially more than originally believed by the original geotechnical and geophysical investigations.

The advancement of 15 tertiary holes was recommended for this project. After the tertiary holes were advanced and grouted a noted reduction of volume per stage was documented at increased sustained pressures; signs that soil improvement in these structural areas occurred. This was further verified through selected SPT sampling in areas between the grout holes. The average SPT blow counts at a depth of 7.62 meters (25 feet) prior to the grouting program was 7, and SPT blow counts after completion of the grouting program was 38. Based on this data, the grouting program was determined to be a success. The final grout program consisted of 359.3 cubic meters (470 cubic yards) of grout with 297.4 cubic meters (389 cubic yards) required for grout columns and 61.93 cubic meters (81 cubic yards) required for cap grouting. This required approximately 853.4 meters (2,800 ft) of drilling and casing installation. Each hole was completed for about $2,500 with an average depth of 7.01 meters (23 feet) and an average grout take of 4 cubic yards per hole. The final cost including tertiary holes was approximately $300,000.

Lessons Learned
From a geotechnical engineering and foundation design perspective, if posed with a similar design challenge such as this in the future the approach would likely change both during the investigation phase and in the grouting phase. In karst prone areas within the Louisville District COE boundaries the transition away from Electrical Resistivity Imaging and Seismic Refraction in favor of Cone Penetrometer Testing (CPT) has been effective and resulted in cost savings on similar projects when used in conjunction with conventional hollow stem auger and rock coring methods. CPT can be useful for determining refusal depths, characterizing subsurface conditions near the soil/bedrock interface, and for the assessment of potential risk posed by karst. On subsequent projects where karst was a concern using this methodology has actually led to the elimination of some anticipated grouting programs through the additional subsurface information provided. Additionally if shear wave velocities are determined through CPT methods some projects have actually been able to justify improvements to the seismic site classification resulting in significant cost savings.

A typical geotechnical investigation utilizing this approach would consist of 2 building borings extending to the top of bedrock, 2 to 4 additional borings extending

Figure 6. Typical Grout Log for One Injection Point.
to the base of the anticipated zone of influence for the building, combined with an appropriate number of site borings. Then an additional 10 to 40 CPT holes could be advanced to refusal to provide a more tangible top of bedrock surface and estimates for soil properties at depth compared to geophysical methods at a similar cost. Had a grid of CPT holes been advanced at the study site with this methodology the data could have been used to better tailor the grouting program to the most karst prone areas with the weakest soils.

The grouting program itself should be modified to restrict the maximum gage pressure to four times the hydrofracture pressure of the soil. The grouting program improved SPT blow counts by a factor of 5. This seems to suggest the program was highly effective, but inefficiencies in the grouting program can result from hydrofracture of the soil. Considering the grouting program allowed the industrial standard of 2,757.9 kPa (400 psi) gauge pressure for grout holes that extended between 6.1 and 13.7 meters (20 and 45 feet) below the ground surface, a pressure 10 to 12 times the hydrofracture pressure of the soil at those depths was routinely introduced at the study site. A reduction of the allowable grouting pressures would have resulted in a lower grout volume required to achieve the desired subsurface improvement.

Several methods of determining the hydrofracture pressure exist. The simplest form of the equation was presented by Mori and Tamura in 1987 where the hydrofracture pressure is represented by the equation:

\[ P_f = \sigma_3 + q_u \]

Where \( P_f \) = Maximum allowable pressure to initiate hydraulic fracture, \( \sigma_3 \) = Horizontal Stress (Minor Principle Stress), and \( q_u \) = Unconfined Compressive Strength. Applying this equation to the case study site at a depth of 7.62 meters (25 feet) below the ground surface you obtain a hydrofracture pressure which is determined as follows:

| Vertical Stress, \( \sigma_1 \) | = 25 ft* (125 pcf) 
| | =3,125 psf 
| | =149.6 kPa 
| Horizontal Stress \( \sigma_3 \) | = \((\mu/1-\mu)( \sigma_1 )\) 
| | = (0.4/1 - 0.4) (3,125psf) 
| | = 2,083 psf 
| | =99.7 kPa 

Where \( \mu \) = Poisson’s Ratio = 0.4 and \( q_u \) = 143.6 kPa (3,000 psf) as determined by laboratory testing. Therefore \( P_f =99.7 \) kPa + 143.6 kPa = 243.4 kPa \( (P_f =2,083 \text{ psf} + 3,000 \text{ psf} = 5,083 \text{ psf} = 35 \text{ psi}) \) and hydraulic fracture of the soils can occur at pressures as low as 243.4 kPa (35 psi). An example of hydraulic fracture is shown in Figure 8.

Some hydrofracture of the soils should be anticipated and in a controlled manner can be effective. However considering one objective of a low mobility grouting program is to densify soft/weak soils, using pressures as high as 2,757.9 kPa (400 psi) results in instances where a lot more grout can be pumped into the ground than is actually needed to improve the soils. This is because severe hydrofracture causes erratic grout travel, excessive grout takes with minimal soil improvement, and sometimes unwanted damage.
Based on this experience, future foundation grouting projects should consider reducing the maximum allowable gage pressure to between 861.8 to 1379 kPa (125 to 200 psi). This was the pressure threshold where tertiary holes were warranted in this case study. However if every stage was terminated at 1379 kPa (200 psi) considerable savings would have been recognized in the 297.4 cubic meters (389 cubic yards) of low mobility column grouting that occurred, possibly as much as 50% of the total grout quantity. Some hydrofracture of the soils would still occur, but a better balance between the total volume of grout expended from hydrofracture and soil improvement would be achieved using reduced pressures at the same flow rate.

**Conclusion**
The grouting methodologies utilized on this project improved soil strength in zones where very low shear strengths were encountered. The methodology allowed for specific assessment and treatment based on the founding conditions at each structural element of the facility. The information gathered during the grouting program allowed for better certainty regarding the top of rock elevations at the site. There are no guarantees that future sinkhole development will not impact this facility, but the extent of potential structural damage to the facility has been minimized. Additionally, a performance specification using industry standards in grouting is functional, but as practices are advanced and evolve designers should be cognizant that more efficient methods and processes can still be explored and refined to minimize costs and maximize benefits associated with pre-treatment of high-risk soils in critical structural areas, and even to practices applied for sinkhole repair.

**Reference**
SUCCESSFUL FOUNDATION PREPARATIONS IN KARST BEDROCK OF THE MASONRY SECTION OF WOLF CREEK DAM

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Abstract

Extensive foundation preparations during construction of the Wolf Creek Dam concrete masonry section precluded the need for additional rehabilitation to mitigate seepage through karstic limestone bedrock. Wolf Creek Dam on the Cumberland River in southern Kentucky has become well known for karst related seepage issues underneath the embankment section, and yet has had little to no seepage issues associated with the concrete masonry portion of the dam. Post-construction efforts to control seepage underneath the embankment began in 1967 and 1968. Emergency grouting commenced and continued through 1970. Between 1975 and 1979 a more permanent solution of a concrete diaphragm cutoff wall was constructed through the centerline of the left portion of the embankment section down to competent bedrock. The wall interrupted the progression of foundation erosion, but post construction monitoring, instrumentation readings, and persistent wet areas downstream showed that seepage paths under or around the wall continued. A second cutoff wall upstream of the first was constructed between 2007 and 2013, extending nearly the entire length of the embankment and up to 75 ft (22.9 m) deeper than the original wall. Cost of the second wall and other concurrent rehabilitation efforts reached nearly $600 million. Exploratory grouting beneath the concrete masonry section of the dam in 2012 resulted in low grout volume takes, so no further remediation efforts below the masonry dam were conducted. The original construction photographs and foundation reports for the concrete masonry section of Wolf Creek Dam instill confidence that the designers and builders of the monoliths took adequate, if not excessive measures to ensure that all the monoliths were founded on competent bedrock. These measures included extensive borehole investigations both prior to and during excavation, efforts to locate, delineate, remove, and clean all karst solution channels, the removal of all loose rock, grouting in the foundation and side vertical faces, large stair-step faces on the left abutment, extended excavations to remove soft beds, final manual cleaning of rock surfaces, and the careful documentation of foundation preparations. These measures do not guarantee that seepage issues will not develop under the concrete dam over time, but they do show with reasonable certainty that the monoliths were originally founded on competent bedrock, and that future seepage issues are either unlikely or will be significantly inhibited by the preparation made to the foundation prior to the construction of the concrete monoliths.

Introduction

Wolf Creek Dam, built and operated by the U.S. Army Corps of Engineers on the Cumberland River in southern Kentucky, is a flood control and hydropower dam that impounds Lake Cumberland, the largest Corps reservoir east of the Mississippi River storing about four million acre-feet (4.9 billion cubic meters), with up to six million acre-feet (7.4 billion cubic meters) maximum storage. The dam has a maximum height of 258 ft (78.6 m) and consists of a 3940 ft (1200.9 m) long compacted clay embankment dam extended from the right, or east abutment, which ties into a 1796 ft (547.4 m) long concrete masonry dam and gated spillway extended from the left, or west abutment (Figure 1). Flow is passed through six turbines rated at 45,000 kW each, and through an additional six sluice gates 4 ft by 6 ft (1.2 m by 1.8 m) each. Floods are passed over the spillway through ten tainter gates 37 ft by 50 ft (11.3 m by 15.2 m) each. The safety of the dam has come into question in recent years and it is estimated that a breach of the dam would result in

Figure 1. Aerial view looking upstream at Wolf Creek Dam.
between 100 and 1,000 fatalities, mostly within the city of Nashville, TN, located 246 miles (395.9 km) downstream of the dam (USACE, 2014).

Designed and constructed from 1938 to 1952 over karstic limestone bedrock efforts were undertaken before, during, and after construction to prevent the seepage of reservoir water through the foundation from compromising the integrity and safety of the dam. These efforts included but are not limited to the construction of an embankment cut-off trench near the upstream toe, grouting before and during construction, emergency grouting of the downstream embankment in the late 1960s and early 1970s, construction of a concrete diaphragm barrier wall in the embankment from 1975 to 1979, and the construction of an additional concrete barrier wall from 2007 to 2013. The latest remediations to the embankment alone cost nearly $600 million. Since construction the masonry section of the dam has had few costs beyond normal operation and maintenance with no issues of seepage, settlement, stability, or other potentially karst related problems. This is due to the extensive foundation preparations undergone during construction which included the removal of all soft and solutioned bedrock.

**Geologic Setting**

The dam was built across the Cumberland River valley within the Highland Rim; a low plateau of nearly horizontal beds of limestone, shale, and chert ranging from Ordovician to Mississippian in age. The dam is located approximately 20 miles east of the crest of the Cincinnati Arch, a broad up-fold extending northeast-southwest across central Kentucky, which gives the bedrock a slight dip of 30 ft per mile (5.7 m/km) to the southeast or upstream direction.

The foundation of the dam is composed of mostly limestone bedrock with some river alluvial deposits left below the embankment portion of the dam (Figure 2). The Catheys Limestone Formation underlies the entire area and is described as hard, thin-to-massive bedded, dark gray, and argillaceous limestone interbedded with thin, well-cemented, calcareous shale. The Leipers Limestone Formation sits unconformably above and is very similar to the Catheys. In general it is thinner bedded, more argillaceous, and more fossiliferous. It forms the valley floor and the lower portion of the abutments. Within the abutments the Leipers is overlain by the Cumberland Limestone Formation, which is a dense, greenish-gray, massive, non-fossiliferous, arenaceous to argillaceous, dolomitic limestone. The Chattanooga Shale sits above that, which is primarily a fairly hard, well cemented, fissile, black, carbonaceous, and silty shale, with a 4 to 5 ft (1.2 to 1.5 m) base of gray shale that is more susceptible to weathering and erosion. The upper abutments are topped off with the Fort Payne Formation, a series of argillaceous limestone, calcareous shale, and thin beds of cherty limestone (USACE, 1940).

**Karst**

Although no faulting is present at the site, relatively close centered jointing is prevalent and follows two well

![Figure 2. Wolf Creek Dam axis in profile with foundation geologic stratigraphy.](image-url)
defined joint sets. Solutioning by water infiltration along these near vertical joints and bedding planes of the limestone has occurred over millions of years, particularly across the valley. The exploration programs and foundation preparation for the masonry dam and cut-off trench revealed a rock foundation riddled with solutioned features ranging in size from a few inches (centimeters) to 40 vertical ft (12.2 m) along the joints (Figure 3). These karstic solution features were found primarily within the Leipers Limestone Formation, which was susceptible to higher groundwater flow being the uppermost bedrock across the valley floor and the formation through which the river channel flows. The interconnected nature of the karst system has been well documented by the foundation preparation during construction, through the various exploration programs, pool responsive piezometers, and wet areas downstream of the dam. These open features within the rock mass have been variably filled or partially filled with residual and alluvial deposits of sands, silts, and clays.

**Performance History**

The reservoir was first impounded in 1950. Seepage issues beneath the dam were first indicated in 1962 by wet areas near the downstream toe toward the right abutment. By 1967 the area had become too wet to mow, and in August of that year a small sinkhole formed near those wet areas. That fall muddy flow began to exit 150 ft (46 m) downstream of the powerhouse into the tailrace of the dam, and the following spring two sinkholes formed near the switchyard. Piezometers were soon installed in the area and dye tests were conducted, which indicated that seepage was occurring under the dam and by-passing the upstream cut-off trench through a system of solution features in the foundation rock that ran generally perpendicular and parallel to the dam axis. This seepage was piping materials from these solution features and transporting embankment material that collapsed into the features.

Emergency grouting of the subsurface solution features began in April 1968 and continued through 1970, installing a series of grout lines within and beneath the downstream embankment where it wraps around the masonry dam. This quick action likely prevented a breach of the dam, but was not considered a permanent fix due to the karst foundation. Between 1975 and 1979 a concrete diaphragm wall was installed across the left portion of the embankment from the crest of the dam down to competent bedrock, to cut-off seepage paths through the karst solution features of the Leipers Formation. A smaller additional wall was also installed between the switchyard and the tailrace to prevent the exit of material from beneath the dam.

The barrier wall was expected to significantly drop the water levels in the downstream piezometers, but only a slight reduction occurred. It was then predicted that over time the water levels would drop, but instead the measurements in the downstream piezometers began to rise over the years. A surface elevation monument installed in 1981 near the embankment and masonry dam interface showed continued settlement of the embankment, with an increase in settlement rate after 1997. Downstream wet areas near the right abutment persisted through the remediation efforts, but those near the left end of the embankment largely disappeared after the grouting and barrier wall construction of the 1970s. Over time the wet areas returned, and those near the right abutment steadily grew in extent from 1990 until they reached a maximum extent in the spring of 2004. In 2002 and 2003 borings drilled into the embankment downstream of the barrier wall encountered zones of soft saturated clay several feet thick at the base of the dam material.

It was decided that additional remediation was necessary to reduce seepage below the embankment dam to maintain safe dam operation. Between 2007 and 2013 an additional concrete barrier wall was constructed in the embankment upstream of the first wall. The new wall was extended up to 75 ft (22.9 m) deeper into bedrock, below the Leipers-Catheys contact, and extended nearly across the entire embankment to the right abutment. New grouting reached depths at least 50 ft (15.2 m) below the new wall and into the right abutment. A deeper extension was also added to the subsurface wall between the switchyard and the tailrace. Exploratory grouting below the masonry dam resulted in small grout volume takes, Figure 3. Karst channels and caves uncovered and cleaned out during excavation of upstream cut-off trench.
so no additional remediation was conducted within the masonry dam foundation (USACE, 2014).

**Masonry Dam Construction**

The seemingly superior performance of the masonry dam foundation over the embankment dam foundation to prevent seepage and safety issues is largely due to the extensive foundation preparations that were conducted prior to the construction of the dam.

The concrete masonry portion of Wolf Creek Dam is divided into 37 primary monoliths, all of which are founded on competent bedrock. The monoliths are grouped into four sections which include 1) the Left Non-Overflow Section (Monoliths 1-7) along the steeply sloped left abutment, 2) the Spillway Overflow Section (Monoliths 8-18) within the Cumberland River channel, 3) the Power Intake Section (Monoliths 19-26) on the shallow right bank of the channel (the powerhouse is located immediately downstream of this section), and 4) the Right Non-Overflow Section (Monoliths 27-37) at the embankment wrap-around area (Figure 1). To reach competent bedrock it was necessary to excavate down to the Lower Leipers Formation or Upper Catheys Formation, which both consist of limestone and interbedded shale. Weathering, karst, and solution features were largely concentrated within the Leipers Formation across the valley floor within the primary groundwater flow regime, but were less prevalent within the abutments.

**Site Investigation and Preparation**

Care was taken during construction of the masonry section to remove all overburden and weathered or deteriorated bedrock. Estimates of depths to competent rock were made based on early 1930s site investigation borings made on 100 ft (30.5 m) centers. Then after the removal of the overburden, additional boreholes at least 16 ft (4.9 m) deep were drilled into the bedrock both parallel and normal to the dam axis on 20 ft (6.1 m) centers covering the entire exposed concrete dam area. Based on the data collected from these holes final excavation depths were determined so that all soft beds and solution channels would be removed from underneath the dam. During bedrock excavation, which was accomplished primarily by blasting and power shovels, the crews continued to look for issues, and some additional borings were ordered for problem areas. These further investigations led to decisions to deepen excavation for Monoliths 24-19, dig out the caves in Monoliths 37 and 36, add more grout than had been originally planned, and other special fixes to ensure the integrity of the bedrock foundation. When large solution features were encountered within the limestone they were dugout, widened, cleaned, and filled with concrete. Final rock preparation consisted of barring and picking, cleaning with air and water pressure hoses, and brooming in a ½-inch layer of grout just prior to the placement of the concrete. After the construction of the grout gallery near the upstream axis of the dam, angled pressure grouting was placed and drains were installed into the foundation bedrock (USACE, 1952).

**Overburden Removal**

Overburden at the dam site consisted largely of sandy and silty alluvial river deposits across the valley, or thin colluvium layers on the abutment slopes. To prepare the monolith foundations all overburden and alluvium were removed by either hydraulic dredging, a dragline, a clamshell, or diesel power shovels (Figure 4). Depths of overburden averaged about 25 ft (7.6 m) for Monoliths 37-27, but increased to an average of 40 ft (12.2 m) for Monoliths 26-14 since the top of bedrock was deeper and closer to the river channel. Overburden depths only averaged 6 ft for Monoliths 13-1 since channel flow limited sediment deposits on bedrock within the river and the steep slopes on the left abutment prevented the deposit of thick layers of sediment, leaving the bedrock exposed or narrowly covered.

The dredge operated from the spring to fall of 1946, between the areas of Monolith 29 to 18. It was then used in the summer and fall of 1948 to remove a sand bar known as Cooper’s Island from the river, and then used in 1949 to fill cells for Cofferdam No. 2. The power shovels, clamshell, and dragline removed all the rest of the overburden, commencing on the right side near the embankment, and then working in various stages near the river channel; as cofferdams were moved, monoliths were constructed, and areas became available for work to proceed (USACE, 1952).

**Bedrock Excavation**

Bedrock excavation was conducted primarily by blasting. As faces were established, new shot holes would be drilled on 3 ft (0.9 m) centers behind the face. Shot hole depths were generally 6 to 8 inches (15.2 to 20.3 cm) above an established bedding plane, but did not extend over 8 ft (2.4 m) deep. Forty percent dynamite was the blasting substance, used in the proportion of 0.75 pounds per cubic yard (0.44 kg/m³) of rock to be removed. Blasting in delayed series was initially tried but quickly abandoned since primary blasts sometimes severed the connections to the secondary blasts, resulting in exploded dynamite and a hazardous situation. Power shovels and a clamshell were used to load the rock onto dump trucks. During the final foundation clean-up operations, rock removal was being conducted by hand labor using picks, shovels, pry bars, and high pressure hoses. All
rock that was not “firmly bedded” had to be removed from the horizontal and vertical surfaces of the foundation. During this operation waste material was loaded into skip pans and hauled away by the cableway that had been constructed above the site (USACE, 1952).

**Non-Overflow Right Bank Section (Monoliths 37-30)**

On 30 April 1946 work began on the foundation of the right end of the concrete dam. Excavation within this section was started at each end, with one group working between Monoliths 37-35 and another working between Monoliths 31-30. Rock removal was more difficult than anticipated due to the irregular patterns of soluble limestone channels and mud filling, especially at Monoliths 37-35. At that end two major mud-filled cavernous solutions were uncovered which were a continuation of the solutions beneath the embankment section that the upstream cut-off trench followed. Stability of the bedrock and the hazards it posed to the workers became a major concern, so both groups proceeded cautiously by conducting only shallow blasts; hoping to shoot out the smaller solution channels and reveal the larger and deeper solution caverns. The patterns eventually revealed themselves as two large channels extending across Monoliths 37, 36, and part of 35. At Monolith 35 the two channels converged into one more narrow channel that eventually was revealed to extend across Monoliths 35, 34, 33, 32, 31, 30, 29, 28, 27, and a portion of Monolith 26. As large channels and caverns were revealed, the side walls were cut back until only satisfactory bedrock remained (the channels were wide enough for a clam-shell bucket to reach the bottom), and the mud and loose rock was removed from the bottom until the channel was completely cleaned out (Figure 5). About 40% of the material removed was mud and 60% rock. Eventually
these channels were filled with concrete. Rock excavation in this area was completed February 1947.

**Power Intake Section (Monoliths 29-19)**

Dredging began in the area between Monoliths 29-19 on 15 April 1946, removing up to 51 ft (15.5 m) of overburden. On 12 July 1946 the dredge was moved downstream and excavation of the final 5 ft (1.5 m) of overburden along with the bedrock commenced with a power shovel and trucks. At Monolith 22 another solution channel was uncovered, and a clamshell was brought in to help define it. It was discovered to extend from Monolith 24 – 18, be mud filled, of irregular pattern, and approximately 8 ft (2.4 m) deep. This prompted additional test drilling along the solution channel to determine the soundness of the surrounding rock. Much of the rock was determined to be inferior, so the test holes became blast holes and the inferior rock was removed taking along with it the solution channel. Investigations in the area continued with drilling several more 6 inch diameter holes and two 30 inch (76.2 cm) Calyx holes, one in Monolith 20 and the other in Monolith 22. These investigations determined that a continuous inferior bed existed at elevation 523.0 ft (159.4 m) above sea level (NGVD29) under the entire area from Monoliths 24 to 19. It was decided to remove all bedrock in that area to elevation 523.0, which was 12 ft (3.7 m) lower than the previously determined elevation of 535.0 (Figure 6). Removal of the additional rock began on 1 December 1946, and was completed on 10 April 1947. In Monolith 20 a flood on 3 January 1947 displaced foundation rock near the Calyx hole, so addi-
tional grout was placed without pressure in a radial pattern across Monolith 20 prior to the placement of concrete (Figure 7). Initial concrete cover of the bedrock was completed on 20 April 1947.

Bedrock excavation in the area of Monoliths 29-25 encountered few if any solution channels, so the excavation was not required to extend deeper. It was even determined to leave in place an upstream rock ledge from Monoliths 29 to 25 (Figure 8). Though the rock in the ledge was determined to be competent, there was some concern for seepage pathways eventually working through the bedding planes to the base of the dam. Contact grouting was pumped into bedding planes of con-

Figure 6. Photograph of the cleaned foundation of Monolith 21. Remnants of the solution feature that was removed are visible on the floor and on the downstream wall.

Figure 7. Photograph of the cleaned foundation of Monolith 20. Remnants of a solution feature is visible on the floor, as are the grout holes placed in a radial pattern from the Calyx hole.
cern along the vertical faces of the ledge (Figure 9) after it had been cleaned and all loose rock removed. Excavation and initial concrete placement was completed by 10 April 1947.

**Spillway Section**

*(Monoliths 18-9)*

Foundation excavation in the area of Monoliths 18-14 began on 3 November 1947. The proximity to the river channel required the partial removal and reconstruction of the cofferdams in order for work in the area to continue. Excavation resulted in few incidents, since weathered rock in the Liepers Formation had likely already been removed by river channel erosion. Up to 48 ft (14.6 m) of overburden had to be removed to reach bedrock. After the removal of the overburden rock excavation was conducted in a single shallow lift, except in the downstream bucket sections where the design required excavation down 6 ft (1.8 m) lower than the abutting monoliths. It was also within the bucket sections that the solution feature from Monolith 24 cut across these monoliths, but it was completely removed by the lower excavation depth in the bucket sections. A shallow sump was excavated below the grout gallery in Monolith 18.

**Spillway Section within River Channel**

*(Monoliths 13-9)*

To begin foundation work within the Cumberland River channel required additional removal and reconstruction of the cofferdams to redirect flow away from the area.

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**Figure 8.** Photograph from above showing the initial placement of concrete monoliths on the foundation, and rock benches that were not excavated.

**Figure 9.** Photograph of the contact grout pipes placed on the vertical face of the rock ledge left in place at Monolith 25. The pipes eventually were connected to the grout gallery so that additional grout could be placed as necessary after construction.
Once dewatered, about 3 ft (0.9 m) of overburden was removed containing boulders, gravel, and sand. A cableway was employed to place excavation equipment within the cofferdam, and work commenced in June 1949, but was stopped on 1 July 1949 due to a general strike. The equipment was removed and the area was re-flooded during the duration of the strike. Work recommenced on 23 September 1949, and all overburden was removed by 1 October 1949. The exposed rock showed a mud-filled solution channel present in the downstream bucket section across Monoliths 13-11. It was determined to be about 8 ft deep and varied in width between 10 and 25 ft (3.0 and 7.6 m). It was determined to remove the entire solution channel. Rock excavation began on 28 September 1949 and was completed on 15 November 1949.

**Non-Overflow Left Bank Section (Monoliths 8-1)**

Since the bedrock was shallow and/or exposed along the left abutment foundation work began early there on 5 April 1946 commencing high on the slope at Monolith 2; removing overburden down the slope towards the river channel. With the overburden and some weathered rock removed, foundation preparation then began at the bottom of the slope at Monolith 8 to better constrain the work limits of the area. Initial concrete placement at Monolith 8 began on 6 December 1947. Work then commenced again from the top of the slope, working downward, by removing rock and cutting in stair step benches on which the laborers could work and concrete could be placed. The left abutment foundations were primarily limestone except for the Chattanooga Shale at the base of Monolith 2. Monoliths were excavated sufficiently deep to maintain a minimum distance of 10 ft (3.0 m) below the top of original rock so that rock exposed to surficial weathering was removed. Between Monoliths 7 and 5 the shallow step-ups were replaced with deeper, larger step-ups (Figure 10) to avoid the Monoliths being founded on structurally weak beds that exhibited conchoidal structure during excavation. Large solution features were not present in the left abutment limestone. Contact grout systems were installed on the vertical rock faces prior to the placement of concrete, to close off bedding planes and joints along the abutment (Figure 10). The systems were left in place and connected into the grout gallery so that future grouting in the abutment could occur.

**Grouting and Drains**

After the foundations were excavated, cleaned, and fully prepared it was decided to drill supplementary grout holes along the dam axis to fill any extensive subsurface openings and crevices, and confine the high pressure grouting that would occur later. The 2 inch (5.1 cm) diameter holes were drilled to a depth of 25 ft (7.6 m) and angled 22.5° from vertical towards the left abutment. The drill holes were washed and cleaned with water, and then compressed air was used to remove all the water. Grout was then poured into the holes without pressure until refusal occurred.

To prepare for the high pressure grouting, 3 inch (7.6 cm) diameter steel casing pipes were installed at 5 ft (1.5 m) centers along the axis of the dam during the foundation preparations (Figure 10). These pipes were angled 7° towards the left abutment, and were positioned such that as they would emerge in the floor of the grout gallery near the dam axis at the base of the masonry dam (Figure 11). Once the grout gallery floor was constructed, grout holes were drilled through the casing and into

![Figure 10. Photograph showing the cut “stair steps” into the left abutment at Monolith 5 and 6, with the pipes in place for the contact grout system on the vertical faces and the casing pipes in place for the grout gallery drains and grout holes.](image-url)
the foundation bedrock. High pressure grout was then pumped into the foundation.

During the foundation preparation, 5 inch (12.7 cm) diameter casing pipes were placed downstream of the grout pipes, angled 12.5° from vertical in the downstream direction. After the high pressure grouting was completed within 100 ft (30.5 m) of these downstream pipes they were drilled to install drains from the foundation into the grout gallery, providing relief points for future pressurized water seeping under the dam foundation to escape without cracking or damaging the structure (USACE, 1952).

**Conclusion**

The original construction photographs and foundation reports for the concrete section of Wolf Creek Dam instilled confidence that the designers and builders of the monoliths took adequate measures to ensure that all the monoliths were founded on competent bedrock. These measures include: extensive borehole investigations both prior to and during excavation; efforts to locate, delineate, remove, and clean all karst solution channels; the removal of all loose rock; grouting in the foundation and vertical faces; the large stair-step faces on the left abutment; the extended excavations to remove soft beds; the final manual cleaning of rock surfaces; and the careful documentation of foundation preparations. These measures do not guarantee that seepage issues will not develop under the concrete dam over time, but they do show that the monoliths were originally founded on competent bedrock and that future seepage issues are either unlikely or will be significantly inhibited by the preparation made to the foundation prior to the construction of the concrete monoliths.

**References**


**Figure 11.** Typical cross-section of non-overflow monolith, highlighting the location of the grout gallery.
HYDROCOMPACTION CONSIDERATIONS IN SINKHOLE INVESTIGATIONS

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Abstract
The cause of ground settlement is a significant concern in sinkhole investigations where the potential for shallow and deep-seated instability in the subsurface is a major focus of the investigation. Complicating the investigation is the occurrence of hydrocompaction of surficial soils caused by introduction of large amounts of surface water particularly from improper maintenance of rainfall runoff. This condition is usually followed by the subsequent loss of soil moisture during dry periods. This manuscript will discuss how hydrocompaction plays a role in the analysis of settlement in the investigation of sinkhole loss and how one can distinguish between hydrocompaction settlement and deep-seated settlement (note, that hydrocompaction is one of many factors that can account for settlement of structures). It will consider the effects of soil density as it impacts hydrocompaction in the investigation of building distress. Also discussed are the results of laboratory tests of simulated hydrocompaction on fine sand samples in loose and dense states. In one of the tests, the formation of a collapse sinkhole occurred at the end of the test. Photographs depicting the sequence of soil failure are attached at the end of this paper.

Introduction
Most of us have seen the effects of hydrocompaction in the settlement of the soil surface in a flowerpot. In this instance settlement occurs during the extended periods of soil saturation followed by periods of soil drying. These saturated and unsaturated conditions are similar to what occurs to the soil around a building. Other, more extreme examples of how hydrocompaction can be a factor in settlement, is seen in the greater than 30 feet (9.1 m) of land subsidence that has occurred during a period of over 50 years in the aeolian soils of the San Joaquin valley in California. In this instance, “aquifer-system compaction and hydrocompaction settlements have significantly lowered the land surface since about the 1920s” (Galloway and Riley, 1999). Settlement results from dewatering and the consequential increases in effective soil weight. Hydrocompaction is one component of the settlement we find in the shallow soils near buildings.

Hydrocompaction

General
Hydrocompaction also referred to as hydro-collapse is a process of settlement and resulting volume change that occurs in fine sand with minor amounts of silt and clay. The term hydrocompaction will be used in this paper to describe this process. Hydrocompaction is driven primarily by the infiltration of water into the soil fabric. During wet periods, the continuing infiltration of water into the soil fabric produces a redistribution of soil particles causing the soil to settle while during dry periods settlement occurs (although to a lesser extent in west central Florida) because of an increase in effective stress (Figure 1).

Soils susceptible to hydrocompaction are generally geologically immature soils that have high void ratios and low densities; they can be aeolian deposits or residual soils. These soils are found throughout the United States and have notably caused significant damage in areas where large amounts of water entered the subsurface from leakage in anthropogenic projects. However, the discussions in this paper will be limited to hydrocompaction of loose fill typically found under and around buildings in which sinkhole investigations are being conducted.

Mechanism
Soils subject to hydrocompaction have one characteristic in common; they have weak structural and chemical bonds between particles. Water infiltrating the soil fabric causes a loss in these bonds. This causes the soil particles to compress in the soil column to more stable positions. As this process continues over time, soil-supporting portions of a structure is lost and a net decrease in soil
Laboratory Analysis of Settlement

To aid in understanding the potential and magnitude of hydrocompaction found in various sinkhole investigations, a series of laboratory tests were performed to simulate conditions found at representative sites. The tests consisted of measuring settlement during application of water to sand samples contained in a polycarbonate (lexan)-lined seepage tank. The tank measured 16 inches (40.6 cm) high by 15¾ (40.0 cm) inches wide by 5 ⅜ (8.6 cm) inches deep and was constructed as shown in Photo 1.

The seepage tank was filled to a height of approximately 13¼ inches (33.7 cm) with fine sandy soil typical of that found in west-central Florida. Soil samples were tested in the loose and dense states. The tests were performed with the application of a continuous supply of water entering at the top of the tank to saturate the soil and maintain a constant state of saturation. Flow into the tank was regulated so that inflow was approximately equal to outflow. An observation well was installed to assure there was no ponded water in the tank; the observation well consisted of a ¼ inch (0.64 cm) neoprene tube mounted in an aluminum channel section. Settlement of the sand surface was measured with an extensometer recording movement to the nearest thousands of an inch. Measurements were recorded at intervals appropriate to establish the time settlement curves shown in Figures 2 and 3. All tests were continued until settlements had reached an essentially constant rate of elastic change.

This test was designed to approximately simulate the settlement that would occur in loose fill that is typically placed around and possibly under building constructed in south-central Florida. The test was run with no strength occurs in soil supporting the footing. Ultimately, differential settlement may occur as footings must span larger areas of soil to support load. This is a phenomenon seen in many sinkhole investigations.

The repeated saturation and drying of these soils subjects the soil to repeated cycles of tensile and compressive forces that exacerbates settlement. Also the movement of water through the soil causes soil particles to move downward due to erosion (Shlemon, 2004) as the water percolates into the ground surface. Loose fine to medium sands are most susceptible to this condition. As mentioned, a good example of this phenomenon is the settlement that occurs in a flowerpot from the alternate saturated and unsaturated conditions that occur. Over a period of months and sometimes years the surface of the soil undergoes settlement of from 3 to 10% of the total depth of loose soil (laboratory and field observations). The rate at which this settlement occurs depends on the amount of fines in the soil. The greater the silt and clay content the slower the rate of settlement.

Figures 1A and 1B (after Cassagrande, 1932), provides a graphical representation of the soil microstructure before and after the addition of water. Figure 1A is an idealized view of how soil particles appear before the application of water while Figure 1B provides a view of how the soil particles may rearrange after water travels through the soil fabric.

The settlement associated with Figure 1B results not only from weak structural bonds but also from weak chemical bonds between particles. As water moves through the soil fabric, capillary tension between particles is lost and there is a weakening of clay bonds between particles. This causes the soil particles to compress resulting in settlement at the surface.

Figure 1. A. Microstructure before application of water. B. Microstructure after application of water.
from the settlement verses time curves in Figures 2 and 3, it is apparent that two distinct curves define the settlement in each sample. The initial, steeper portion of the curve represents the inelastic component of hydrocompaction settlement. This is the settlement that is characterized by relatively rapid movement of sand as shown in Figure 1A. The second part of the curve represents the elastic component of hydrocompaction shown in Figure 1B.

The distinction between the inelastic and elastic portions of settlement is readily apparent from the abrupt change of slope shown in the time verses settlement curves shown in Figures 2 and 3. This change in slope represents the point when most soil particles have shifted to more stable positions in the soil mass as shown in Figure 1B. The trend is most apparent in loose soil because of the greater ease with which particles can transition to the more stable state characterized by elastic settlement.

A good illustration of settlement associated with hydrocompaction and the subsequent settlement associated with the development of a sinkhole is shown the photographs in Figure 4. In this hydrocompaction experiment, it was found after hydrocompaction settlement was essentially complete, additional settlement occurred associated with the development of a cover collapse sinkhole. The experiment was continued while a small void developed in the sand matrix as soil particles moved to lower positions in the soil section. With the passage of time, at a constant rate of seepage (inflow was approximately equal to discharge through the bottom of the section), the void continued to enlarge until a cover collapse sinkhole developed in the surface of the test section (Photos 1 through 6 in Figure 4).

Ok, so what have we shown? We have shown that hydrocompaction settlement is, for the most part, non-

### Test Results

As one would expect, we have found the loose sand had the greater hydrocompaction settlement. Settlement begins with the first application of water and continues at a relatively rapid rate until soil particles move to more stable positions in the soil matrix as shown in Figure 1. After reaching this state, only elastic settlement occurs at a greatly reduced magnitude (see Table 1).

Based on the test results shown in Figures 2 and 3, it is seen that the less dense the soil (high void ratio) the greater the settlement. Table 1 provides a summary of the test results for the two densities corresponding to a loose and dense soil.

<table>
<thead>
<tr>
<th>No.</th>
<th>Soil State</th>
<th>Moisture Saturated (wSat)</th>
<th>Sp Gravity</th>
<th>Void Ratio (e)</th>
<th>Rate of Settlement (inches per foot)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Loose</td>
<td>20.8%</td>
<td>2.61</td>
<td>0.54</td>
<td>0.26</td>
</tr>
<tr>
<td>2</td>
<td>Dense</td>
<td>17.5%</td>
<td>2.61</td>
<td>0.46</td>
<td>0.012</td>
</tr>
</tbody>
</table>

### Analysis

From study of the settlement verses time curves in Figures 2 and 3, it is apparent that two distinct curves define the settlement in each sample. The initial, steeper portion of the curve represents the inelastic component of hydrocompaction settlement. This is the settlement that is characterized by relatively rapid movement of sand as shown in Figure 1A. The second part of the curve represents the elastic component of hydrocompaction shown in Figure 1B.

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Photo 1. Void begins with partings in sand at course zones.

Photo 2. Partings enlarge to form voids.

Photo 3. Voids coalescence to form one large void.

Photo 4. Void enlarging to a more stable configuration.

Photo 5. Fully developed void. Note the top geometry has remained constant while sides have expanded to relatively stable position.

Photo 6. Failure of void roof.

Figure 4. Development of a cover collapse sinkhole.
The depth affected by hydrocompaction has additional significance in sinkhole investigations as seen from the wording of the Florida sinkhole statute §627.706 (Florida Statutes 2014). The statute states that sinkhole activity is present if settlement or weakening of earth supporting the building has occurred (see Figure 5). In some instances, investigators have considered low N-values caused by hydrocompaction as evidence of sinkhole activity (Zisman, 2013). The distinction between hydrocompaction caused by surficial conditions verses raveling caused by deep-seated conditions can, in

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**Figure 5.** “Sinkhole activity” according to the Florida statute.
many cases, be distinguished from the change in N-value that occurs with increasing depth. That is, there must be some evidence that low-density material is present at depth and is related to movement of soil “into subterranean voids created by the effect of water on a limestone or similar rock formation.” An obvious distinction would be the change in soil density that occurs at increasing depth below the soils affected by hydrocompaction.

By in large, the greatest damage from hydrocompaction is in buildings constructed without proper compaction of the sandy soils prominent in our area. Deposition of rainwater runoff next to the walls of buildings not equipped with gutters and downspouts to direct water away from the building results in a condition favorable to hydrocompaction settlement.

Settlement Theories
Although not strictly true, it is acceptable to assume that water and soil grains are incompressible; therefore, the only way settlement can occur in a soil is by movement of the soil grains or collapse of the soil structure. Accordingly, in classical analysis of settlement we generally consider three types of settlement: 1) immediate settlement, 2) consolidation settlement and 3) secondary settlement. However, in hydrocompaction two mechanisms for settlement are present: inelastic and the elastic. The elastic component can be determined by classic methods related to void ratio while the inelastic component can be determined by empirical methods related to void ratio change. Table 2 provides a comparison.

Discussion and Conclusions
1. From the testing completed, we find that two components of settlement have occurred; one inelastic, the other elastic. The inelastic settlement is characterized by a steep curve while the elastic settlement is characterized by the flatter curve indicative of elastic settlement. Depending on soil density, the magnitude of inelastic settlement is at least an order of magnitude greater than elastic settlement.

2. The location and depth of soils subject to hydrocompaction can be determined by a use of a handheld penetrometer. In general soils having a penetrometer reading of less than 25 kg/sq cm are susceptible to hydrocompaction. The location for penetrometer testing can be expedited by first testing with a probe rod to determine the locations loose/soft soil material.

3. A preliminary estimate of the magnitude of hydrocompaction settlement can be estimated from the value in Table 1 of 0.37 inches / foot (0.94 cm/30.5 cm) of thickness.

<table>
<thead>
<tr>
<th>Settlement Mechanisms</th>
<th>Classic</th>
<th>Hydrocompaction</th>
<th>Compaction</th>
</tr>
</thead>
<tbody>
<tr>
<td>(Driven by static and dynamic loading conditions under various moisture conditions)</td>
<td>(Driven essentially by movement of water)</td>
<td>(Driven by static and dynamic loading under non-saturated water conditions)</td>
<td></td>
</tr>
<tr>
<td>Phase 1. Immediate or elastic</td>
<td>Phase 1. Immediate or elastic</td>
<td>Phase 1. Immediate or elastic</td>
<td></td>
</tr>
<tr>
<td>A. Settlement is determined from elastic theory</td>
<td>A. Immediate settlements are inelastic and can best be approximated from empirical data related to void ratio</td>
<td>A. Settlement is determined from elastic theory</td>
<td></td>
</tr>
<tr>
<td>B. Occurs in all types of soil because of elastic compression</td>
<td>B. Occurs in geologically immature fine sandy soils with less than about 10% silt and clay</td>
<td>B. Occurs in all types of soil because of elastic compression</td>
<td></td>
</tr>
<tr>
<td>Phase 2. Consolidation</td>
<td>Phase 2. Consolidation</td>
<td>Phase 2. Consolidation</td>
<td></td>
</tr>
<tr>
<td>A. Occurs by the process of expulsion of water from soil matrix</td>
<td>A. Consolidation is not a factor in hydrocompaction.</td>
<td>A. Consolidation can be a factor in settlement.</td>
<td></td>
</tr>
<tr>
<td>B. Settlement is determined from the theory of consolidation</td>
<td>B. Not applicable</td>
<td>B. Settlement is determined from the theory of consolidation</td>
<td></td>
</tr>
<tr>
<td>A. Excess pore water pressure is zero</td>
<td>A. Excess pore water pressure is zero</td>
<td>A. Excess pore water pressure is zero</td>
<td></td>
</tr>
<tr>
<td>B. Creep settlement occurs from deformation of soil particles to load.</td>
<td>B. Creep settlement occurs from deformation of soil particles to load.</td>
<td>B. Creep settlement occurs from deformation of soil particles to load.</td>
<td></td>
</tr>
</tbody>
</table>

Table 2. Settlement theories.
4. Photographs are shown in Photos 1 through 6 in this paper of a sinkhole that developed in the settling tank when the water was allowed to continue infiltration into the sand soil ultimately resulting in a cover collapse sinkhole.

5. Hydrocompaction occurs in loose sandy soils common in west-central Florida. Buildings constructed without gutters and downspouts to direct water away from the building are vulnerable to hydrocompaction settlement in loose soil.

References
Florida Statutes. 2014. Chapter 627.706 Sinkhole insurance; catastrophic ground cover collapse; definitions.; Division of Law Revision and Information Services, Tallahassee, Florida 32399.