Hydrologic Controls on Salinity in Mangroves and Lagoons

by

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DEDICATION

This dissertation is dedicated to my family—especially my sisters, who have supported and encouraged me every step of the way.
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Although this is ultimately my dissertation, this work could not have been possible without the contributions of many people, especially my committee members. First and foremost is Mark Rains who, aside from guiding all of my research efforts, had the patience to give me what I didn’t even know I needed at first…the time to believe in myself and my abilities. Sarah Kruse has been a wonderful mentor and a constant source of encouragement. Mark Stewart and H.L. Vacher have been enthusiastic supporters and greatly expanded my knowledge of geological concepts. Thomas Crisman had the graciousness to step in and fill a void late in the game. My family, friends, and fellow graduate students all helped to positively shape my doctoral experience. Many students and volunteers have aided in data collection over the years; this work simply would not have happened without the benefit of their time. Beth Fratesi has worked diligently to help me create, improve, and format the majority of my figures. During the course of my doctoral research I have received financial support from the United States Geological Survey, The Smithsonian Institution, The National Science Foundation, The University of South Florida Graduate School, The Great Basin Institute, and the Patel Center for Global Solutions.
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Hydrologic Controls on Salinity in Mangroves and Lagoons

Christina Elaine Stringer

ABSTRACT

This dissertation explores the hydrologic controls on salinity within mangroves and lagoons at sites in Florida and Mexico. The main objective of this research is to better understand hydrologic controls on mangrove ecosystem structure and develop ideas that will be useful to land managers attempting to regulate and conserve these critical habitats. This study was conducted at sites in Ft. Pierce, FL and Costalegre on the central Pacific coast of Mexico.

We examined controls on water levels and salinity in a mangrove on a carbonate barrier island along the Indian River Lagoon, east-central Florida. Spectral analysis of water levels showed that mangrove groundwater levels are not tidally influenced. Salinities vary spatially, with values of ~10 in upland environments to ~75 psu in irregularly-flushed mangroves. Water chemistry indicates that water salinities are largely controlled by enrichment due to evapotranspiration. An electrical resistivity survey showed that the freshwater lens is restricted to uplands and that hypersaline waters extend deeply below the mangrove. These results indicate that evapotranspiration lowers water levels in the mangrove, which causes Indian River Lagoon water to flow into the mangrove where it evapoconcentrates and descends, forming a thick layer of high-salinity water below the mangrove.
Spatial variability of terrain conductivity in the Ft. Pierce mangrove varied under two hydrologic management regimes, breached rotational impoundment management and rotational impoundment management. The difference in coefficient of variation (ΔCV) between the breached RIM and RIM data was calculated to examine spatial variability in both the shallow and deep layers. A null-hypothesis model was employed to examine the statistical significance of the ΔCV results. The average water levels were -0.06 m amsl and 0.49 m amsl during the breached-RIM and RIM regimes, respectively. The average shallow (EM31) layer terrain conductivity shifted slightly from 1868 mS m\(^{-1}\) to 1825 mS m\(^{-1}\) after the alteration in management regime, yet the standard deviation of these averages decrease from 656 mS m\(^{-1}\) to 216 mS m\(^{-1}\). The average deep (EM34) layer terrain conductivities were 328 mS m\(^{-1}\) and 255 mS m\(^{-1}\) during the breached-RIM and RIM regimes, respectively. The temporal ΔCVs were 0.23 and -0.04 for the shallow and deep layers, respectively. The null-hypothesis model for the shallow layer illustrates that the difference in spatial structure is statistically significant. The deep layer ΔCV was not statistically significant. These results indicate that the transition from breached RIM to RIM resulted in changes to both the physical and chemical hydrologic character of the impoundment, especially in the shallow layer.

The second study sites were three mangrove communities along the central Pacific Mexican coast. Salinities varied by water type, with values of ~9 in La Manzanilla, ~17 in La Vena, ~33 in Barra de Navidad, ~0.4 in the fresh waters, and ~34 in the seawater. Sodium and Chloride concentrations and isotopic signatures, as well as salinity, were used as tracers in mass-balance mixing models to quantify estimates of relative fresh-water and seawater contributions to each site. La Manzanilla, a basin mangrove, had mean fresh-water contribution estimates of 63-84%. La Vena, a riverine mangrove, had fresh-water estimates of 39-51%. Barra de Navidad, a fringe mangrove,
had low fresh-water contributions of 0-5%. These results illustrates that the role groundwater plays in mangrove hydrodynamics is dependent on the site hydrogeomorphology.
CHAPTER 1:
INTRODUCTION

Mangroves are an important coastal ecosystem that provides numerous ecological functions, as well as goods and services, such as storm and erosion buffering and fish hatchery habitat. Mangroves cover ~180,000 km$^2$ of sheltered subtropical and tropical coasts between latitudes 24° north and south where mean annual temperatures are >20°C (Lugo 1990; Dawes 1998, Alongi 2002). Worldwide, mangrove ecosystems are just one example of a coastal wetland that is being lost quickly, with approximately 35% disappearing since the 1980s at a rate of 2.1% per year globally (Alongi 2002, Martinuzzi et al. 2009). In the Americas, this rate is as high as 3.6% per year with an approximately 38% loss in coverage area (Valiela et al. 2001, Valiela et al. 2009). These high rates of loss make mangroves the most threatened major coastal habitat in the world (Valiela et al. 2009). As in the case of other coastal habitats, anthropogenic activities are largely responsible for the loss and degradation of mangrove ecosystems (Alongi 2002; Gunawardena and Rowan 2005; Islam and Haque 2004; Vijay et al. 2005).

Hydrology is a master variable in coastal mangroves, directly or indirectly controlling ecosystem structure and function. Plant community composition can be controlled by salinity and/or water levels (Odum and McIvor 1990), while primary productivity is largely controlled by salinity, nutrients, and/or water levels.
The hydrology of coastal wetlands is complex, with climatically- and geologically-controlled fluxes of fresh water interacting with oceanographically- and geologically-controlled fluxes of sea water. Water levels and fluxes can be controlled by tidal variations, which can be propagated rapidly through surface-water and ground-water environments (Hughes et al. 1998; Ataie-Ashtiani et al. 2001). However, water levels and fluxes also can be controlled by episodic direct precipitation and associated surface-water and ground-water inflows during storms (Hughes et al. 1998) and/or steady ground-water inflow between storms (Drexler and De Carlo 2002). In some cases, water levels and fluxes also can be controlled by evapotranspiration, which can draw down water levels in the mangroves and draw water in from the surrounding surface-water and ground-water environments (Hughes et al. 1998).

Salinity and specific solute concentrations in coastal wetlands vary primarily as functions of water sources, and additionally as functions of evaporation and water-rock interaction. The salinity of direct precipitation is typically ~0.01 psu; the salinities of fresh surface water and ground water typically are ~0.03–0.60 psu; and the salinity of sea water is ~35 psu (Maidment 1993). Therefore, the degree of sea-water intrusion strongly controls salinity. However, solutes remain in solution when water evaporates, so salinity also may be greatly influenced by the rate at which water flows through the coastal
wetlands and the rate at which water evaporates while in the coastal wetlands (Hughes et al. 1998; Twilley and Chen 1998)

This dissertation explores the hydrologic controls on salinity within mangroves and lagoons at sites in Florida and Mexico. The main objective of this research is to better understand hydrologic controls on mangrove ecosystem structure and develop ideas that will be useful to land managers attempting to regulate and conserve these critical habitats. This dissertation is organized into three core content chapters, each of which is an independent manuscript prepared for journal publication. Chapter Two examines the role of evapotranspiration on the hydrology and salinity of a mangrove mosquito impoundment in Ft. Pierce, FL. This manuscript was published in *Wetlands* volume 30 in September 2010. Chapter Three utilizes electromagnetic measurements to examine the change in salinity spatial variability after an adjustment in management strategy at the Ft. Pierce mangrove mosquito impoundment. Before this manuscript is suitable for publication, another round of electromagnetic measurements is most likely necessary. This field work is scheduled for November or December 2010. After this information is added to the paper, the manuscript will be submitted to *Journal of Wetlands Ecology and Management*. Chapter Four looks at relative source water contributions in three different mangrove sites on the Central Pacific coast of Mexico, each with a different hydrogeomorphic character. This manuscript will be submitted to *Estuarine, Coastal, and Shelf Science* or *Journal of Wetlands Ecology and Management*.
References


CHAPTER 2:

CONTROLS ON WATER LEVELS AND SALINITY IN A BARRIER ISLAND MANGROVE, INDIAN RIVER LAGOON, FLORIDA

Abstract

We examined controls on water levels and salinity in a mangrove on a carbonate barrier island along the Indian River Lagoon, east-central Florida. Piezometers were installed at 19 sites throughout the area. Groundwater was sampled at 17 of these sites seasonally for three years. Head measurements were taken at the other two sites at 15-minute intervals for one year. Water levels in the mangrove are almost always lower than lagoon water levels. Spectral analysis of water levels showed that mangrove groundwater levels are not tidally influenced. Salinities vary spatially, with values of ~10 psu in uplands, ~30 psu in regularly-flushed mangroves, and ~75 psu in irregularly-flushed mangroves. Cation and anion concentrations and stable isotope compositions indicate that water salinities are largely controlled by enrichment due to evapotranspiration. A shore-perpendicular electrical resistivity survey showed that the freshwater lens is restricted to uplands and that hypersaline waters extend deeply below the mangrove. These results indicate that evapotranspiration lowers water levels in the mangrove, which causes Indian River Lagoon water to flow into the
mangrove where it evapoconcentrates and descends, forming a thick layer of high-salinity water below the mangrove.

Introduction

Mangroves, which cover ~240,000 km² of sheltered subtropical and tropical coasts between latitudes 24° north and south where mean annual temperatures are >20°C (Lugo 1990; Dawes 1998), provide numerous ecological functions as well as goods and services. Mangroves support estuarine and near-shore marine productivity, in part by providing critical habitat for juvenile fish and through the export of nutrient-rich water (McKee 1995; Rivera-Monroy et al. 1998) or plant, algal, or animal biomass (Zetina-Rejón et al. 2003). Mangroves also protect coastal habitats against the destructive forces of hurricanes, typhoons, and tsunamis (e.g., Kandasamy and Narayanasamy 2005; Granek and Rutenberg 2007; Alongi 2008).

Hydrology is a master variable in coastal mangroves, directly or indirectly controlling ecosystem structure and function. Plant community composition can be controlled by salinity and/or water levels (Odum and McIvor 1990), while primary productivity is largely controlled by salinity, nutrients, and/or water levels (Feller 1995; Chen and Twilley 1999; Suarez and Medina 2005; Cardona-Olarte et al. 2006; Lovelock et al. 2007). Typically, primary productivity can be greatly enhanced by additions of nutrient-rich water (McKee 1995; Rivera-Monroy et al. 1998).
The hydrology of coastal wetlands is complex, with climatically- and geologically-controlled fluxes of fresh water interacting with oceanographically- and geologically-controlled fluxes of sea water. Water levels and fluxes can be controlled by tidal variations, which can be propagated rapidly through surface-water and ground-water environments (Hughes et al. 1998; Ataie-Ashtiani et al. 2001). However, water levels and fluxes also can be controlled by episodic direct precipitation and associated surface-water and ground-water inflows during storms (Hughes et al. 1998) and/or steady ground-water inflow between storms (Drexler and De Carlo 2002). In some cases, water levels and fluxes also can be controlled by evapotranspiration, which can draw down water levels in the mangroves and draw water in from the surrounding surface-water and ground-water environments (Hughes et al. 1998). In many coastal mangroves, natural hydrologic patterns have been modified for a variety of reasons. On the east coast of Florida, for example, most of the mangroves have been impounded for purposes of mosquito control or the creation of habitat for migratory and resident waterfowl (e.g., Montague et al. 1987).

Salinity and specific solute concentrations in coastal wetlands vary primarily as functions of water sources, and additionally as functions of evaporation and water-rock interaction. The salinity of direct precipitation is typically ~0.01 psu; the salinities of fresh surface water and ground water typically are ~0.03–0.60 psu; and the salinity of sea water is ~35 psu (Maidment 1993). Therefore, the degree of sea-water intrusion strongly controls salinity. However, solutes remain in solution when water evaporates, so salinity also may
be greatly influenced by the rate at which water flows through the coastal wetlands and the rate at which water evaporates while in the coastal wetlands (Hughes et al. 1998; Twilley and Chen 1998).

This study is part of a broader series of investigations of the controls on species composition, primary productivity, and nutrient cycling in mangroves on a carbonate barrier island on the east-central coast of Florida. Our investigation is specifically focused on identifying and quantifying the controls on water levels and salinity, because there is little fresh water on the barrier island, infrequent tidal inundation over much of the mangrove, and greater than sea-water salinities in much of the mangrove. We hypothesized that evapotranspiration plays a major role in controlling water levels and salinities, by lowering the water table in the mangrove and creating a hydraulic gradient that draws surface water and ground water from the lagoon into the mangrove.

**Study Site**

The study was conducted in SLC-24, a mosquito control impoundment located at N27°33', W80°33' on the west shore of North Hutchinson Island. This barrier island is ~35-km in length and ~0.2–2 km in width, and is part of the system of barrier islands that bound the Indian River Lagoon, a 250-km estuarine system located on the east-central coast of Florida (Figure 2-1). The study site is ~9 km north of the Ft. Pierce Inlet, the northern-most of five channels that connect the Atlantic Ocean and the Indian River Lagoon.
The impoundment was developed in 1970 when a dike was constructed around an existing mangrove-dominated wetland (Rey et al. 1990; Rey and Kain 1991). The impoundment was hydrologically isolated from the lagoon between 1970 and 1985. Tidal exchange was minimally restored in 1985 when a culvert was placed through the dike to remove water that had accumulated during two tropical storms. The culvert was later closed and the impoundment was again hydrologically isolated until 1987 when the culvert was reopened and four additional culverts were added.

The climate is subtropical (mean annual, maximum, and minimum temperatures are approximately 23, 28, and 18°C, respectively). Annual precipitation is approximately 1340 mm and it is distributed over a November–May dry season and a shorter June–October rainy season.
North Hutchinson Island sediments vary in size and texture, with a high concentration of shell debris and a mean CaCO$_3$ concentration of 65% (Wang and Horwitz 2007).

Vegetation cover in the impoundment decreased from 75% to near 30% between 1970 and 1985 but began to recover following the installation of the culverts in 1987 (Rey et al. 1990). Black mangrove (*Avicennia germinans* (L.) L.) is the dominant mangrove, but red (*Rhizophora mangle* L.) and white (*Laguncularia racemosa* (L.) C.F. Gaertn.) mangrove also are common.

Buttonwood (*Conocarpus erectus* L.) also occurs, as do understory plants such as *Batis maritima* L., *Salicornia virginica* L., *Salicornia bigelovii* Torr., and *Borrichia frutescens* (L.) DC. The abundance, density, and sizes of mangroves vary across the impoundment from highly saline salt pans that have no mangroves or are fringed by dwarf black mangroves, areas that have dense and relatively short black mangroves, areas that have larger and more widely spaced black mangroves, to areas associated with open water where red mangroves are most abundant. Adjacent to and up gradient from the mangrove is a woody upland dominated by *Sabal palmetto* (Walter) Lodd. Ex Schult. & Schult. F. and *Schinus terebinthifolius* Raddi (Brazilian Peppertree). Therefore, for purposes of this study, five community types based on species composition and structure were identified: (1) salt pan, (2) sparse black mangrove, (3) dense black mangrove, (4) red mangrove, and (5) woody upland.
Methods

Physical Hydrology

Precipitation was measured continuously with a tipping-bucket rain gage (HOBO Event Rainfall Logger, Onset Computer, Pocasset, Massachusetts, USA). Stage in the Indian River Lagoon was modeled on 15-minute intervals using XTide 2 (http://www.flaterco.com/xtide/index.html), which was developed to predict tides and currents using an algorithm developed and used by the National Oceanic and Atmospheric Administration, National Ocean Service. The modeling program uses station-specific harmonic constants provided by the National Ocean Service and based on local tide gauge data, resulting in predictions that are accurate to ± 1 minute and ± 0.03 m of measured high and low tides assuming no episodic storm-related surge (e.g., surge related to expansion of water during periods of extremely low pressure during the passage of extratropical storms).

Piezometers were installed at 19 locations. At 17 locations, 2.5 cm inside diameter stainless-steel piezometers were installed in nests. These piezometers were screened over 15 cm beginning at 0.6, 1.2, and 1.8 m below the ground surface, with the exception of two sites in the woody upland that were screened over 15 cm beginning at 1.2 and 1.8 m below the ground surface. These piezometers were used for ground-water sampling. At two locations, 5.0 cm inside diameter PVC piezometers were installed. These piezometers were screened over 30 cm beginning at 2.0 m below the ground surface. Heads at these piezometers were measured on 15-minute intervals with pressure
transducers and data loggers (Model 3001 Levelogger, Solinst, Georgetown, Ontario, Canada). The pressure transducers had a 1.5 m range and a ±1 mm accuracy. Head data were compensated for changes in barometric pressure using barometric pressure data also measured on 15-minute intervals with a pressure transducer and data logger (Barologger, Solinst, Georgetown, Ontario, Canada).

Frequencies in the long-term water-level data from the piezometers and the Indian River Lagoon were determined using a fast Fourier Transform (FFT) algorithm in MATLAB (Version R2008A, Mathworks Inc., Natick, Massachusetts, USA). Prior to the FFT, a Tukey window function was applied to taper the data (Gubbins 2004). A Butterworth filter with a low-frequency cutoff of 33 hours was applied to the data to remove any long-term trends in the data. Mean daily head and stage cycles were also calculated using MATLAB. Daily evapotranspiration was calculated by quantifying daily fluctuations in the long-term water-level record following the procedure outlined in White (1932).

Daily evapotranspiration was computed as:

$$ET = S_y (24r \pm s)$$

where $S_y$ is the specific yield of the sediments ($10^{-1}$ in this setting), $r$ is the rate of water table rise between 00:00 and 04:00, and $s$ is the net change of the water level during the daily period (White 1932).

**Chemical Hydrology**

Water samples were collected and analyzed for temperature, pH, salinity, cations (e.g., Na$^+$, K$^+$, Mg$^{2+}$, Ca$^{2+}$), anions (Cl$^-$ and SO$_4^{2-}$), and stable isotopes
(e.g., deuterium and oxygen-18) in the wet and dry seasons in 2005 and temperature, pH, and salinity in the wet and dry seasons in 2006 and 2007. Water samples were collected from the rain collector, the Indian River Lagoon, the primary perimeter ditch inside the impoundment, and piezometers in which water was found. The rain collector consisted of 1-L bottle contained within a plastic cylinder with a funnel top. A ping pong ball was placed in the funnel to limit evaporation. Rainfall sampling was opportunistic and immediately following rainfall, with only five samples collected throughout the study period.

Approximately three volumes of water were pumped from each piezometer prior to the collection of samples for chemical analyses. Samples were pumped through 0.45 μm in-line filters (Whatman, Maidstone, Kent, England) directly into pre-cleaned, acid-washed HDPE sample bottles. Anion samples were acidified with 1 ml of nitric acid, and cation and anion samples were stored at the method-required range of 4 (±2) ºC prior to analyses. Stable isotope sample bottles were filled completely with negligible head space and sealed with Parafilm (American National Can, Chicago, Illinois, USA) to prevent the sample from equilibrating with ambient air.

Temperature, pH, and salinity of the surface-water and ground-water samples were measured in the field with a YSI 556 MPS (YSI Inc., Yellow Springs, Ohio, USA). Major cation and anion analyses were conducted at the University of South Florida Center for Water Analysis. Major cation concentrations were determined by inductively coupled plasma-emission spectroscopy following the EPA 200 method (Clesceri et al. 1998), and major
anion concentrations were determined by ion chromatography following the EPA 300 method (Clesceri et al. 1998). Analytical precision of the laboratory analyses were better than 1%.

Stable isotope analyses were conducted at the UC Davis Department of Geology Stable Isotope Laboratory. Deuterium analyses were performed using the chromium reduction method (Donnelly et al. 1991), while oxygen-18 analyses were performed using the carbon dioxide equilibration technique (Epstein and Mayeda 1953). Deuterium and oxygen-18 are reported in the conventional, delta notation ($\delta$):

$$
\delta = \left( \frac{R_{\text{sample}}}{R_{\text{standard}}} - 1 \right) \times 1000
$$

where $R$ is the ratio D/H or $^{18}O/^{16}O$ for deuterium and oxygen-18, respectively (Craig 1961). The resulting sample values of $\delta$D and $\delta$18O are reported in per mil (‰) deviation relative to Vienna Standard Mean Ocean Water (VSMOW) and, by convention the $\delta$D and $\delta$18O of VSMOW are set at 0‰ VSMOW (Gonfiantini 1978). Analytical precisions were ±1.0‰ and ±0.05‰ for $\delta$D and $\delta$18O, respectively.

**Evapoconcentration and Isotope Enrichment Modeling**

Evapoconcentration is the process by which solute concentrations increase as water evaporates and solutes are retained in the remaining solution. An evapoconcentration model with Na⁺ and Cl⁻ as conservative natural tracers was used to determine if evapoconcentration could explain solute concentrations within the five habitat types. The evapoconcentration model was run
encompassing both dry and wet seasons, using seasonal mean Na\(^+\) and Cl\(^-\) concentrations of the precipitation or Indian River Lagoon Water. The evapoconcentration model was:

\[
C_{RES} = \frac{C_{INI}}{f_{RES}}
\]

where \(C\) is the modeled Na\(^+\) or Cl\(^-\) concentration in mg/L, \(f\) is the fraction of water remaining as it evaporates, and the subscripts ‘RES’ and ‘INI’ refer to residual water (e.g., evaporated ground-water in the mangrove) and initial water (e.g., precipitation or Indian River Lagoon Water), respectively.

Isotope enrichment due to evaporation is the process by which \(\delta D\) and \(\delta^{18}O\) increase as water evaporates and heavier isotopes are retained in the remaining solution. Isotope enrichment was modeled through the application of a Rayleigh distillation model (Clark and Fritz 1997). The isotope enrichment model was:

\[
R = R_0 f^{(\alpha-1)}
\]

where \(R\) is the modeled isotope composition, \(R_0\) is the initial isotope composition, \(f\) is the fraction of water remaining as it evaporates, and \(\alpha\) is the equilibrium fractionation factor for evaporation (Majoube 1971). Kinetic effects due to high humidity were taken into account by applying a correction factor to the fractionation factor (Gonfiantini 1986). Two separate models were completed using theoretical precipitation and sea water as source waters. The precipitation source term was taken from the mean isotopic composition of precipitation at Kennedy Space Center, located ~113 km north and in the same physiographic
region as the study site (http://www.uaa.alaska.edu/enri/usnip/isotope_1989-2001.cfm), while the sea water source term was taken from the isotopic composition for theoretical sea water (Gonfiantini 1978).

**Resistivity Survey**

A 400-m resistivity survey was conducted on an east-west transect from the Indian River Lagoon to the dune crest of North Hutchinson Island (Figure 2-1). The shoreward half of the survey was completed on September 28, 2006 and the upland half on March 3, 2007. Data were collected using an Advanced Geosciences Incorporated Marine SuperSting R8 multichannel system connected to an external switching box that controlled the flow of current along the 56-electrode cable (AGI, Austin, Texas, USA). Electrodes (metal spikes) were spaced at 1.5 m intervals. Contact resistance was minimized as far as possible by driving the electrodes as deep as practical (typically ~ 20-30 cm deep) and watering the electrode-ground contact with sea water. Three individual surveys were completed, with a 20% overlap between surveys. Resistivity surveys measure terrain resistivity, which is a weighted average of the deposit and fluid resistivities. Therefore, resistivity is not solely a function of salinity. However, the extreme contrast in resistivity between fresh water (high resistivity/low conductivity) and sea water (low resistivity/high conductivity) makes the freshwater/sea water interface a strong component of measured terrain resistivities in any setting. Following acquisition, the data were inverted using the EarthImager 2D inversion software from Advanced Geosciences Incorporated.
Results

Physical Hydrology

Raw water-level data were converted to equivalent fresh-water head data to allow direct comparison of heads in fluids with varying densities. With the exception of a few excursions, heads in the mangrove habitats were lower than stages in the Indian River Lagoon (Figure 2-2). Mean annual heads for the two continuously-monitored piezometers in the impoundment were -0.09 m and 0.02 m amsl, while mean annual stage of the Indian River Lagoon was 0.21 m amsl.

Figure 2-2. Equivalent fresh-water head (m) of tide in the Indian River Lagoon and groundwater level in each of the piezometers. Measurements were recorded on 15-minute intervals with a pressure transducer and data logger.
Spectral analysis of the piezometer and tidal data shows that the time frequencies of patterns in heads in the mangrove and Indian River Lagoon differ (Figure 2-3). Heads in the mangrove have strong spectral peaks at 24.0 hours and weak spectral peaks at 12.0 hours, while stage in the Indian River Lagoon has a strong spectral peak at 12.4 hours and weak spectral peaks at 24.0 and 25.8 hours.

The direct effect of evapotranspiration on water levels can be seen in the examination of a segment of the long-term water-level record (Figure 2-4A). The water-level data between November 25, 2007 and November 26, 2007 show that heads decline from the early morning to a minimum at midday, and then recover through the afternoon and night to a maximum in the early morning. These water-level fluctuations correspond with daily evapotranspiration rates of 2 mm and 9 mm, which are within the range of evapotranspiration values (0–10 mm) reported for the Indian River Lagoon in the month of November from a previous study (Sumner and Belaineh 2005). The same trend in water levels is found throughout the long-term record and can be seen in the mean daily head cycle (Figure 2-4B).

**Chemical Hydrology**

Ground-water salinities varied by community type (Figure 2-5). Mean ± SD ground-water salinity was 78 ± 9 psu in the salt pan (n = 48), 55 ± 2 psu in the sparse black mangrove (n = 59), 44 ± 5 psu in the dense black mangrove (n = 56), 32 ± 4 psu in the red mangrove (n = 52), and 10 ± 2 psu in the upland forest (n = 19). Ground-water salinities varied significantly between community types.
(ANOVA, \( p < 0.01 \); all Tukey post-hoc comparisons \( p < 0.01 \)) but showed no apparent seasonal or annual variations or trends throughout the course of the 2.5-year study.

Figure 2-3. Fourier transform results showing frequency spectra of equivalent fresh-water head data for the tide and piezometer ground-water levels. The y-axis shows the relative amplitude of the peaks, with the most intense peak having a value of 1. The x-axis is the frequency of the water-level pattern (1/hours).
Figure 2-4. (A) Representative long-term water level data from November 25–26, 2007 showing daily changes due to evapotranspiration. (B) Average daily cycle of entire equivalent fresh-water head (m) data set for the tide and piezometers showing lack of tidal influence on piezometer water levels. Y-axis is in the number of hours after midnight.
Figure 2-5. Average ground-water salinity values for each wet and dry sampling event in each of the five habitat classifications. Bars are the average ground-water salinity for all depths at all sites sampled in that habitat and are arranged in ascending date from left to right. Error bars are the standard deviations of those averages (± 1σ).

There were substantial differences in the absolute concentrations of the cations and anions in precipitation, surface water in the primary perimeter ditch inside the impoundment and the Indian River Lagoon, and ground water in each of the five community types (Table 2-1; Figure 2-6). In all but one case (Ca^{2+}), cation and anion concentrations were greatest in salt pan ground water, followed by the sparse black mangrove ground water, dense black mangrove ground water, red mangrove ground water, surface water in the primary perimeter ditch inside the impoundment and Indian River Lagoon, woody upland ground water, and precipitation. Although the absolute concentrations varied, the relative
Table 2-1. Average cation and anion concentration (ppm) and isotopic composition (‰) of each water type sampled during the investigation. Standard deviations shown are ±1σ. Isotopic averages for precipitation are long-term records at Kennedy Space Center, located ~113 km north and in the same physiographic region as the study site.

<table>
<thead>
<tr>
<th>Water Type</th>
<th>Precipitation (n=5)</th>
<th>Surface Water (n=6)</th>
<th>Salt Pan (n=18)</th>
<th>Sparse Black (n=21)</th>
<th>Dense Black (n=19)</th>
<th>Red (n=12)</th>
<th>Woody Upland (n=9)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Na⁺ (ppm)</td>
<td>5 ± 4</td>
<td>8600 ± 900</td>
<td>33480 ± 60</td>
<td>23000 ± 2000</td>
<td>16600 ± 500</td>
<td>11200 ± 500</td>
<td>2500 ± 700</td>
</tr>
<tr>
<td>K⁺ (ppm)</td>
<td>0.5 ± 0.2</td>
<td>320 ± 30</td>
<td>1050 ± 30</td>
<td>750 ± 90</td>
<td>558 ± 7</td>
<td>410 ± 10</td>
<td>112 ± 2</td>
</tr>
<tr>
<td>Ca²⁺ (ppm)</td>
<td>0.5 ± 0.3</td>
<td>380 ± 30</td>
<td>1020 ± 10</td>
<td>1000 ± 100</td>
<td>920 ± 60</td>
<td>600 ± 100</td>
<td>210 ± 60</td>
</tr>
<tr>
<td>Mg²⁺ (ppm)</td>
<td>1 ± 1</td>
<td>1100 ± 100</td>
<td>4220 ± 70</td>
<td>3000 ± 200</td>
<td>2260 ± 60</td>
<td>1450 ± 10</td>
<td>350 ± 30</td>
</tr>
<tr>
<td>Cl⁻ (ppm)</td>
<td>10 ± 10</td>
<td>15000 ± 2000</td>
<td>64000 ± 1000</td>
<td>42000 ± 2000</td>
<td>31000 ± 1000</td>
<td>20500 ± 500</td>
<td>4900 ± 600</td>
</tr>
<tr>
<td>SO₄²⁻ (ppm)</td>
<td>2 ± 1</td>
<td>2300 ± 500</td>
<td>7800 ± 400</td>
<td>5600 ± 400</td>
<td>4400 ± 700</td>
<td>2700 ± 300</td>
<td>700 ± 20</td>
</tr>
<tr>
<td>dD (‰)</td>
<td>-20.8</td>
<td>9 ± 1</td>
<td>8 ± 2</td>
<td>7 ± 3</td>
<td>5 ± 4</td>
<td>6 ± 4</td>
<td>-15 ± 5</td>
</tr>
<tr>
<td>d¹⁸O (‰)</td>
<td>-3.85</td>
<td>0.9 ± 0.2</td>
<td>1.4 ± 0.5</td>
<td>0.8 ± 0.6</td>
<td>0.3 ± 0.8</td>
<td>0.3 ± 0.7</td>
<td>-3.3 ± 0.6</td>
</tr>
</tbody>
</table>

Figure 2-6. Schoeller diagram showing the average cation and anion concentrations (meq/L) of each water type collected in both the wet and dry seasons.
proportions of the cations and anions were relatively constant across precipitation, surface water in the primary perimeter ditch inside the impoundment and Indian River Lagoon, and ground water in each of the five community types (Figure 2-6).

Precipitation anion and cation concentration averages were comparable to those reported by the National Atmospheric Deposition Program monitoring location at Kennedy Space Center (http://nadp.sws.uiuc.edu/sites/siteinfo.asp?id=FL99&net=NTN). The cation and anion averages reported by the monitoring program all fall within the standard deviation of the averages of samples collected during this study.

Isotope data indicate that most of the surface water and ground water was evaporated (Figure 2-7). Ground water in the woody upland had a mean isotopic composition of -15.49‰ and -3.32‰ for δD and δ18O, respectively, which plotted on the Global Meteoric Water Line (Craig 1961) near the mean isotopic composition for regional precipitation (http://www.uaa.alaska.edu/enri/usnip/isotope_1989-2001.cfm). Surface water in the primary perimeter ditch inside the impoundment/Indian River Lagoon and ground water in the four remaining community types plotted along a local evaporative trend line that intersected the Global Meteoric Water Line near the mean isotopic composition for regional precipitation and the mean isotopic composition of sea water. The slope of this local evaporative trend line was 5.0, which is indicative of evaporation in an environment with a relative humidity of ~70%, which is the approximate mean annual humidity of this region (Southeast
Figure 2-7. $\delta^D$ and $\delta^{18}O$ (‰) of precipitation from long-term records at Kennedy Space Center, located ~113 km north and in the same physiographic region as the study site, theoretical sea water, and surface water and ground water collected during this study. The symbol shows the mean composition and the error bars represent the standard deviations (± 1σ) of those mean values. The solid black line is the global meteoric water line as defined by Craig (1961). The dashed line is the local evaporative trendline ($y = 5.6x + 2.7$; $R^2 = 0.98$), determined via linear regression of the sample isotopic compositions.

Regional Climate Center data for Vero Beach, Florida). Salt-pan ground water was the most evaporatively enriched, with mean isotopic composition of 7.95‰ and 1.41‰ for $\delta^D$ and $\delta^{18}O$, respectively.

**Evapoconcentration and Isotope Enrichment Modeling**

The evapoconcentration model results show that Na$^+$ and Cl$^-$ concentrations in mangrove ground water can be produced through the evaporation of precipitation and/or Indian River Lagoon water (Figure 2-8A). Mean Na$^+$ and Cl$^-$ concentrations in the red mangrove, dense black mangrove, sparse black mangrove, and salt pan community types can be produced by
evaporating >99% of the precipitation or 21%, 46%, 61%, and 74% of the Indian River Lagoon water, respectively. Similarly, the isotope enrichment model results also show that mangrove ground water can be produced through the evaporation of precipitation and/or Indian River Lagoon water (Figure 2-8B). Mean isotopic compositions in the red mangrove, dense black mangrove, sparse black mangrove, and salt pan community types can be produced by evaporating 23%, 23%, 25%, and 28% of the precipitation, respectively, or 3%, 3%, 6% and 7% of the Indian River Lagoon water, respectively.

![Figure 2-8](image)

**Figure 2-8.** (A) Evapoconcentration modeling results showing that sodium and chloride concentrations (ppm) in each of the community types can be achieved by evaporating precipitation and/or Indian River Lagoon water. (B) Isotope enrichment results showing that $\delta^D$ and $\delta^{18}O$ (‰) in each of the community types can be achieved through the Rayleigh distillation of precipitation and/or sea water. In panel A, the trendlines are the calculated increases in those concentrations, assuming that sodium:chloride ratio remains constant, as water is evaporated. In both plots, the points represent the mean composition (sodium and chloride in Figure 8A and $\delta^D$ and $\delta^{18}O$ in Figure 8B) and the error bars represent the standard deviations (± 1σ) of those mean values.
**Resistivity**

The resistivity results were somewhat noisy, most likely due to poor contact between electrodes and hard or resistive soils, but larger-scale spatial patterns were clear (Figure 2-9). There was a discernable transition between the mangroves (i.e., the salt pan, sparse black mangrove, dense black mangrove, and red mangrove) and the woody upland at a distance of approximately 220 m along the profile (Figure 2-9). Below the mangrove, terrain resistivities were primarily ≤1 ohm-m to depths of at least 17 m, the maximum depth of the survey. Below the mangrove a lower-resistivity layer persists in the depth range of 1–4 m, thinning somewhat towards the woody uplands. Below the low resistivity layer lies a zone with higher resistivities between ~5–10 m depth. Any vertical

![Figure 2-9](image-url)

**Figure 2-9.** (A) Profile showing resistivity inversion results. Horizontal axis is distance along the transect (m) from the coastline. (B) General interpretation of resistivity results showing corresponding large-scale spatial patterns in conductivity. Resistivities are shown in ohm-m. Black line and arrow indicate the transition from mangrove to upland vegetative communities. Dashed line represents a confining clay layer located under the mangrove that tapers out at the vegetative upland transition.
stratification is less clear below the woody upland, where terrain resistivities are overall higher, with values generally ≥1 ohm-m.

Discussion

Annual water levels in the hydrologically altered impoundment appear to be largely controlled by evapotranspiration. Long-term water-level records show that water levels in the mangrove are generally lower than those in the Indian River Lagoon (Figure 2-2). Evaporation is likely higher in the Indian River Lagoon (~1500 mm year\(^{-1}\)) than evapotranspiration in the mangrove (~950 mm year\(^{-1}\)) (Twilley and Chen 1998; Sumner and Belaineh 2005). Nevertheless, drawdown is greater in the mangrove than in the Indian River Lagoon because (a) the Indian River Lagoon receives fresh-water inflows from the mainland and sea-water inflows from the Atlantic Ocean and (b) the Indian River Lagoon has an effective specific yield of 10\(^0\) while the mangrove sediments have a specific yield of 10\(^{-1}\) (Johnson 1967), which means that equal evapotranspiration will lead to a 10-fold greater drawdown in the mangrove. This greater drawdown in the mangrove maintains the hydraulic gradient from the Indian River Lagoon to the mangrove.

Daily water-level variations in the impoundment also appear to be largely controlled by evapotranspiration. The Indian River Lagoon Fourier spectrum shows a strong peak at 12.4 hours and weak peaks at 24.0 and 25.8 hours (Figure 2-3). The 12.4-hour peak corresponds to the tide’s diurnal cycle, while the 24.0 and 25.8 hour peaks correspond to the solar and lunar phases,
respectively, of the tide’s harmonic signal. The mangrove Fourier spectrum shows a strong peak at 24.0 hours and a weak peak at 12.0 hours (Figure 2-3). The 12.4-hour peak corresponding to the tide’s diurnal cycle and the 25.8 hour peak corresponding to the lunar phase of the tide’s harmonic signal are not evident. The 24.0 hour peak corresponds to the daily solar cycle, which controls both the solar phase of the tide’s harmonic signal and the mean daily peak evapotranspiration.

These peaks can be seen in the mean daily head and stage cycles (Figure 2-4B). Mean stages in the Indian River Lagoon follow a sinusoidal curve, as might be expected for stages in a lagoon so close to a large inlet. Conversely, mean heads in the mangrove follow a pattern indicative of evapotranpiration control. Heads rise most steeply between 24:00 and 06:00 when plants are not transpiring, and decline most steeply between 06:00 and 12:00 when plants are transpiring. Interestingly, heads begin to rise after 12:00, although the rate at which they rise is slower than later in the evening. The transition between the heads declining and rising is abrupt. This suggests that evapotranspiration slows considerably and perhaps even ceases in some species in the afternoon, which results in the rate of inflow exceeding the rate of outflow (i.e., the rate of evapotranspiration). It has been shown that mangrove evapotranspiration rates decrease when salinities are increased (Ball and Farquhar 1984; Biber 2006). However, it is unclear if high salinity results in evapotranspiration rates that are generally lower throughout the day or in xylem dysfunction and evapotranspiration rates that are lower in the afternoon when atmospheric
demand is highest (Tyree 1997). These results suggest the latter may be occurring, although this is merely speculative and deserving of independent study.

The observed Na$^+$ and Cl$^-$ concentrations and $\delta$D and $\delta^{18}$O compositions can both be modeled by theoretically evaporating precipitation and/or Indian River Lagoon water (Figure 2-8). However, the amount of evaporation necessary to model the observed Na$^+$ and Cl$^-$ concentrations far exceeds the amount of evaporation necessary to model the observed $\delta$D and $\delta^{18}$O compositions. This implies that transpiration plays an important role in controlling salinities because transpiration concentrates solutes by exclusion of solutes at the roots, extrusion of solutes from glands, and/or storage of solutes in leaves that ultimately drop and decompose (Dawes 1998), but does not result in isotope fractionation between the uptaken water and the remaining ground water (Gat 1996).

Collectively, these results indicate that evapotranspiration plays an important role in controlling water levels and salinities in the barrier island mangrove on the east-central coast of Florida (Figure 2-10). Evapotranspiration lowers water levels in the mangrove, which creates a hydraulic gradient that drives water from the Indian River Lagoon to the mangrove. While infrequent tidal inundation of the mangrove does occur, the regular, daily pathway for this water movement is subsurface. Water evaporates and is transpired but solutes remain in solution, so mangrove water evapoconcentrates, becomes more dense, sinks, and creates the layering of high-salinity waters in the subsurface 1–4 m below
Figure 2-10. Conceptual model showing importance of evapotranspiration in controlling water levels and salinities in the barrier island mangrove. As water evaporates a head gradient is created causing Indian River Lagoon water to flow into the mangrove. The high-density, hypersaline waters that result from the evapoconcentration sink and cause layering in the subsurface below the mangrove. The solid line represents the sub-marine and terrestrial land surface and the dashed line represents the clay layer that acts as a confining unit for the hypersaline waters.

the mangrove. However, the ultimate fate of the sinking, high-salinity water remains unclear.

Acknowledgments

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CHAPTER 3:
CHANGES IN THE SPATIAL VARIABILITY OF TERRAIN CONDUCTIVITY IN A MANGROVE, INDIAN RIVER LAGOO N, FLORIDA

Abstract

We examined the differences in spatial variability of terrain conductivity in a mangrove on a carbonate barrier island along the Indian River Lagoon under two hydrologic management regimes, breached rotational impoundment management and rotational impoundment management. Continuous water-level measurements were recorded in a piezometer for two years. EM31 and EM34 measurements were taken throughout the mangrove during both regimes and then inverted to obtain terrain conductivity for the shallow and deep layers, respectively. The difference in coefficient of variation (ΔCV) between the breached RIM and RIM data was calculated to examine spatial variability in both the shallow and deep layers. A null-hypothesis model was employed to examine the statistical significance of the ΔCV results, using the 90% confidence-interval as a benchmark. The average water levels were -0.06 m amsl and 0.49 m amsl during the breached-RIM and RIM regimes, respectively. The average shallow (EM31) layer terrain conductivity shifted slightly from 1868 mS m\(^{-1}\) to 1825 mS m\(^{-1}\) after the alteration in management regime, yet the standard deviation of these averages decrease from 656 mS m\(^{-1}\) to 216 mS m\(^{-1}\). The average deep
(EM34) layer terrain conductivities were 328 mS m\(^{-1}\) and 255 mS m\(^{-1}\) during the breached-RIM and RIM regimes, respectively. The temporal ΔCVs were 0.23 and -0.04 for the shallow and deep layers, respectively. The null-hypothesis model for the shallow layer illustrates that the ΔCV of 0.23 is outside the 90% confidence interval and determines that the difference in spatial structure is statistically significant. The deep layer ΔCV was not statistically significant. These results indicate that the transition from breached RIM to RIM resulted in changes to both the physical and chemical hydrologic character of the impoundment, especially in the shallow layer.

**Introduction**

Mangroves, which cover ~240,000 km\(^2\) of sheltered subtropical and tropical coasts between latitudes 24° north and south where mean annual temperatures are >20°C (Lugo 1990; Dawes 1998), provide numerous ecological functions as well as goods and services. Mangroves support estuarine and near-shore marine productivity, in part by providing critical habitat for juvenile fish and through the export of nutrient-rich water (McKee 1995; Rivera-Monroy et al. 1998) or plant, algal, or animal biomass (Zetina-Rejón et al. 2006). Mangroves also protect coastal habitats against the destructive forces of hurricanes, typhoons, and tsunamis (e.g., Kandasamy and Narayanasamy 2005; Granek and Rutenberg 2007; Alongi 2008).

Mangroves also have historically provided breeding grounds for numerous species of mosquito (Ritchie and Addison 1992). In Florida, early accounts tell of
mosquitoes so numerous that early European explorers were forced to sleep on the beach covered with sand (Connelly and Carlson 2009) and mosquito-borne disease outbreaks were common, with the 1888 outbreak of yellow fever in Jacksonville killing 400, sickening 5,000, and causing 10,000 to flee the city, resulting in the temporary loss of ~40% of the total population (Mulrennan 1986). Consequently, insect-control efforts have long been undertaken, including ditching and draining, the widespread application of pesticides, and the construction of impoundments (Connelly and Carlson 2009). Impoundments were shown to be effective in the 1930s (Hull and Dove 1939), but were not widely constructed from the 1950s to the 1970s (Rey and Kain 1991; Connelly and Carlson 2009). By the mid-1970s, more than 16,400 ha of mangroves had been impounded on the Indian River Lagoon (Rey and Kain 1991).

Most impoundments consist of perimeter levees constructed between mean lower low and mean higher high tide. Culverts are typically placed beneath the levees to facilitate control over inflow and outflow. In some cases, gates are installed to either maintain continuous or seasonal flooding; in other cases, gates are either not installed or are installed but not closed to allow free inflow and outflow. More recently, pump systems have been increasingly installed and operated to allow better control over seasonal inflow and outflow and associated flood durations and depths (Connelly and Carlson 2009).

Impoundment can result in numerous environmental impacts, including changes in the occurrences of certain species of fish (Harrington and Harrington 1982), birds (Provost 1959), and vegetation (Rey et al. 1990). In some cases,
monotypic stands of red mangrove (*Rhizophora mangle*) replace spatially complex stands of red mangrove, black mangrove (*Avicennia germinans*), herbaceous halophytes (e.g., *Batis maritima*), and unvegetated salt pans (Rey et al. 1990). Porewater salinities can differ between the different habitat types in these spatially complex stands (Odum and McIvor 1990; Stringer et al. 2010), suggesting that impoundment both reduces overall salinities—thereby favoring less salt-tolerant red mangrove—and reduces the spatial complexity of salinity distributions—thereby favoring monotypic stands of red mangrove.

This study is part of a broader series of investigations of the controls on species composition, primary productivity, and nutrient cycling in an impoundment on a carbonate barrier island on the east-central coast of Florida. A previous investigation showed that evapotranspiration plays a major role in controlling salinities by lowering the water table in the mangrove and creating a hydraulic gradient that draws surface water and ground water from the lagoon into the mangrove (Stringer et al. 2010). Since the conclusion of that study, management has changed, with culverts without gates allowing free exchange between the impoundment and the adjacent lagoon being replaced by culverts with gates and pumping of water into the impoundment from the adjacent lagoon. Therefore, this subsequent investigation tests the hypothesis that flooding the impoundment both reduces overall salinities and reduces the spatial complexity of salinity distributions.
Study Area

The study was conducted in SLC-24, a mosquito-control impoundment located at N27°33’, W80°33’ on the west shore of North Hutchinson Island. This barrier island is ~35-km in length and ~0.2–2 km in width, and is part of the system of barrier islands that bound the Indian River Lagoon, a 250-km estuarine system located on the east-central coast of Florida (Figure 3-1).

The impoundment was created in 1970 when a dike was constructed around an existing mangrove-dominated wetland (Rey et al. 1990; Rey and Kain

Figure 3-1. Local setting showing EM31 and EM34 sampling points, piezometers and instrumentation, as well as cross-section location.
Culverts were not installed so there was no surface-water exchange between the impoundment and the lagoon between 1970 and 1985. Surface-water exchange was minimally restored in 1985 when one culvert was installed to facilitate the drainage of water that had accumulated in the impoundment during two tropical storms. The culvert was later closed so there was no surface-water exchange between the impoundment and the lagoon again until 1987 when the existing culvert was reopened and four additional culverts were installed. From 1987 to 2009, these culverts remained open, allowing focused but otherwise free exchange of surface water between the impoundment and the lagoon, a management strategy called breached rotational impoundment management (breached RIM) (Middleton et al. 2008). In 2009, pumps and tide gates were installed and have since been operated to maintain surface-water inundation during the summer months, a management strategy called rotational impoundment management (RIM) (Middleton et al. 2008).

Continuous core collected to depths of approximately 10 m in four locations across the barrier island indicate that there is ~5 m of fine sand, which is underlain by ~1.5 m of clayey sand, which is underlain shelly fine sand (Figure 3-2). The clay-rich layer apparently acts as an aquitard, with surface water and ground water above the aquitard evaporating and concentrating to develop hypersaline environments, particularly in areas that are infrequently flooded by high tides (Stringer et al. 2010).

Vegetation cover in the impoundment decreased from 75% to near 30% between 1970 and 1985 but began to recover following the installation of the
Figure 3-2. Geologic cross section characterizing the site subsurface on a west-east transect. The matrix is dominated by sand and sand with clay closest to the ground surface, with a confining layer of clayey sand and sandy clay occurring at 3 m below mean sea level.

culverts in 1987 (Rey et al. 1990). Black mangrove (Avicennia germinans (L.) L.) is the dominant mangrove, but red (Rhizophora mangle L.) and white (Laguncularia racemosa (L.) C.F. Gaertn.) mangrove also are common. Prior to the change to RIM, the abundance, density, and sizes of mangroves varied across the impoundment from highly saline salt pans that had no mangroves or were fringed by dwarf black mangroves, areas that had dense and relatively short black mangroves, areas that had larger and more widely spaced black mangroves, to areas associated with open water where red mangroves were most abundant.
Methods

**Water Levels and Ground-Water Conductivities**

Piezometers were installed at 18 locations. At 17 locations, two piezometers were installed, with piezometers being constructed of stainless steel, having 2.5 cm inside-diameters, and being screened over 15 cm beginning at 1.2 m and 1.8 m below the ground surface. Ground-water conductivity was measured at these piezometers using a YSI 556 MPS (YSI Inc., Yellow Springs, Ohio, USA). At one location, one piezometer was installed, with the piezometer being constructed of PVC, having a 5.0 cm inside-diameter, and being screened over 30 cm beginning at 2.0 m below the ground surface. Ground-water head was measured at this piezometer on 15-minute intervals with Model 3001 Levelogger (Solinst, Georgetown, Ontario, Canada). The pressure transducers had a 1.5 m range and a ±1 mm accuracy. Head data were compensated for changes in barometric pressure using barometric pressure data also measured on 15-minute intervals with a Barologger (Solinst, Georgetown, Ontario, Canada).

**Electromagnetic Survey Measurements**

Electromagnetic surveys were conducted while the impoundment was under breached RIM in April 2007 and under RIM in September 2009 (Figure 3-1). Electromagnetic measurements were made throughout the site using EM-31 and EM-34 instruments (Geonics LTD., Mississauga, Ontario, Canada). For each instrument, a transmitter coil was placed at the ground surface to generate alternating currents that induced eddy currents in the subsurface. Those eddy
currents generated secondary magnetic fields that were measured, along with the primary magnetic fields, by receiver coils placed specific distances away from the transmitters. The ratios of the secondary to primary magnetic fields are proportional to the apparent conductivities (McNeill 1980). Inversion modeling software, EMIX 34 (Interpex LTD, Golden, Colorado, USA), was then used to invert the apparent conductivities to provide terrain conductivities, which are the combined conductivities of the sediments and the porewater fluids. The EMIX 34 inversion modeling software uses given values for instrument height, dipole mode, frequency, and coil spacing to invert the set of instrument responses to find the best-fit values of one or more layer conductivities using a ridge-regression estimation algorithm (Inman 1975).

The EM-31 was operated with a fixed coil spacing of 3.7 m at a frequency of 9.8 kHz. At each location, measurements were taken in the horizontal dipole and vertical dipole modes, providing an effective penetration of 3 m and 6 m, respectively. Two measurements were taken in each mode; one on the ground surface and one at hip height, ~0.75 m above ground surface. Data were interpreted using the EMIX 34 inversion modeling software with a one-layer model.

The EM-34 was operated at 10 m and 20 m spacings at frequencies of 6.4 kHz and 1.6 kHz, respectively. Measurements were only taken in the horizontal dipole mode, resulting in an effective penetration of 7.5 m and 15 m for the 10 m and 20 m spacing, respectively. No measurements were taken in the vertical dipole mode because the terrain was too conductive for the range of the
instrument in this configuration. Data were interpreted using the EMIX 34 inversion modeling software with a one-layer model.

**Ground-Water vs. Terrain Conductivity**

Ground-water conductivities were measured at the same time that EM-31 measurements were taken in both vertical and horizontal dipole modes adjacent to the piezometers. The measured ground-water and modeled terrain conductivities were then compared using least-squares regression in Excel (Microsoft Corporation, Redmond, Washington, USA) to ascertain the validity of using terrain conductivity as a proxy for ground-water conductivity.

**Null-Hypothesis Modeling**

The terrain conductivity data were used to test the hypothesis that there was greater spatial variation in terrain conductivities while the impoundment was under breached RIM in April 2007 than while the impoundment was under RIM in September 2009. First, the impoundment was stratified into shallow and deep layers, with all EM-31 data assigned to the shallow layer and all EM-34 data assigned to the deep layer. The coefficient of variation (CV), the ratio of the standard deviation to the mean, was calculated for each layer in each year. For each layer, the CV from September 2009 was subtracted from the CV from April 2007, which yielded the difference in the coefficient of determination ($\Delta$CV). Subsequently, a null model was used to randomly reassign terrain conductivities within each layer, with measured values in any location in any year randomly reassigned to any location in any year (Lewis et al. 2006; Lewis et al. 2007). This reassignment was completed $5 \times 10^5$ times, with a new $\Delta$CV calculated for each
of the iterations. The observed $\Delta$CVs were then compared to the frequency
distribution of the new $\Delta$CVs generated by the null model, and the observed
$\Delta$CVs were determined to be significantly different than the new $\Delta$CVs generated
by the null model if they fell outside of the 90% confidence interval.

**Results**

**Physical Hydrology**

Continuous water-level measurements from water years 2007 (i.e.,
October 1, 2006 to September 30, 2007) and 2009 (i.e., October 1, 2008 to
September 30, 2009) show the change in water levels due to the change in
impoundment management approach. In 2007, mean ± standard deviation head

![Figure 3-3. Water level in the mangrove (A) under breached RIM regime in water year 2007, and (B) under RIM regime in water year 2009. The management regime change occurred in March 2009. The arrow in panel A indicates the first sampling period in April 2007. The arrow in panel B indicates the sampling period in September 2009. All measurements were recorded on 15-minute intervals with a pressure transducer and data logger.](image)
was -0.09 ± 0.10 m amsl from October to March and -0.08 ± 0.11 m amsl from April through September (Figure 3-3A). In 2009, mean ± standard deviation head was -0.06 ± 0.19 from October to March and 0.49 ± 0.14 m amsl from April through September (Figure 3-3B).

**Relationship between Ground-Water and Terrain Conductivities**

Ground-water conductivities ranged from 1,700 to 15,000 mS m⁻¹, while the modeled terrain conductivities ranged from 330 to 3,800 mS m⁻¹ (Figure 3-4). Least-squares regressions of ground-water conductivity (dependent variable) against terrain conductivity (independent variable) were performed for both the 1.2-m and 1.8-m piezometers data sets.

![Figure 3-4. Correlation between ground-water conductivity and modeled terrain conductivity. The black diamonds and open squares represent ground-water conductivities taken at piezometers with a depth of 1.2 m and 1.8 m, respectively. The solid black and grey lines represent the least-squares regression for the 1.2-m and 1.8-m conductivities, respectively. The dashed black and grey lines represent the standard deviation (SD) lines for the 1.2-m and 1.8-m conductivities, respectively.](image-url)
The standard deviation lines (SD lines) were calculated and plotted to illustrate the correlation that would occur with a perfect fit ($R = 1$). With actual values of $R = 0.79$ ($p = 0.02$) and $R = 0.87$ ($p < 0.01$) for the 1.2-m and 1.8-m piezometers, respectively, the slopes of the fitted regression lines are reduced accordingly (by 21% and 13% respectively). The $R^2$ values mean that the fitted lines capture 62% and 76%, respectively, of the variance of the ground-water conductivity, assuming a linear relationship between ground-water conductivity and terrain conductivity, and no error in the measurements of terrain conductivity.

Inspection of the residuals from the regression lines gives no indication that the assumption of a linear relationship for these data is inappropriate. The most likely sources of the unexplained variability in ground-water conductivity, therefore, is interpreted to be measurement error and differences in the volumes sampled during the measurement of the two conductivity variables. With regards to the latter source, ground-water conductivities were measured on fluids drawn from a small volume immediately surrounding the screened intervals 1.2 m and 1.8 m below the ground surface, while terrain conductivities were integrated over a depth of ~6 m and sample volumes of ~490 m$^3$. There could be heterogeneity in the ground-water conductivities within that larger volume, resulting in different values.

**Terrain Conductivity**

Modeled terrain conductivities for the 2007 EM-31 survey had a mean ± standard deviation of 1,868 ± 656 mS m$^{-1}$ and ranged from 116 to 2,917 mS m$^{-1}$ (n=38). Analysis of the spatial arrangement of these values shows an axis of
anisotropy at 143.6° (Figure 3-5A). The highest conductivities occurred along this NW to SE axis, while the lowest conductivities occurred in the northeastern corner of the study area. The terrain conductivity values of the 2009 EM-31 survey had a mean ± standard deviation of 1,825 ± 216 mS m\(^{-1}\) and ranged from 1,489 to 2,430 mS m\(^{-1}\) (n=43). Forty-two percent of the values were in the range of 1,800 to 1,900 mS m\(^{-1}\), resulting in almost no spatial variability or structure in the data (Figure 3-5B).

The terrain conductivities for the 2007 EM-34 survey had a mean ± standard deviation of 328 ± 40 mS m\(^{-1}\) and ranged from 246 to 404 mS m\(^{-1}\) (n=18). Analysis of the spatial arrangement of these values reveals an axis of anisotropy at 152.2° (Figure 3-6A). The lowest conductivities occurred along this NW to SE axis, with values increasing gradually toward the edges of the study area. The terrain conductivities for the 2009 EM-34 survey had a mean ± standard deviation of 255 ± 42 mS m\(^{-1}\) and ranged from 201 to 378 mS m\(^{-1}\) (n=37). Analysis of the spatial arrangement of these values shows an axis of anisotropy at 7.7° (Figure 3-6B). The lowest values occurred along this W to E axis, with values increasing toward the N end of the study area.

**Temporal Variability in Terrain Conductivity**

The \(\Delta CV\) values for the random reassignment of EM31 data from 2007 and 2009 ranged from -0.21 to 0.23, in a normal distribution (Kolmogorov-Smirnov non-normality test, \(p = 0.19\)) (Figure 3-7A). The \(\Delta CV\) of the actual data, 0.23, had a z-score of 3.35. Because the z-score exceeds the 1.96 threshold (i.e., the 95% confidence interval), we reject the null hypothesis and conclude
that the $\Delta CV$ was not produced by a random rearrangement of the 2007 and 2009 data. There was a statistically significant change in the spatial structure of near-surface terrain conductivity values from the 2007 to 2009 sampling periods.

Figure 3-5. Contoured surfaces of shallow (EM31) terrain conductivities under (A) breached RIM in 2007 and (B) RIM in 2009. All units are in $\text{mS m}^{-1}$. The average conductivities were $1,868 \pm 656 \text{ mS m}^{-1} (n=38)$ in 2007 and $1,825 \pm 216 \text{ mS m}^{-1} (n=43)$ in 2009.
Figure 3-6. Contoured surfaces of deep (EM34) terrain conductivities under (A) breached RIM in 2007 and (B) RIM in 2009. All units are in mS m\(^{-1}\). The average conductivities were 328 ± 40 mS m\(^{-1}\) (n=18) in 2007 and 255 ± 42 mS m\(^{-1}\) (n=37) in 2009.
Figure 3-7. Histograms showing null-hypothesis modeling results for (A) shallow (EM31) $\Delta CV$ randomized values and (B) deep (EM34) $\Delta CV$ randomized values. Both plots illustrate the distribution of 500,000 randomizations. The arrows show the position of the observed $\Delta CV$ values in relation to the randomized values. The observed $\Delta CV$ for the EM31 measurements had a $z$-score of 3.35 indicating there was a statistically significant change in spatial structure in terrain-conductivity values from 2007 to 2009.

The $\Delta CV$ values for the random reassignment of EM34 data from 2007 and 2009 ranged from -0.12 to 0.11, in a normal distribution (Kolmogorov-Smirnov non-normality test $p = 0.38$) (Figure 3-7B). The $\Delta CV$ of the actual data, -0.04, had a $z$-score of -1.56. Because the $z$-score falls outside the ±1.96 threshold (i.e., the 95% confidence interval), we do not reject our null hypothesis. There was not a statistically significant change in spatial structure of deep terrain conductivity values from the 2007 and 2009 sampling periods.

Discussion

The transition from breached RIM to RIM resulted in changes to both the physical and chemical hydrologic character of the impoundment. Overall, there
were clear changes in both water levels and the variability and spatial structure of porewater salinities.

Under breached RIM, water levels generally ranged between -0.5 m to 0.1 m relative to mean sea level throughout the year (Figure 3-3A). Under RIM, water levels generally ranged between -0.5 m to 0.1 m relative to mean sea level when the impoundment was not flooded (i.e., October 2008 to February 2009), but greatly increased and generally ranged between 0.1 m to 0.6 m relative to mean sea level when the impoundment was flooded (i.e., March 2009 to the end of the study in September 2009) (Figure 3-3B). Mean ± standard deviation elevations at the 18 piezometers were -0.02 ± .22 m relative to mean sea level, so this relatively slight change in water level resulted in a relatively large change in water levels relative to the ground surface, with water levels being generally below ground surface under breached RIM and water levels being generally above ground surface under RIM when the impoundment was flooded.

Under breached RIM, there were diverse shallow salinity environments throughout the impoundment, ranging from salinities lower than sea water under the uplands to salinities higher than sea water in the infrequently flooded parts of the mangrove (Figure 3-5A). Stringer et al. (2010) attributed both, at least in part, to infrequent flooding by tidal waters, with the former being lower than sea water because of the effects of a fresh-water lens fed by rainwater and the latter being higher than sea water because of the effects of the evapoconcentration of ground water that flowed in from the Indian River Lagoon.
The extensive and long-term flooding under RIM changed these conditions, resulting in significant changes in shallow salinity distributions throughout the impoundment. Overall, there was little change in the shallow mean terrain conductivities (i.e., 1,868 mS m\(^{-1}\) to 1,825 mS m\(^{-1}\)). However, there was significant change in the variability and spatial structure of the shallow terrain conductivities. Under breached RIM, shallow terrain conductivities ranged from 116 to 2,917 mS m\(^{-1}\) with the highest shallow terrain conductivities being located along a NW to SE axis on the western side of the shoreline, closest to Indian River Lagoon (Figure 3-5A). Under RIM, shallow terrain conductivities ranged from 1,489 to 2,430 mS m\(^{-1}\) with little spatial structure to the shallow terrain conductivity distributions (Figure 3-5B). The change in structure was so significant that 500,000 random reassignments of the data could not produce the same \(\Delta CV\) that was calculated using the observed values (Figure 3-7A).

Conversely, the extensive and long-term flooding under RIM has not yet resulted in significant changes in deep salinity distributions throughout the impoundment. Though the spatial structure of the deep terrain conductivities visibly changed (Figure 6A and 6B), the change in the structure was not significant because 500,000 random reassignments of the data routinely produced the same \(\Delta CV\) that was calculated using the observed values (Figure 3-7B). However, the mean deep terrain conductivities did decrease from 328 to 255 mS m\(^{-1}\), suggesting that there may be a trend toward lower salinities and perhaps less spatially structured salinities in the deep subsurface of the impoundment.
References


CHAPTER 4:
WATER SOURCES IN MANGROVES IN THREE HYDROGEOMORPHIC SETTINGS ON THE COSTALEGRE, JALISCO, MEXICO

Abstract

We examined the fresh-water and seawater contributions to three distinct mangrove communities along the Costalegre on the central Pacific coast of Mexico. Surface-water, ground-water, fresh-water, and seawater samples were all collected over two sampling periods representing the wet and dry seasons. Salinities varied by water type, with values of ~9 in La Manzanilla, ~17 in La Vena, ~33 in Barra de Navidad, ~0.4 in the fresh waters, and ~34 in the seawater. Sodium and Chloride concentrations and isotopic signatures, as well as salinity, were used as tracers in mass-balance mixing models to quantify estimates of relative fresh-water and seawater contributions to each site. La Manzanilla, a basin mangrove, had mean fresh-water contribution estimates of 63-84%. La Vena, a riverine mangrove, had fresh-water estimates of 39-51%. Barra de Navidad, a fringe mangrove, had low fresh-water contributions of 0-5%. These results illustrate the varying role that groundwater plays in mangrove hydrodynamics, depending on the site hydrogeomorphology.
Introduction

More than one-third of the world’s population lives in coastal areas and small islands, which make up only 4% of Earth’s total land area (Brown et al. 2006). These large coastal populations stress the very natural resources on which they depend for the goods and services they provide (Barbier et al. 2008). For example, recent estimates report up to a 30% loss of coastal wetlands worldwide, largely due to anthropogenic activities (Solomon et al. 2007).

Mangroves are coastal ecosystems that cover ~180,000 km$^2$ of sheltered subtropical and tropical coasts between latitudes 24° north and south where mean annual temperatures are >20°C (Dawes 1998). Mangroves are just one example of coastal wetlands being rapidly lost. In recent decades, global mangrove area has declined ~35% overall at a rate of ~2.1% per year (Alongi 2002; Martinuzzi et al. 2009), while North and South American mangrove area has declined ~38% overall at a rate of ~3.6% per year (Valiela et al. 2001; Valiela et al. 2009). These high rates of loss make mangroves the most threatened major coastal habitat in the world (Valiela et al. 2009). As in the case of other coastal habitats, anthropogenic activities are largely responsible for the loss and degradation of mangrove ecosystems (Alongi 2002; Gunawardena and Rowan 2005; Islam and Haque 2004; Vijay et al. 2005).

The central Pacific coast of Mexico has ~64.5 km$^2$ of mangrove habitat, ~22 km$^2$ of which are in the state of Jalisco (RAMSAR 2010; CONABIO 2007). Much of this is on the Costalegre, a 260-km coastline from Puerto Vallarta, Jalisco in the north to Manzanillo, Colima in the south, that is 400 km from...
Guadalajara and 950 km from Mexico City. This coastal region has traditionally been home to small agricultural and fishing communities but, in recent years, has been one of eight targeted for development by the National Fund for Tourism Development and the Secretariat of Tourism, federal agencies responsible for tourism development throughout the nation (Wilson 2008). As a result, tourism has dramatically increased, with many of these small communities experiencing almost a complete transformation from economies based upon local agriculture and fishing to more complex economies based on tourism, real estate, and commerce (Everitt et al. 2008). Much of the related population growth is due to the addition of European and North American expatriates and retirees, most of whom expect higher standards of living than do the traditional residents. These shifts in population and demographics have increased the stresses placed on the natural resources, including the aforementioned mangrove environments.

These mangroves play vital roles in supporting both the regional ecology and economy, but sound management and conservation strategies are essential to ensure that they can continue to do so in spite of the rapid development and rapidly shifting demographics (Valiela et al. 2001; Alongi 2002; Martinuzzi et al. 2009). This need is amplified, because approximately one-third of the mangroves in Jalisco are designated as wetlands of international importance because of the significance of the ecological goods and services they provide (RAMSAR 2010).

The flux of water into and out of mangroves controls the physical and chemical hydrological characteristics of mangroves and facilitates the exchange
of mass, energy, and organisms between mangroves and the surrounding hydrological landscape (Twilley and Chen 1998). Mixing between fresh water and seawater plays a role in controlling salinity (Drexler and De Carlo 2007; Stringer et al. 2010). Plant community composition (Odum and McIvor 1990; Stringer et al. 2010) and primary productivity (Suarez and Medina 2005; Suarez and Medina 2006; Cardona-Olarte et al. 2006) are correlated with salinity. Also, nutrient-rich fresh water flowing into mangroves can enhance mangrove primary productivity (McKee 1995) while nutrient-rich mangrove water flowing into lagoons and near-shore marine environments can enhance water-column primary productivity (Rivera-Monroy et al. 1998).

As such, understanding the water sources in mangroves is a critical first step in developing sound management strategies. Water sources can vary, with the relative contributions of fresh water and seawater varying depending on climate, geology, and the proximity to the coast (Drexler and De Carlo 2002; Barlow 2004; Stringer et al. 2010). Source water contributions also can vary tremendously between different mangrove functional environments in close proximity to one another (Drexler and De Carlo 2002). However, detailed studies are time consuming and expensive, so there is a need for simple conceptual models that can be used to rapidly infer water sources and better inform land-use planning and decision-making, particularly lesser-developed nations.

One such simple conceptual model might be developed from the hydrogeomorphic-based mangrove classification first proposed by Lugo and Snedaker (1974) and modified thereafter, most recently by Woodroffe (2002).
The classification is typically displayed as a ternary diagram, with the vertices representing interior, river-dominated, and tide-dominated mangroves, where interior mangroves are closed-basins, river-dominated mangroves are dominated by unidirectional river flows, and tide-dominated mangroves are characterized by bidirectional tidal flows. The purpose of this study is to quantify the relative contributions of fresh water and seawater to each of these different mangroves classes, using three mangroves on the Costalegre as case studies. To that end, we used basic water parameters (salinity), conservative ion concentrations (sodium and chloride), and stable isotopes (deuterium and oxygen-18) as input data to mass-balance mixing models to determine the relative contributions of fresh water and seawater to each mangrove setting.

**Study Site Description**

This study was conducted in three separate mangroves located on the Costalegre on the central Pacific coast of Mexico in the state of Jalisco (Figure 4-1). This region lies at the western edge of the Sierra Madre del Sur, which is characterized by alternating basins and ridges with few large alluvial plains. During the middle Tertiary period, the region was subject to widespread volcanic activity, resulting in massive intrusions of magma reaching the surface in batholiths-sized domes and shields (Barrera and Vargas 2007). All of this material cooled and solidified, forming the rhyolites, dacites, adesites, basalts, igneous breccias, and tufts that now make up the rugged basin and ridge
Figure 4-1. Site map showing general location of the Costalegre along the central Pacific coast of Mexico. The three insets show each of the three mangrove study sites: A) La Vena, B) La Manzanilla, and C) Barra de Navidad. Each sampling point is labeled with the name used to identify it throughout the paper.
topography of the modern Sierra Madre del Sur. The bays on the Costalegre are formed by the intersection of the alternating basins and ridges and the Pacific Ocean, with the basins providing protected depositional environments in which form coastal alluvial plains on which Quaternary stream and oceanic sediments are deposited (Barrera and Vargas 2007). All of the study sites are located in these protected bays, with two located in the Bahía of Tenacatita and one located in the Bahía de Navidad. The regional average temperature is 25.3 °C and yearly rainfall is 700-800 mm, occurring mostly in the summer months (SMN 2010).

The La Manzanilla mangrove is an interior mangrove typically impounded by a semi-permanent beach ridge (Figure 4-1). The La Manzanilla mangrove is ~2.5 km² in size, being ~4.3 km parallel to the shoreline and extending ~0.6 km inland (RAMSAR 2010). The Rio Purificación and other smaller, unnamed streams flow toward the La Manzanilla mangrove. However, these streams flow only ephemerally in response to wet-season (i.e., June-October) storms, so infrequently that defined channel beds and banks do not extend across the alluvial fan to the La Manzanilla mangrove. During particularly intense wet-season storms, such as tropical cyclones, the beach ridge can be breached forming an intermittent connection between the La Manzanilla mangrove and the Bahía de Tenacatita. The predominant mangrove species are the white mangrove (*Laguncularia racemosa* L.) and the red mangrove (*Rhizophora mangle* L.) along with the mangrove associate, buttonwood (*Conocarpus*
erectus) (RAMSAR 2010). The La Manzanilla mangrove is a RAMSAR site, having received this status in 2008.

The La Vena mangrove is a river-dominated mangrove perennially connected to the Bahía de Tenacatita (Figure 4-1). The La Vena mangrove is ~0.6 km$^2$ in size but is long and linear, being ~5.0 km in length and ~0.1 km in width. No streams flow into the La Vena mangrove. Instead, the La Vena mangrove is fed by ground-water discharge from a headwater spring which maintains perennial flows and a perennial connection between the La Vena mangrove and the Bahía de Tenacatita. The predominant mangrove species is the red mangrove (*Rhizophora mangle* L.).

The Barra de Navidad mangrove is a tide-dominated mangrove located on Laguna de Navidad, a small, shallow lagoon perennially connected to the Bahía de Navidad (Figure 4-1). The lagoon is ~3.7 km$^2$ in area and is fringed by ~5.7 km$^2$ of mangroves, much of which being located on a broad delta prograding into the lagoon (CONABIO 2007; Mendez-Linares et al. 2007; RAMSAR 2010). The Rio Arroyo Seco, the Canal de Marabasco, and other smaller, unnamed streams flow into the Barra de Navidad mangrove, with a combined watershed area of 3.9 km$^2$ (RAMSAR 2010). The predominant mangrove species are the white mangrove (*Laguncularia racemosa* L.) and the red mangrove (*Rhizophora mangle* L.) along with the mangrove-associate buttonwood (*Conocarpus erectus*) (RAMSAR 2010). The Barra de Navidad mangrove is a RAMSAR site, having received this status in 2008.
Methods

Water samples were collected and analyzed for temperature, pH, and salinity in the dry and wet seasons (i.e., January and July, respectively) in 2007; and temperature, pH, salinity, sodium and chloride concentrations, and stable isotopes (deuterium and oxygen-18) in the dry and wet seasons (i.e., January and July, respectively) in 2008. The lone exception was that no water samples were collected from the Barra de Navidad study site in the dry season of 2007. Water samples were collected from both in, and the areas surrounding, the mangroves. In the mangroves, water samples were collected from surface-water bodies in all three mangroves and from shallow piezometers in the La Manzanilla mangrove (i.e., mangrove waters). Around the mangroves, water samples were collected opportunistically in a variety of irrigation and drinking-water wells, springs and seeps, and rivers (i.e., fresh waters) and in the open waters in the Bahía de Tenacatita (i.e., seawaters) (Figure 4-1). For each piezometer and well sample, approximately three volumes of water were pumped prior to the collection of samples for chemical analyses. Samples were pumped through 0.45 μm in-line filters (Whatman, Maidstone, Kent, England) directly into pre-cleaned, acid-washed HDPE sample bottles. Chloride samples were acidified with 1 ml of nitric acid, and chloride and sodium samples were stored at the method-required range of 4 (±2) ºC prior to analyses. Stable isotope sample bottles were filled completely with negligible head space and sealed with Parafilm (American National Can, Chicago, Illinois, USA) to prevent the sample from equilibrating with ambient air.
Temperature, pH, and salinity of the surface-water and ground-water samples were measured in the field with an YSI 556 MPS (YSI Inc., Yellow Springs, Ohio, USA). Sodium analysis was conducted by the University of South Florida Center for Water Analysis, with concentrations determined by atomic absorption spectrometry following EPA method 273.1 (Clesceri et al. 1998). Chloride analysis was conducted by Advanced Environmental Laboratories, Inc. in Tampa, Florida, with concentrations determined by the ferricyanide method following EPA method 325.2 (Clesceri et al. 1998). Based upon replicate analysis of 10% of the samples, analytical precision was typically better than 5 percent. All sodium and chloride concentrations are reported as ppm.

Stable isotope analyses were conducted at the UC Davis Department of Geology Stable Isotope Laboratory. Deuterium analyses were performed using the chromium reduction method (Donnelly et al. 2001), while oxygen-18 analyses were performed using the carbon dioxide equilibration technique (Epstein and Mayeda 1953). Deuterium and oxygen-18 are reported in the conventional, delta notation (δ):

\[ \delta = \left( \frac{R_{\text{sample}}}{R_{\text{standard}}} - 1 \right) \times 1000 \]

where \( R \) is the ratio D/H or \(^{18}\text{O} / ^{16}\text{O}\) for deuterium and oxygen-18, respectively (Craig 1961). The resulting sample values of \( \delta D \) and \( \delta ^{18}\text{O} \) are reported in per mil (‰) deviation relative to Vienna Standard Mean Ocean Water (VSMOW) and, by convention the \( \delta D \) and \( \delta ^{18}\text{O} \) of VSMOW are set at 0‰ VSMOW (Gonfiantini
Analytical precisions were ±1.0‰ and ±0.05‰ for δD and δ\textsuperscript{18}O, respectively.

The relative contributions of fresh water and seawater to surface-water samples taken within the mangrove were estimated using a series of two-end, mass-balance mixing models using salinity, sodium, chloride, δD, and δ\textsuperscript{18}O as conservative natural tracers. The mass-balance mixing model was

\[ C_{FW}(f_{FW}) + C_{SW}(f_{SW}) = C_M \]

where \( C \) is salinity (psu), sodium or chloride concentration (ppm), or δD or δ\textsuperscript{18}O (per mil), and \( f \) represents the fraction of end member present. The subscripts \( FW, SW, \) and \( M \) refer to fresh water, seawater, and mangrove, respectively.

**Results**

**Salinity**

The salinities of the mangrove waters varied between study sites and were generally intermediate between the fresh waters and seawaters (Table 1; Figure 4-2). Salinity is the ratio of the electrical conductivity of a given water sample to the electrical conductivity of a standard potassium chloride solution. Ratios have no units, but salinity nevertheless is typically expressed in terms of practical salinity units (psu), and that convention is followed here. Mean ± SD salinities were 9.1 ± 9.3 psu in the La Manzanilla mangrove (\( n = 50 \)), 16.6 ± 10.4 psu in the La Vena mangrove (\( n = 33 \)), and 33.0 ± 3.0 psu in the the Barra de Navidad mangrove (\( n = 19 \)). Conversely, the salinities of the seawaters and
The sodium and chloride concentrations of the mangrove waters varied between study sites and were generally intermediate between the fresh waters and seawaters (Table 1; Figure 4-3). Mean ± SD sodium and chloride concentrations were, respectively, 2,770 ± 2,850 ppm and 6,940 ± 7,150 ppm in the La Manzanilla mangrove (n = 31), 4,700 ± 3,210 ppm and 10,100 ± 6,850 ppm in the La Vena mangrove (n = 15), and 9,040 ± 500 ppm and 19,500 ± 2,300 ppm in the the Barra de Navidad mangrove (n = 15). These were bracketed by the fresh waters and seawaters, with mean ± SD sodium and chloride concentrations being, respectively, 48 ± 40 ppm and 123 ± 156 ppm for fresh waters (n = 25) and 8,800 ± 180 ppm and 20,500 ± 580 ppm for seawater (n = 5). Linear regression of the sodium and chloride concentrations yields a local mixing line ($R^2 = 0.95; p<0.01$).
Figure 4-2. Average ground-water and surface-water salinity values for each wet and dry sampling event in each of the three mangroves and in the source waters. Bars are the mean salinity for all sites sampled in that mangrove and are arranged in ascending date from left to right. Error bars are the standard deviations of those means (± 1σ). No data were collected in the Barra de Navidad mangrove in January 2007.

Stable Isotopes

The δD and δ¹⁸O of the mangrove waters varied between study sites and were generally intermediate between the fresh waters and seawaters (Table 1; Figure 4-4). Mean ± SD δD and δ¹⁸O were, respectively, -39.86‰ ± 39.33 and -5.44‰ ± 5.35 in the La Manzanilla mangrove (n = 31), -26.55‰ ± 16.99 and -3.75‰ ± 2.43 in the La Vena mangrove (n = 15), and -2.01‰ ± 3.36 and -0.17‰ ± 0.42 in the the Barra de Navidad mangrove (n = 15). These were bracketed by the fresh waters and seawaters, with mean ± SD δD and δ¹⁸O being -61.65‰ ± 4.04 and -8.83‰ ± 0.66 for fresh waters (n = 25) and -0.83‰ ± 1.08 and -0.03‰
Figure 4-3. Sodium and chloride concentrations (ppm) of fresh waters, seawaters, and mangrove waters collected from each of the three study sites during the course of the study. The symbol shows the mean concentration and the error bars represent the standard deviations (± 1σ) of those mean values. The dashed line is the local mixing line ($R^2 = 0.95; p<0.01$) determined via linear regression of the sample concentrations.

± 0.10 for seawaters. Isotope data plot along a local mixing line ($R^2 = 0.99$, $p<0.01$) anchored by the fresh waters and seawaters, with fresh waters plotting on the Global Meteoric Water Line (Craig 1961).

**Mass-Balance Mixing**

The contribution of the fresh water end member was determined for each site using each of the five tracers: salinity, sodium concentration, chloride concentration, δD, and δ¹⁸O (Figure 4-5). The La Manzanilla mangrove had the highest fresh-water contributions, with site-averaged means ranging from 63-84%; and the La Vena mangrove had intermediate fresh-water contributions, with
Figure 4-4. $\delta^D$ and $\delta^{18}O$ (‰) of fresh waters, seawaters, and mangrove waters collected from each of the three study sites during the course of the study. The symbol represents the mean composition and the error bars represent the standard deviations ($\pm1\sigma$) of those mean values. The solid black line is the global meteoric water line as defined by Craig (1961). The dashed line is the local mixing line ($R^2 = 0.99; p<0.01$), determined via linear regression of the sample isotopic compositions.

site-averaged means ranging from 39-51%; and the Barra de Navidad mangrove had the lowest fresh-water contributions, with site-averaged means ranging from 0-5%. At each site, the salinity-based mixing models yielded the highest mean estimates and the isotope-based mixing models yielded the lowest mean estimates.

**Spatial Variability**

The mass-balance mixing estimates for the La Manzanilla and La Vena mangrove surface waters show a large variability in fresh-water contributions,
while the Barra de Navidad mangrove have a relatively narrow range (Figure 4-6). In La Manzanilla mangrove, the largest fresh-water contributions were to surface waters furthest from the ocean at the toeslope of the mountainous terrain (~90%), while the largest seawater contributions were to surface waters closest to the beach ridge (~30-40%). In the La Vena mangrove, the largest fresh-water contributions were to surface waters near the spring vent (~75-80%), while the largest seawater contributions were to surface waters closest to the mouth (~82-99%), with this large variation possibly being due to varying stream and tidal flows at different sampling times. In the Barra de Navidad mangrove, there was little fresh-water contribution, with seawater contributions high throughout the mangrove (~99-100%).

Discussion

Fresh-water contributions to mangroves exhibit both an intra- and inter-site variability, even when those sites are in close proximity to one another, as is the case with the La Manzanilla, La Vena, and Barra de Navidad mangroves (Figure 4-5 and Figure 4-6). The La Manzanilla mangrove has a largest percentage of fresh-water contribution; the La Vena mangrove has a slightly lower percentage of fresh-water contribution, with amounts varying in relation to their position on the river between the headwater spring and mouth leading to Bahia Tenacatita; and the Barra de Navidad mangrove, has a negligible percentage of fresh-water contribution at any sampling location.
Figure 4-5. Fresh-water fraction of mangrove waters in each of the three study sites using each of the five tracers. Bars are the mean fresh-water contribution for all sample locations. Error bars are the standard deviations of those means (± 1σ).

The fresh-water contributions to these mangroves can be understood in the context of hydrogeomorphic setting, as defined by Woodroffe (2002). The La Manzanilla mangrove is an interior mangrove, with a semi-permanent beach ridge and only ephemeral connections to the Bahía de Tenacatita. These ephemeral connections typically occur following low-frequency/high-magnitude storms, such as tropical cyclones, during which fresh water inflows to the mangrove accumulate and ultimately break through the beach ridge. During these connections, fresh-water outflow dominates exchange between the mangrove and the Bahía de Tenacatita. For most of the year, and for all of the
year in some years, the beach ridge remains intact and fresh-water inflows, presumably from ground-water discharge, dominate mangrove-water salinities. The La Vena mangrove is a river-dominated mangrove, with ground-water discharge from a headwater spring maintaining perennial flows and a perennial

Figure 4-6. Fresh-water fraction of mangrove waters at each individual sampling location. Fractions are means of fractions from all five tracers.
connection to the Bahía de Tenacatita. Fresh water dominates at the headwaters, seawater dominates at the mouth, and fresh water and seawater mix in the middle reaches of the river. The Barra de Navidad mangrove is a tide-dominated mangrove, with fringing mangroves located on the margins of a small, shallow lagoon perennially connected to the Bahía de Navidad. Though there are some inflowing streams and canals, i.e., the Rio Arroyo Seco, the Canal de Marabasco, and other smaller, unnamed streams, these have a combined watershed area of 3.9 km$^2$ so large surface-water inflows are ephemeral, largely following low-frequency/high-magnitude storms such as tropical cyclones. The mouth is wide and deep, which allows a free exchange of seawater between the ocean and the mangrove so fresh-water inflows are quickly mixed into the seawater and seawaters dominate mangrove-water salinities. Therefore, hydrogeomorphic setting, as represented in the simple conceptual model most recently proposed by Woodruffe (2002), can be used as a tool to rapidly estimate the importance of fresh-water inflows to a given mangrove and therefore the potential susceptibility of a given mangrove to changes in fresh-water inflows due to water-resources development.

Fresh-water inflow to the mangrove can follow a few basic pathways: direct precipitation, surface-water runoff, and ground-water discharge. Ground-water discharge almost certainly plays a large role in this region. Rainfall and runoff are limited, ephemeral, and focused in the wet season (July-October). Despite the seasonality of the rainfall, the mangrove-water salinities lack a clear
seasonal trend (Figure 4-2), meaning that there must be another consistent source of fresh water to the mangrove maintaining the mangrove-water salinities.

Mangroves are commonly found inset into high-transmissivity coastal plain aquifers. The La Manzanilla mangrove, for example, is inset into a coastal plain with an unconfined aquifer with a shallow water table and a transmissivity of 3,360 m² d⁻¹ (M.C. Rains, unpublished data). Additionally, mangroves are located at the terrestrial-marine interface where ground water is forced upward to ground-water discharge points near the shoreline (Cady 1941; McBride and Pfannkuch 1975; Harvey and Odum 1990; Rains et al. 2004). Both of these characteristics illustrate how geologic conditions favor ground-water discharge into these mangrove environments. Additionally, ground-water discharge can also be directly observed in these sites, such as the headwater spring in the La Vena mangrove and the many seeps and springs in the mountainous terrain surrounding the La Manzanilla mangrove.

There are few perennial streams and surface-water storage reservoirs in the region. Therefore, ground-water is an important source of water for irrigation and/or municipal supply. There is extensive heuristic evidence of the depletion of ground-water resources in these small coastal plain aquifers. For example, salinity increased in two water-supply wells in La Manzanilla between January 2007 and July 2008. One well is under municipal control, and ground-water pumping appears to be causing salt-water intrusion, with wet-season salinity increasing from 1.1 in January 2007 to 2.2 in January 2008 and dry season salinity increasing from 2.5 in July 2007 to 2.9 in July 2008 (M.C. Rains,
unpublished data). The other well is under private control and is lacking in salinity data, but ground-water pumping nevertheless appears to be causing salt-water intrusion, with the well having been abandoned and replaced with a shallower well as of January 2010. If ground-water resources continue to be stressed by burgeoning populations, then heads in the coastal plain aquifers will be expected to decline potentially resulting in decreases in the ground-water discharge to mangroves. This might lead to changes in salinity and related ecosystem structure and function in those mangroves in which fresh water inflows play an important role.

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