Imaging Wetland Hydrogeophysics: Applications of Critical Zone Hydrogeophysics to Better Understand Hydrogeologic Conditions in Coastal and Inland Wetlands and Waters

Christine Marie Downs
University of South Florida, cmcniff@mail.usf.edu

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Imaging Wetland Hydrogeology: Applications of Critical Zone Hydrogeophysics to Better Understand Hydrogeologic Conditions in Coastal and Inland Wetlands and Waters

by

Christine Marie Downs

A dissertation submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy
School of Geosciences
College of Arts and Sciences
University of South Florida

Major Professor: Sarah E. Kruse, Ph.D.
Mark Rains, Ph.D.
Chester Weiss, Ph.D.
Thomas Crisman, Ph.D.
Mahmood Nachabe, Ph.D.

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Abstract

This dissertation consists of three projects utilizing electric and electromagnetic (EM) methods to better understand critical-zone hydrogeologic conditions in select Florida wetlands and waters. First, a time-lapse electrical resistivity (ER) survey was conducted in section of mangrove forest on a barrier island in southeast Florida to image changes in pore-water salinity in the root zone. ER data show the most variability in the root zone over a 24-hour period, and, generally, the ground is more resistive during the day than overnight. Second, a suite of three-dimensional forward models, based on varying lateral boundaries and conductivities typical of a coastal wetland, were run to simulate the EM response of a commercial electromagnetic induction instrument crossing over said boundaries. Normalized profiles show the transition is sharper in a hypersaline regime than one where freshwater and clay are present. Furthermore, enough variability exists in hypersaline regimes to justify collecting profile measurements in multiple coil configurations to constrain the nature of a lateral boundary. Also, under certain circumstances, there are kinks in the EMI response even across abrupt boundaries due to concentrated current density at a layer’s edge. Lastly, geophysical surveys were conducted at six wetlands in west-central Florida to characterize potential hydrostratigraphic units and compare/contrast them to the current conceptual model for cypress dome wetlands. ER was used to image the geometry of the top of limestone; ground penetrating radar (GPR) was used to image stratigraphy beneath and surrounding wetlands. These wetlands can be grouped into two models. Topographic highs surrounding wetlands are controlled by the undulating top of limestone at sites where the region is characterized by limestone ridges. In contrast, topographic highs are controlled by thick sand packages at sites regionally characterized by sand dunes over scoured limestone.
1. Introduction

1.1 Critical Zone Hydrogeophysics

This dissertation consists of three projects utilizing electric and electromagnetic (EM) methods to better understand critical-zone hydrogeologic conditions in coastal and inland wetlands and waters. Although these are three distinct studies, the ability to utilize various near-surface geophysical methods in critical zone hydrogeological research is the overarching theme. The chapters are organized in terms of the depth of investigation. Chapter 2 focuses on pore water in the uppermost five meters of a coastal wetland environment. Chapter 3 consists of models where the signal response is a function of up to 10 meters of subsurface and lateral boundaries 2-7 meters deep. The data in chapter 4 image hydrostratigraphic units as deep as 30 m.

1.2 Mangrove forest, North Hutchinson Island, FL

Chapter 2 focuses on pore water in the uppermost five meters of a coastal wetland environments. This study seeks to characterize the possible role mangrove trees play in diel changes in soil salinity with a time-series electrical resistivity survey and continuous groundwater sampling.

Mangrove forests, often simply referred to as “mangroves”, consist of woody, medium height trees and shrubs growing in dense, saline woodland habitats. Mangrove trees are integral components of tropical intertidal forest communities and are essential to a number of coastal stabilization and ecological processes. They can survive in environments of a wide range of salinities, but are often found in saline to hypersaline habitats because there is minimal competition for resources. The movement of salt within an individual tree is well documented in the literature (Ball, 1988; Esteban et al., 2013; Parida and Jha, 2010, among
Mechanisms vary among species, but generally excess salt is either excluded at the leaf, stored in succulent leaves, or excreted at the leaf. Black mangrove (*Avicennia germinans* (*L.*)) is the dominant mangrove species at the site of the study described in chapter 2. This species relies on salt excretion at the leaves to cope with high salt volumes (Drennan and Pammenter, 1982; Esteban et al., 2013; Parida and Jha, 2010; Scholander et al., 1962). Though root exclusion is minimal, it produces a shallow hypersaline zone around the root system, which can be laterally extensive due to the pneumatophore roots (Ball, 1988). A maximum amount of salt is excreted during stressful periods (dry season, noon) (Esteban et al., 2013). The salt dissolves during cooler, moister periods. It will eventually return to the soil during rain events or via excised leaves. Overall, the salt balance in a given tree is purportedly maintained, but the subsurface experiences a net gain in salt concentrations over time, in the absence of additional processes that mobilize saline pore water.

Typical estuarine water circulation models show tidal forcing (high tide pushing water landward both above and below the surface and reversing gradient at low tide) to be the dominant mechanism of pore water (and salt) movement in saturated and intertidal zones (Wolanski, 1992; Hughes et al., 1998; Sam and Ridd, 1998). Evapotranspiration (ET) rates, including the processes described above, are highly influential in pore water salinities specifically during hot, dry periods. However, salinity balances sustained under natural conditions can be altered through anthropogenic and climate-change processes.

The current study area is tidally “restricted” due to an earthen dike acting as a barrier between the brackish lagoon and the mangrove forest. Developed in 1970, Imp-24 is part of a system of impoundments constructed around mangrove forests that dominated the lagoon side of North Hutchinson Island—a siliciclastic barrier island of the east coast of southeast Florida (Rey et al., 1990b; Brockmeyer Jr et al., 1996; Rey et al., 2009; Rey and Rutledge, 2009). Tidal exchange was partially restored in 1985 via an open culvert installed through the dike (a practice called breached rotational impound management) and four more culverts were installed in 1987. In 2009, pumps were installed so brackish lagoon water could
be pumped into the impoundment during the summer season (a practice called rotational impound management). Maximum water levels are maintained throughout the summer and the impoundment is allowed to dry up during the winter.

The impoundment’s barrier decreases mangrove growth rates and diversity (Crase et al., 2013; Harris et al., 2010; Rey et al., 1990a; Stringer et al., 2010) and shuts out fish populations (Rey et al., 1990a). Additionally, since strong tidal currents are reduced, groundwater must play a major role in hydrodynamics, and pore water salinities are more heavily affected by ET rates (Stringer et al., 2010). Throughout the mangrove, ET lowers the water table and creates a gradient that drives lagoon water to the mangrove via groundwater flow. Both processes introduce salt into the deeper subsurface. Dense, saline pore water sinks and creates a hypersaline layer about 1-4 m below the mangrove (Stringer et al., 2010). The ultimate fate of this excess local salt is, to our knowledge, poorly understood.

In an effort to capture a spatiotemporal changes in this pore-water salt, an electric resistivity array was repeated along a constant transect fourteen times over a 24-hour period. Direct groundwater sampling during the survey provided pore-water temperature, conductivity, and water levels. An improved understanding of processes at this site will help further assess the long-term effects of diking on a mangrove forest and their ability to adapt to and mitigate climate change.

### 1.3 Three-dimensional Electromagnetic Induction Modeling

Chapter 3 presents a suite of 3D forward models from a finite volume electromagnetic induction (EMI) solver. The response of the commercial EM31 instrument by Geonics, Inc. was simulated for a profile going over lateral boundaries in regimes where the low induction number may or may not be valid. The conductivity models were designed to simulate what one would expect to see a coastal wetlands (i.e. clay lenses, saltwater- saturated sands).

In coastal wetland environments, where the clay lenses, discontinuous layered material, saltwater or freshwater lenses exist, it is important to have a good understanding of subsurface heterogeneities that may influence local flow regimes. Obtaining a dense enough
dataset via direct measurements to quantify this heterogeneity is labor–, time–, and resource–

intensive. EMI is a useful reconnaissance method that can alleviate this effort. EMI data can

serve as good discriminators between conductivity contrasts (i.e. saltwater versus freshwater,

clay versus sand). However, the nature of a boundaries (i.e. sharp versus gradational) are

not easily discernable in EMI data.

To further complicate the problem, most of the readouts of EMI commercial instrumentation

and data interpretation are designed to work in environments that satisfy conditions referred to as the low induction number (LIN). The LIN approximation is valid when

the transmitter-receiver coil spacing is much smaller than the depth at which the attenuated transmitted signal is approximately $1/e$ its amplitude at the transmitter. For a given instrument with a fixed coil spacing the LIN is a primarily a function of operating frequency and bulk ground electrical conductivity. Generally, zone with conductivities higher than 80 mS/m will violate the LIN assumption. For a setting with clay layering within sand deposits

the LIN is valid. However, for zones of saltwater-saturated sand in a freshwater background

where saltwater conductivity is on the order of 5000 mS/m, the LIN is clearly not valid.

A forward solution is necessary to address the breakdown of the LIN and has already been done in 1D as a layered Earth (McNeill, 1980; Ward and Hohmann, 1988; Callagery et al., 2007) for a 1D homogeneous halfspace. Since the world is 3D, many EMI mapping surveys are appropriately conducted as a series of 2D profiles and a 2D or 3D forward model is desirable. The author is not aware of 2D solutions for 3D EMI problems, such as a profile over a 3D layered Earth. Thus a suite of 3D forward models were created to simulate the EMI response of conductivity contrasts typical of coastal, wetland environments.

The EMI forward modeling program by Weiss (2013) was used to run the simulations. APhiD is a finite volume numerical solver that uses a staggered grid (both nodes and cell edges are assigned values) over a Cartesian model domain. The user assigns true conductivity values to cells in a Cartesian model domain and the location, orientation and frequency of
the EM source. APhiD outputs the x, y, and z components of the total electric field (edge-centered) and magnetic field (face-centered).

The focus was on the EMI response from the conductivity structures in the first few meters beneath the ground surface so the mesh was extended to a depth of at least twice the skin depth (~50 m) to avoid any source energy reflecting back to the area of interest (a.k.a. edge effects.) A similar approach is taken to determine the minimum lateral extent of the model so interpolated profiles are not affected by reflecting energy.

Simulating a survey profile required at least one model run per data point along the profile. Each run considered one source frequency, location, and orientation. So simulating a profile of, for example, 20 data points with four soundings each required 80 models. The input and command files were built iteratively via a Python script and multiple models with different source input were run simultaneously via OpenMPI.

Each run provided a three-dimensional response to the source for the whole domain. Out of the whole domain, only the total magnetic field (B) value at the presumed receiver was needed. This was a distance equal to the intercoil spacing from the source (the simulated transmitter). Again, this had to be done for each model run.

1.4 West-central Florida Wetland Hydrogeology

Chapter 4 investigates how a subset of wetlands in xeric environments that are surrounded by uplands in west-central Florida may or may not differ from the accepted model for regional wetlands in terms of the hydrostratigraphic unit beneath and surrounding wetlands.

West-central Florida is characterized by a mantled karst terrain composed of Cenozoic sediments and sedimentary rocks and wetlands are very common throughouth the region. Fine sands and clays overlie highly-weathered limestone with a regionally-discontinuous clay unit that is sometimes present between them. The top of limestone is extremely irregular and can outcrop at locations with little to no overburden.
This overall description of the geology is a good summary of regional trends, but may or may not account for local variations beneath and surrounding wetlands, in particular, a subset of wetlands in xeric environments surrounded by uplands. The underlying lithology, and thus how it relates to hydrogeology, is not well understood.

In an effort to better contrain the underlying geology beneath these wetlands, a series of ground penetrating radar (GPR) and electric resistivity (ER) surveys were conducted at five wetland sites (Figure 4.1). Geophysical data and borehole logs were used to identify the geometry of strata and locate the top of weathered limestone beneath the wetlands. Interpretations were also used to compare and contrast these wetlands to cypress dome wetlands also found in the region.

1.5 References


2. Time-lapse Electric Resistivity Imaging Subsurface Salt Mobilization in an Impounded Mangrove Forest

2.1 Abstract

A time-lapse electrical resistivity (ER) survey was conducted in section of mangrove forest on a barrier island in southeast Florida. The objective was to image changes in pore water salinity potentially due to mangrove tree physiology. Groundwater measurements show the ground to be saturated for the duration of the survey. ER data show that over a 24 hour period there is more variability in the root zone than elsewhere along the transect. Around the roots, a decrease in resistivity (increase in salinity) is observed two hours after sunset. An increase in resistivity (decrease in salinity) is observed one hour after sunrise. Changes in resistivity may indicate water uptake or distribution from tree roots. However, hydraulic flow models, constrained by the geophysical data, are necessary before biological mechanisms can be identified.

2.2 Introduction

Mangrove forests are integral components of tropical and subtropical intertidal forest communities and essential to coastal stabilization and ecological processes. Due to their location between land and sea, mangrove forests play a pivotal role in climate change adaptation and mitigation. They show a great deal of ecological stability (Alongi, 2015) through their ability to elevate land surface via sediment accretion to potentially keep pace with sea level rise (McIvor et al., 2013; Krauss et al., 2014) as well as their capacity to act as CO2 sinks (Duarte et al., 2013). Mangrove trees have developed salt tolerance mechanisms to survive harsh coastal environments (e.g. Ball, 1988; Esteban et al., 2013; Parida and Jha, 2010). Black mangroves rely primarily on salt excretion at the leaves (Drennan and Pam-
menter, 1982; Esteban et al., 2013; Parida and Jha, 2010; Scholander et al., 1962). Excreted salt dissolves during cooler, moister periods and returns to the soil during rain events or via excised leaves. Red and white mangroves, on the other hand, exclude salt at roots, including pneumatophores, which further contributes to a local salt excess (Ball, 1988).

Tidal forcing is the typically the dominant mechanism of pore water and salt movement in the saturated and intertidal zones of coastal wetlands (Wolanski, 1992; Hughes et al., 1998; Sam and Ridd, 1998). However, impoundment practices alter the salinity balance maintained under natural conditions. Earthen dikes surround sections of wetlands and hydrologically isolate them from tidal surface water and each other. Thus evapotranspiration (ET) can become the most influential factor in pore water salinities, particularly during hot, dry periods (Stringer et al., 2010). This has the potential to create hypersaline conditions in the subsurface. While the salt balance in a given tree is presumably maintained, the subsurface experiences a net gain in salt concentration over time. Among the major factors determining the growth and production of mangrove species, groundwater salinity significantly influences mangrove forest production (Sam and Ridd, 1998; Ridd and Sam, 1996; Gordon, 1993; Naidoo et al., 2011; Ball, 1988; Esteban et al., 2013; Parida and Jha, 2010; Lovelock et al., 2006).

Resolving the spatiotemporal nature of groundwater salinity can be done through direct pore water measurements. However, capturing small scale spatial variability, like that expected in a vegetated area, requires labor–, time–, and resource–intensive data collection. Shallow geophysical methods can alleviate such an effort. Electric resistivity methods can evaluate ground resistivity (the inverse of conductivity), which is a function of lithology, degree of saturation, pore-water salinity, and temperature, among other factors. Repeated (time-lapse) electrical resistivity measurements along a constant transect can capture both spatial structure of ground resistivity due to lithology, pore-water salinity, water table, and temporal changes due temperature changes, shifts in the water table, and movement of salts in pore water.
This chapter reports on the observations made with a time-lapse electric resistivity (ER) survey coupled with real-time groundwater data and tidal records. ER has proven useful in saline environments and time-lapse surveys (Leroux and Dahlin, 2006; Kemna et al., 2002; Hayley et al., 2009; Urish and Frohlich, 1990; Attwa et al., 2011; Sutter and Ingham, 2016; Cassiani et al., 2006). The objective was to illustrate the spatiotemporal changes in saline pore water structures as a function of diel and tidal cycles in the absence of surface interaction between the impounded mangrove forest and lagoon. A diel cycle is defined as a 24 hour period. Improved understanding of processes at this site will help further assess the long-term effects of diking on a mangrove forest and their ability to adapt to and mitigate climate change.

Since these data were collected over a diel cycle, it is important to note how changes in salinity distribution relate to tide and sunrise/sunset. Transpiration significantly decreases at night when air temperatures drop. At this time, the water potential in the roots will rise to slow down or cease the intake of soil water. At the start of the day, transpiration resumes (reaches a maximum rate around midday) and the roots’ water potential drops for root uptake. Figure 2.1 provides a schematic. The ER data and interpretations were compared to this general model to see if the ER methods were able to image such variabilities around mangrove tree roots.

2.3 Field Site

The study was conducted in Imp-24, a mosquito-control impoundment on the west side on North Hutchinson Island, Florida (Figure 2.2). This is a siliciclastic barrier island approximately 35 km long and 0.7 km wide where the impoundment is located. The Indian River Lagoon, a tidal estuarine system, borders the west shore of the island; the Atlantic Ocean borders the east side. The lagoon connects to the Atlantic Ocean about 9 km south of Imp-24 and a lagged and dampened tidal signature is observed in lagoon water levels.

For the Indian River Lagoon, average precipitation rates are 1180 mm/yr (Sumner and Belaineh, 2005). Open waters evaporation rate is 1502–1614 mm/yr with the lowest
rate occurring in the winter months (Sumner and Belaineh, 2005; District, 2006). Estimates for vegetated areas are 3.5 mm/day of evapotranspiration (Lugo et al., 1975). Mangrove evapotranspiration (ET) rates strongly influence pore water salinities during hot, dry periods. Changes in water management practices in a mangrove forest along the lagoon have produced changes in the pore water salinity structure. The forest is separated from the neighboring intertidal lagoon by an earthen dike along the shoreline. Maximum water levels are maintained during the summer season (April-October) and allowed to dry during the winter (November-March) (Middleton et al., 2008; Rey and Rutledge, 2009; Connelly and Carlson, 2009). In the dry season, pore water salinities are more heavily affected by ET rate (Stringer et al., 2010), which lowers the water table and creates a gradient that drives lagoon water to the mangrove via groundwater flow.

The cross section in figure 2.3 was constructed using existing borehole data from Stringer et al. (2010). The portion that coincides with this study’s survey transect is mostly fine-grained sand composed of quartz and calcareous grains underlain by sandy clay interbedded with calcareous silts and sands. The depth of investigation for all resistivity data is ~11 m meaning one can expect to detect horizontal anomalies due to background geology in addition to pore water conductivity structures.

The very shallow subsurface can be partitioned into three vegetative zones. From the western end to seven meters along the transect, the surface is dominated by dense, dwarf mangroves. The central portion is a salt pan with standing roots sparse along the edges. At the center of the salt pan (about 30 m along the transect) there is a small shrub of mangrove very near the transect. From 40 m along the transect to the eastern end the vegetation is characterized by sparse, dwarf trees. Several trees were dug around to measure the depth to which the roots extend. The root zones of dense trees extend as far as 1.5 meters; smaller trees’ root zone extend up to about a meter.

Current water management practices pump brackish water into the impoundment to keep it inundated from April to October. It is then allowed to completely dry up from
November to March. It is already found that impoundment practices decrease mangrove growth rates and diversity (Crase et al., 2013; Harris et al., 2010; Rey et al., 1990a; Brockmeyer Jr et al., 1996; Rey et al., 1990b; Berrenstein et al., 2013; Middleton et al., 2008; Rey et al., 2009; Rey and Rutledge, 2009), which implies, among other things, a hypersaline subsurface.

2.4 Methods

Repeated electrical resistivity surveys were performed over a 24-hour cycle with an AGI Supersting R8 resistivity meter, using a combination of a dipole-dipole and inverse Schlumberger array. The electrodes remained stationary throughout the survey—56 electrodes at 1 meter spacing (Figure 2.4). There was standing water at certain times on the eastern half of the array so electrodes were cased in PVC pipes pushed approximately 3 cm into the ground to avoid losing current to surface water. Contact resistance tests confirmed the ground was wet enough to deem watering the electrodes with salt water unnecessary. Topography was measured along the transect using an auto-level, however, the very small relief was in the envelope of error, or 10 cm. Instead topography presented here comes from LiDAR of the area (District, 2007).

In an effort to capture both diel and tidal effects on salinity structures in the subsurface, an array was collected approximately every hour for a 24-hour period. However, a 2 hour gap exists due to a battery issue. It was important to have high lateral sensitivity in the root zone (0-2 m depth) as well as moderate vertical sensitivity in the first one to two meters. Thus the combination array was chosen for its balance between high sensitivity to lateral variations, moderate sensitivity to horizontal structures, and acquisition time (35-45 minutes) (Loke, 1999).

The resistivity data were inverted via time-lapse inversion using the RES2DINV software package from Geomoto, Inc. Due to a failing set of batteries, there are three surveys that are incomplete. These data were not included. Bad data points in each dataset were removed following initial inversion of all data points and removing points where the percent
difference between the logarithms of the measured and calculated apparent resistivity values are greater than 60%. The first of these trimmed datasets—collected just before solar noon on day 1—acts as a reference model. Damping factor in the inversion was set to increase with depth due to decreased resolution and at the sides where measurement outliers were observed. A small weight for the vertical filter was selected due to horizontally elongated anomalies that dominate the geology. The objective here is to track temporal changes rather than present absolute resistivity images.

The robust, L1 norm time-lapse inversion method was used where the first time step served as a reference model. L1 norm methods better account for the large contrasts in the observed data (Chambers et al., 2004) and expected at boundaries such as lithological interfaces and between vegetated and non-vegetated zones. Robust constraints were introduced on both the data and the parameters. Time-lapse inversion is preferred because outliers or noise inherent in each dataset has less influence in the solution versus individual solutions (Binley et al., 1996; Loke, 1999). A time-lapse regularization parameter of 2 was used, which reduces with each time step. This value is a balance between minimizing the difference between time steps and data misfit (Rucker et al., 2011).

Direct groundwater data were also collected using three CTD (Decagon) sensors deployed down PVC wells. The PVC wells, installed four months earlier, are screened at 0.6 m, 0.8 m, and 1.2 m, and reside at the western edge of the transect. The sensors measure electric conductivity, and water level at 15-minute intervals over the same time period as resistivity data. These data allow us to estimate a formation factor and position of the water table. Sensors were calibrated the day before by taking measurements at room temperature with tap water.

2.5 Results

The average resistivity over the entire survey (all time steps) is $0.6323 \, \Omega \cdot m$ with a min/max of $0.168/6.178 \, \Omega \cdot m$. A six meter thick conductive layer overlays a resistive layer and is observed throughout the 24-hour period. This corresponds to a change in lithology—
shell-rich sand overlaying clayey sand (see figure 2.3). Although the resistivity around roots change over the course of 24 hours, they are consistently more resistive than the neighboring salt pan.

Water level data at the far western end of transect indicate the ground was saturated for the duration of the survey (Figure 2.5). There was approximately 4 cm of standing water at the western end of the transect at the start of the study. Surface waters levels gradually dropped until there was no surface water, but the ground was completely saturated throughout the study. Ground temperature was $23^\circ$ C at the start of the survey and decreased to $21^\circ$ C by the end (Figure 2.5). Using the linear temperature dependent model from Hayley et al. (2007) and the fractional change coefficient ($m=0.0187$) by Hayashi (2004) resistivity varies between 0 and 0.018 $\Omega \ast m$, which is within the envelope of noise for the data (0-0.15 $\Omega \ast m$).

Given full saturation the time-lapse resistivity data reflect pore water conductivity changes corresponding to salinity changes. Data are shown as the percent change between a given time step and the previous. An increase in resistivity is a positive change; decrease in resistivity is a negative change (figure 2.6).

Within the uppermost meter, the salt pan does not experience much variability from one time step to the next (figure 2.7, 0.5 m). The salt pan has a low resistivity due to high rates of evaporation concentrating salts in the immediate subsurface, however, that process which increases pore water salinity is not captured in this diel cycle.

The root zone of trees, measured as far as 1.5 meters down, experiences more local increases and decreases in resistivity compared to the salt pan until sunset (figure 2.6, 19:00-20:00 and 2.7, 1.5 m) when there is little change across both zones. By midnight, the area around roots experiences a decrease in resistivity. These low resistivity areas get larger with successive time steps and reach approximately six meters depth where a lithology contact exists (figure 2.6, 03:30-05:00).
This trend of decreasing resistivity and spreading continues until sunrise (figure 2.6, 07:00-08:00) at which point the uppermost five meters becomes more resistive. After the initial increase in resistivity, there is, once again, little to no change through to the end of the time-lapse survey.

2.6 Discussion

Throughout the diel period, salt pan conductivities remained relatively constant. Most of the significant temporal changes in the uppermost two meters occur in the root zone around mangrove trees. It is known that salt excretion rates are highest over night (e.g. Ball, 1988; Gordon, 1993; Naidoo et al., 2011; Esteban et al., 2013), and there was no rain event to rinse large amount of salt crystals on leaves or decrease the salinity of groundwater.

During the daylight hours, mangrove trees maintain a low water pressure potential through maximum evapotranspiration. This creates a negative pressure gradient which allows groundwater to enter the roots. Evapotranspiration slows down and/or ceases after sunset and the roots’ water pressure potential equals that of the soil or is higher. This results in no water intake or a positive gradient where water and salts in a tree enter the soil.

The overnight reduction in evapotranspiration rates and increased water pressure potential in the roots may allow water and salts to return back to the soil, which would decrease the resistivity of the ground. Overnight the decrease in resistivity (increase in salinity) from midnight to sunrise may suggest water and salts are (re)introduced to the subsurface via tree roots or the influx of saline groundwater in the direction normal to the ER transect. At six meters, the potentially sinking saline water extends laterally, which it likely due to flow being diverted at the sand-clayey sand interface.

The roots’ water pressure potential drops at sunrise, as evapotranspiration resumes, and water and salts are taken up by the roots. Increased resistivity (lower salinity) at this time may be due to fresher water being pulled in from below or in the direction normal to the ER transect.
Little change in resistivity at sunrise and sunset may suggest the rate of water uptake or return in the roots slowing down or ceasing when the trees’ water pressure gradient changes.

Figure 2.8 offers a possible interpretation of mangrove trees’ role in pore-water salinity. The mangrove trees appear to reverse their pressure gradient overnight rather than simply reduce water uptake. This is suggested by the decreased resistivity (increase salinity) from midnight to 07:00. This dense saline water returns to the soil and sinks down until it reaches the sand-clayey sand interface and is diverted laterally. Evapotranspiration and root water uptake resumes after sunrise, which pulls fresher groundwater up from below or from the a direction normal to the ER transect.

This process of water returning to the soil via tree roots is a process known as hydraulic redistribution, and has been observed in dry environments (e.g. Caldwell et al., 1998; Richards and Caldwell, 1987; Caldwell and Richards, 1989; Ludwig et al., 2003), temperate forests (e.g. Meinzer et al., 2004), and savannas (e.g. Meinzer et al., 2004; Scholz et al., 2002).

Only one study observed redistribution in a saturated, saline wetland environments (Hao et al., 2009). Hao et al. (2009) found that reverse flow in red mangroves in south Florida may help to relieve the adverse effects of hypersaline conditions in a given tree, but also may be one of multiple mechanism causing dwarfism (impeded growth) in mangrove trees. While the hypersaline conditions in the impounded mangrove forest discussed here are detrimental to mangrove tree growth, hydraulic redistribution may be an important factor for its survival.

This chapter is the first time potentially similar observations have been made using geophysics. However, cross sectional hydraulic flow models, constrained by the geophysical data presented here, are necessary to confirm the suggested interpretations.

2.7 Conclusions

Time-lapse electrical resistivity surveys are able to capture dynamic processes over a diel cycle. With the aid of groundwater water sampling, one can narrow down the mechanism
responsible for resistivity changes. Transects that cross both vegetated and non-vegetated areas of the forest allow one to see changes that may be related to tree root activity. Hydraulic models, constrained by geophysical data, can provide additional information necessary to better understand the role in mangrove physiology in pore-water salt distribution.

2.8 References


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2.9 Figures

Figure 2.1: Simplified diel salt cycle with a black mangrove (water and dissolved salts enter the root) as described in the literature. Varying amounts of salts remain at the roots to maintain a negative water potential gradient (pore water flow from soil into roots). The root’s water pressure potential reaches a minimum around noon and maximum around midnight. Evapotranspiration reaches a maximum around noon and minimum around midnight. Salt excretion rates at the leaves is highest overnight. Original image from (Alongi and Brinkman, 2011).
Figure 2.2: Digital elevation map of Imp-24: a section of a predominant black mangrove forest on a silicilastic barrier island off the Atlantic coast of Florida. The impoundments are earthen dikes, which prevent surface water exchange between impoundments to the north and south and the brackish, tidal intracoastal lagoon to the west.
Figure 2.3: Extent of resistivity image in reference to surficial geology. Cross section modified from Stringer et al. (2010).

Figure 2.4: Schematic illustration of survey design. Fifteen complete datasets were collected using 1 m electrode spacing and a combined dipole–dipole—inverse Schlumberger array with the AGI SuperSting. CTD sensors in drive points collected conductivity, temperature, and water depth every 15 minutes. Topographic relief is very small, but the salt pan does sit 20-30 cm lower than surrounding vegetation with the exception of a mangrove scrub at mid-transect.
Figure 2.5: Groundwater specific conductivity and water level collected every 15 minutes with Decagon CTD sensors. PVC drive points were installed with screened intervals at 60 cm, 80 cm, and 120 cm though the sensor at 80 cm malfunctioned during the survey.
Figure 2.6: *left to right; top to bottom* The percent change in apparent resistivity from a given time step to the next. A positive change (red) reflects an increase in resistivity; a negative change (blue) reflects a decrease.
Figure 2.7: *top to bottom; left to right* Percent change in resistivity since the previous time step for a given depth as a function of time step. Horizontal black lines indicate sunset and sunrise. Vertical lines indicate the boundary between vegetation zones.
Figure 2.8: Suggested diel salt cycle for a black mangrove contains a period of water uptake that slows down at sunset after which little change occurs for two hours. Water pressure potential in the roots increases enough for water and salts to flow back into the soil (referred to as hydraulic redistribution.) Hydraulic redistribution starts before midnight and continues until sunrise. Original image from (Alongi and Brinkman, 2011).
3. Forward Models to Guide Interpretations of EMI Data in Highly Conductive Environments (Wetlands)

3.1 Abstract

There is a non-linear relationship between an electromagnetic induction (EMI) instrument response and terrain conductivity in the presence of brackish or saltwater. A series of geo-electric models are evaluated to assess the utility of the EMI method for mapping lateral variations in wetland hydrostratigraphy and in particular to understand the signatures of lateral transitions in very high conductivity terrain. While the response crossing over the edge of a clay lens versus a saltwater lens is different in amplitude, the normalized profiles also show the transition is much sharper in a hypersaline regime. There is enough variability in hypersaline regimes to justify collecting profile measurements in inline, transverse, vertical dipole, and horizontal dipole coil configurations to constrain the nature of a lateral boundary. Also, under certain circumstances, there are kinks in the EMI response even across abrupt boundaries due to concentrated current density at a layer’s edge.

3.2 Introduction

Electromagnetic induction (EMI) is a useful reconnaissance method for mapping a wide variety of subsurface features (e.g. Everett and Meju, 2005; Nobes, 1996; Tezkan, 1999; Pellerin, 2002). In the geosciences, EMI is employed to map groundwater extent and salinity (e.g. Siemon et al., 2009; McNeill, 1990; Greenwood et al., 2004; Sherlock and McDonnell, 2003; Viezzoli et al., 2010; Rubin and Hubbard, 2006; Gunnink and Siemon, 2015; MacNeil et al., 2007; Siemon, 2009; Buchanan and Triantafilis, 2009) and clay content (e.g. Williams and Hoey, 1987; Weller et al., 2007; Saey et al., 2008; Triantafilis and Lesch, 2005). Soil salinity, which is of great concern to ecologists and the agricultural industry, also can be
mapped with the fast acquisition of ground conductivity surveys (e.g. Doolittle and Brevik, 2014; Cambardella et al., 1994; Zhu et al., 2010; Yao and Yang, 2010; Triantafilis et al., 2009; André et al., 2012; Robinson et al., 2012; Allred et al., 2008; Herrero et al., 2011; Misra and Padhi, 2014; Thiesson et al., 2011). EMI additionally has its uses in archaeological settings (e.g. Scollar et al., 1990; Simon et al., 2015; Saey et al., 2014; Simpson et al., 2009) and contamination studies (e.g. Nearing et al., 2013; Martinelli and Duplaá, 2008; Martinelli et al., 2012).

There is an interest in coastal wetland environments in resolving conductivity structure that present themselves as layers and lateral transitions—namely clay layering within sand deposits and saltwater saturated zones in freshwater saturated background or vice versa. Direct ground measurements are seldom sufficient to capture of the lateral and vertical heterogeneity of pore-water salinity or clay content. Obtaining a dense enough data sets via direct measurements to quantify this heterogeneity is labor-, time-, and resource-intensive.

EMI can alleviate this mapping effort. In coastal environments EMI data serve as good discriminators between conductivity contrasts (i.e. saltwater versus freshwater, clay versus sand) and are particularly valuable as the instrument does not require contact with ground (e.g. Stewart, 1982; Ridd and Sam, 1996; Sam and Ridd, 1998; Mansoor et al., 2006; Greenwood et al., 2004). Tools using the EMI methods are easy to employ, acquire data quickly, and very common in the environmental sciences for mapping soil and groundwater conductivity.

In hypersaline conditions, like that potentially found in coastal wetlands, EMI can present challenges. Most of the readouts of EMI commercial instrumentation and data interpretation are designed to work in environments that satisfy conditions referred to as the low induction number (LIN). The LIN approximation is valid when the transmitter-receiver coil spacing is much smaller than the depth at which the attenuated transmitted signal is approximately $1/e$ of its amplitude at the transmitter. For a given instrument with a fixed coil spacing, the LIN is primarily a function of operating frequency and bulk
ground electrical conductivity. Generally, zones with conductivities higher than 80 mS/m will violate the LIN assumption. For a setting with clay layering within sand deposits the LIN is valid. However, for zones of saltwater saturated sand in a freshwater background where saltwater conductivity is on the order of 5000 mS/m, the LIN is clearly not valid.

A forward solution valid when the LIN breaks down has been done for a 1D layered Earth (McNeill, 1980; Ward and Hohmann, 1988; Callagery et al., 2007). However, since the world is 3D, many EMI mapping surveys are appropriately conducted as a series of 2D profiles and 2D or 3D forward models are desirable. (2D is defined as a cross-sectional space of the subsurface and 3D as a volume of the subsurface.) There is a need for solutions of 2D and 3D problems that can serve as a guide for data interpretations, particularly in settings that violate the LIN approximation. The aim in this approach is to gain understanding while avoiding the high computational cost of the 3D inverse problem and the deficiencies of lower-dimensional inversion in a 3D world. It is worth noting the existence of 2.5D EMI models, which are characterized by the use of a 3D EM source on a 2D subsurface.

The shape of response gradients across lateral boundaries, including differences between coil configurations, could help constrain the nature of lateral boundaries and improve our understanding of lateral transition in very high-conductivity terrain. The goal in producing the synthetic datasets presented in this chapter is to focus on forward modeling the EMI response of wetland hydrostratigraphy so that one might recognize these structures where they arise.

3.3 Methods

3.3.1 The primary field tool

The field tool considered here is the popular EM31 by Geonics—a two-coil coplanar system that can be operated in vertical or horizontal dipole mode (as described by McNeill (1980)). During profile surveys, the boom holding the coils may be held parallel to the direction of the profile or perpendicular to it, referred to inline and transverse profiles, respectively (Figure 3.9).
Given the EM31 operating frequency of 9.8 kHz ($\omega = 6.16e4 \text{ rad/s}$), the responses of a real conductivity model and complex conductivity model are indistinguishable (Figure 3.9). For instruments with even slightly higher operating frequencies the imaginary component of complex conductivity, $i\omega\varepsilon\varepsilon_0$, should be considered.

Sensitivity of the EM31 to ground structures depends on the height of the instrument, dipole mode, and terrain conductivity. Generally speaking, the instrument’s depth of sensitivity is 1-10 m below the coils in vertical dipole mode with maximum sensitivity at 2-3 m; 0-5 m in horizontal mode with maximum sensitivity immediately below the coils. In highly conductive environments, the transmitted signal is strongly attenuated in the uppermost 5 meters in any mode.

The instrument reports an apparent conductivity ($\sigma_a \text{ mS/m}$), which is given by the formula

$$\sigma_a = \frac{4}{\omega\mu_0s^2} \left( \frac{H_s}{H_{p \text{ imag}}} \right)$$

where $H_s$ is the secondary magnetic field and $H_p$ is the primary (source) magnetic field. Apparent conductivity is the conductivity of the uniform halfspace for which the response matches the recorded measurement, under the LIN assumption.

To simulate the EM31 response to a given Earth conductivity structure, it is necessary to compute the $\frac{H_s}{H_p}$ ratio at the receiver coil and use equation 3.1 to find the instrument read out.

### 3.3.2 Forward Solver

The forward solver is a three-dimensional finite volume EMI solver for fully heterogeneous media which produces a staggered Yee (1966) grid of magnetic vector and electric scalar potentials on a Cartesian grid. Where $\nabla \cdot \mathbf{A} = i\omega\mu_0\hat{\sigma}\phi = k^2\phi$, the system of equations are
\[
\begin{bmatrix}
-\nabla^2 + k^2 & \nabla k^2 - k^2 \nabla \\
\nabla \cdot k^2 - k^2 \nabla \cdot & -\nabla \cdot k^2 \nabla + k^4
\end{bmatrix}
\begin{bmatrix}
A \\
\phi
\end{bmatrix}
= \begin{bmatrix}
\mu_0 J_s \\
\mu_0 \nabla \cdot J_s
\end{bmatrix}
\tag{3.2}
\]

where \( \omega \) is the angular frequency \((2\pi f)\), \( \mu_0 \) is the permeability of free space, \( \tilde{\sigma} \) is the complex conductivity, \( \phi \) is the electric scalar potential, \( A \) is the magnetic vector potential, and \( J_s \) is the current source density. The quasi-minimal residual method is used to solve the system of equations Freund et al. (1992). The total electric field \((E)\) is computed by \( \nabla \phi \), and the total magnetic field \((B)\) by \( \nabla^* A \). Refer to Weiss (2013) for more details.

### 3.3.3 Defining the mesh

Typical conductivities in a coastal wetland like a mangrove forest are controlled by the presence of clay layers (3-30 mS/m) and/or saltwater-saturate sand (defined here as 1380 mS/m) in a freshwater–sand, relatively resistive background (~1 mS/m) Palacky (1988). To represent fully saturated sand, the bulk conductivity was calculated with Archie’s Law Archie (1942). Stringer et al. (2010) determined the formation factor (a coefficient controlled by grain shape and porosity) for the underlying sand deposits in a mangrove forest in Florida to be \( F = 3.6 \). Applying Archie’s Law and the formation factor, the conductivity of seawater (5 S/m) in the pore space of sand (0.001 S/m) is 1.38 S/m. The skin depth, which is defined as the distance to which the signal has attenuated to \( 1/e \) its original amplitude (Kaufman and Keller, 1983), is a maximum of 3.5 meters. Thusly, the mesh is 50 m * 50 m * 100 m to 1) avoid scattering from the grid boundaries and 2) incorporate an air-Earth interface and air halfspace.

Although the actual EM31 coils are rectangular, the modeled transmitter coil was a square set by assigning current along the four edges of a cell face such that the source has a dipole moment of \( \bar{1}A - m^2 \). This is equal to that of the EM31 and avoids excessive mesh discretization. The receiver antenna is 3.66 m from the transmitter and coplanar with the transmitter antenna. Discrete representation of the receiver antenna is not necessary. Instead, instrument readings are computed by a simple interpolation of the component of
the magnetic field, B via \( \text{curl}(A) \), in the direction perpendicular to the plane containing the transmitter/receiver coils (Figure 3.1).

Since the focus is on the response of shallow conductivity structures, the mesh size (node spacing) must capture the geologic detail desired. To represent thin horizontal layers in the uppermost 5 meters of the numerical model, vertical node spacing from the air-Earth interface to a depth of 5 meters is 0.5 m; 1.0 m node spacing is used elsewhere. There are 101 nodes in the \( x \) and \( y \) direction and 106 in the \( z \) direction.

Each transmitter location requires solving the previously mentioned system of equations once. This costs us up to about 40 minutes each time for a 101*101*106 node mesh, which itself represents a linear system (equation 3.2 of size 2e6 with \(~2\) million unknowns representing the scalar electric potential and three components of magnetic vector potential).

A profile over a prospective model thus requires repeated forward solutions, which were done cell-by-cell for 70 transmitter locations. The computation can be parallellized so that each solve is simultaneous with the others. This means an entire profile can be computed on a cluster at no additional wall-clock time over that for a single station location.

### 3.3.4 Geo-electric models

The instrument response curve as a function of depth for vertical and horizontal dipole modes are well known for LIN conditions (Callagery et al., 2007). Here how much sensitivity is lost in scenarios of interest where LIN is not valid, and the nature of the response to lateral boundaries, are assessed.

Three groups of 2D models are examined that describe simplified scenarios that might be expected in coastal wetlands.

For the LIN case, the model response is sensitive to limited to material in the top 10 meters, meaning there is no significant change in the ‘average’ apparent conductivity between an interface at 9.5 meters and 10 meters depth. For the non-LIN case, the presence of a lower conductivity beneath a higher one is only detectable to \(~3\) meters (Figure 3.9).
Model group A considers a 2.5 m thick clay lens. The objective here is to model the change in the EM31 response as the instrument crosses the vertical boundary defined by the presence or lack of clay in sand. In model A1, the lens extends 2-4.5 meters in depth and abruptly ends in the center of the mesh (Figure 3.2). Model A2 is the same lens is shifted down to 4-6.5 meters depths (Figure 3.4). Model A3 returns the lens to 2-4.5 meters depth and replaces the sharp lens boundary with zones of progressively lower conductivity in the direction of its edge. In all A models the background is 1 mS/m to represent freshwater-saturated sand. Note that all group A models satisfy the LIN assumption.

Model group B replaces the clay lens with a saltwater-saturated sand lens. The emphasis here is on the ability of an EM31 to resolve the high ground conductivity due to the presence of saltwater and adequately capture a vertical boundary. In model B1 the lens abruptly ends in the center of the mesh (Figure 3.5); model B2 replaces the sharp lens boundary with zones of progressively lower conductivity. Model B3 contains a lens in which saltwater mixes with freshwater so that conductivities are progressively lower towards its edge. Like A models, the background is 1 mS/m to represent freshwater-saturated sand. Although saltwater over freshwater is physically unstable, it can occur during times of pumping (i.e. lateral saltwater intrusion at a pumping well or blocking a saltwater intrusion at an injection well). In a dynamic system like a mangrove forest, a zone of shallow saline pore water can also develop from mangrove roots returning saline water to the ground and rain events rinsing salt off vegetation. The high conductivities in this model group violate the LIN assumption.

Model C considers a clay lens in a saltwater-saturated sand. Here the lens has a lower conductivity than the surroundings. The objective is to quantify the EM31 response in a highly non-LIN environment and determine the form of the instrument response to the lens edge.

Refer to the appendix for benchmark calculations that influenced the choices in conductivity, layer depths, and start and end positions of 2D profiles.
3.3.5 1D Inversion

Given these synthetic readings, how well a traditional 1D inversion scheme recovers the edge of the clay lens was considered (e.g. Hohmann, 1975; Constable and Weiss, 2006). A commonly used, commercial 1D inversion algorithm is IX1D (Interprex, Ltd.), which uses a fast Hankel transform as described in Anderson (1989) in its forward calculations and a ridge regression inversion procedure similar to that described by Inman (1975). The solver does not make a low induction number assumption other than to calculate the apparent conductivity of the solution for display purposes. Sets of soundings are split between inline and transverse style and inverted as independent profiles. Each 1D sounding uses the same starting model– a three layer system where the first and third layer have fixed conductivities. The unknowns in the inversion are set to be the first and second layers thicknesses and the second layer’s conductivity.

3.4 Results

3.4.1 Forward Models

In all profiles, the EM response is larger and more positive over the more conductive portion of the model, which is to be expected. As a numerical check the inline and transverse responses for any given EM31 coil configuration are essentially equivalent (1% difference), as expected, at the ends where the local subsurface is laterally continuous. Furthermore, at these ends the 3D forward solutions should be in close agreement with 1D forward calculations. An average of 16% difference (Table 3.1) indicates the 3D solutions are still experiencing slight edge effects and/or discrepancies due to a square loop in the numerical solution and circular loop in the analytical solution.

EM31 coil orientation will from here on in be distinguished by the orientation of the coplanar coils relative to the ground, the orientation of the boom relative to the profile direction, and the height of the instrument above the ground. For example, a profile that used a vertical dipole configuration at hip height (meaning the user holds the instrument
at their hip, which is approximated as 1 m) with the boom transverse to the profile is \( VD_{\text{ground}}^{\text{transverse}} \).

### 3.4.1.1 Model A

The profiles in this model group relate to a 2D survey over a discontinuous clay lens in a sand background. Conductivities in these models are in the LIN range. The base model, A1, of a clay lens at 2 m depth (Figure 3.2) is compared to a deeper clay lens (Figure 3.4) and a lens with a gradational edge (Figure 3.3). As expected, the EM response is higher over the clay lens for all coil configurations. The \( VD \) configuration returns the largest response, which is expected since the skin depth over sand-clay-sand is just over 2 m. This coil configuration also experiences the largest change in response from sand-clay-sand to sand. \( VD_{\text{ground}}^{\text{inline}} \) and \( VD_{\text{ground}}^{\text{transverse}} \) responses drop from about 10 mS/m to 1.5 mS/m. This is expected as the instrument is most sensitive to a depth range coinciding with the top of clay. There is also a large drop in response from \( VD_{\text{hip}}^{\text{inline}} \) and \( VD_{\text{hip}}^{\text{transverse}} \) (-5 mS/m). Interestingly, \( VD_{\text{ground}}^{\text{transverse}} \) exhibits a small kink in response not seen in the other profiles. This will be discussed later. In profiles with \( HD \) orientations the change in response is smaller than that of \( VD \), however the gradients are larger. \( HD_{\text{ground}}^{\text{transverse}} \) produces the largest gradient (-0.6 mS/m²); the response at the conductivity boundary is larger from \( HD_{\text{ground}}^{\text{transverse}} \) than \( VD_{\text{ground}}^{\text{transverse}} \). Furthermore, the transition zone is centered over the conductivity boundary in the center of the model in both \( HD_{\text{ground}}^{\text{transverse}} \) profiles.

Model A2 shifts the clay lens to 4 m depth. The response from the layered portion of the model for all profiles is 2-4 mS/m smaller than model A1 (Figure 3.4). With the response from the non-layered portion remaining the same, the deeper clay lens produces a subdued version of the model A1 results. \( VD \) profiles give the greatest change in response (-3 mS/m) across the conductivity boundary and \( HD_{\text{ground}}^{\text{transverse}} \) profiles show the largest gradient (-0.125 mS/m²) centered around the boundary. Additionally, the \( HD_{\text{ground}}^{\text{transverse}} \) response is never larger than \( VD_{\text{ground}}^{\text{transverse}} \).
As a general rule of thumb for potential-field geophysical methods, a deeper target will produce a longer wavelength signal. However, when the clay layer is shifted 2 meters down a wavelength increase is not observed. When the response of model A2 is scaled and compared with A1, there is only a slight increase in wavelength (Figure 3.8).

The profiles of the response in model A3, again, are similar to those of model A1 (Figure 3.3) except that there is a wider drop in response crossing the conductivity boundary. In the profiles with $VD$ and $HD_{inline}$ orientations, the response decreases 10 meters ahead of the boundary; 5 m with $HD_{transverse}$.

Despite the difference in gradient width, there is still a point 2-3 m ahead of the conductivity boundary where $HD_{ground}$ is greater or equal to $VD_{hip}$.

3.4.1.2 Model B

Model group B replaces the clay lens in model group A with a saltwater- saturated sand lens (Figure 3.5, 3.6, and 3.7). Such zones of saline water can develop in wetlands as mangrove roots release water and salts to the soil, or where saltwater lenses may be introduced by pumping of adjacent seawater into impounded wetlands.

In profiles with $VD$ orientation the edge of the lens is nicely captured by a significant drop in response ($-275$ mS/m for $VD_{ground}$ and $-175$ mS/m for $VD_{hip}$) and a narrower transition zone than that of models with a clay lens. The slope break at the ‘bottom’ of the slope corresponds well to the location of the lens edge.

For profiles with a $HD$ orientation the responses are similar in shape to those for a clay lens in sand, but over a much larger range ($-140$ mS/m for $HD_{ground}$ and $-80$ mS/m for $HD_{ground}$). The transition zone is wider for $HD_{transverse}$ than $HD_{inline}$ profiles. In the $HD_{transverse}$ profiles the signal shows a partial drop ($15$ m) before the contact, before flattening and then a steep drop over the contact. This sharp corner occurs in $VD_{ground}$ as well though it is a relatively small amplitude.
The $HD_{\text{ground}}^{\text{inline}}$ response is larger than $VD_{\text{hip}}^{\text{inline}}$ when the receiver is over the end of the lens. Both $HD_{\text{ground}}^{\text{inline}}$ and $HD_{\text{hip}}^{\text{inline}}$ are larger than $VD_{\text{ground}}^{\text{inline}}$ and $VD_{\text{ground}}^{\text{inline}}$ when the receiver is just past the lens end. Due to the sharp corners that $HD_{\text{transverse}}^{\text{transverse}}$ profiles exhibits, the responses are larger than $VD_{\text{transverse}}^{\text{transverse}}$ on either side of the boundary.

In the same vein as model A group, the response for a saltwater-saturated sand lens is modeled. A set of profiles are generated for a lens with a gradational edge, model B2, and a lens that pinches out over 10 m, model B3.

As observed are between model A1 and A2, the 10 meter wide gradational lens edge of model B2 produces a wider transition zone and gentler response gradients (Figure 3.6). A sharp corner is still observed in $HD_{\text{transverse}}^{\text{transverse}}$ profiles, but it is shifted to the left where the largest vertical conductivity contrast exists.

For profiles over a lens that pinches out rather than a constant thickness with decrease in conductivity, the transition zones are comparable in width to those of the gradational boundary Figure 3.7. Furthermore, $HD_{\text{transverse}}^{\text{transverse}}$ still exhibits a sharp corner over the lens where its thickness is decreasing, but this feature is very subdued.

Overall, subtle differences exist between the curves produced by the gradient boundary and the tapering boundary.

### 3.4.1.3 Model C

For a final run of simulations, a clay lens in a saltwater-saturated sand like one might find in a coastal wetland environment is considered. The response, regardless of the configuration, is an order of magnitude higher than those of the previous models (Figure 3.8).

In this extremely conductive environment, conditions that violate the LIN approximation dominate. Numerous differences to group A and B results are observed due to the conductive ‘background’. First, significantly difference apparent conductivities are measured in each of the forward models over the right side uniform halfspace. Second, even though the conductivity contrast between lens and background in model C is similar to that in model
group B the amplitude of the response to the contrasting lens is smaller (20-50 mS/m) for the resistive lens in the conductive background (50-250 mS/m). Third, the instrument response to the abrupt lens edge is much more complex.

All four profiles with VD orientations and $HD_{transverse}$ experience intermediate corners near the conductivity boundary. The corners in the $HD_{transverse}$ profile occur once to the right of the boundary whereas there are two VD profile exhibit corners that are more centered over the boundary.

3.4.2 Current flow near the lens edge

For some profiles, a change in response is not an indication that the EM31 is passing over a conductivity boundary. Although only one conductivity boundary exists, the EM31 can return local ‘kinks’ within a longer wavelength response. In the field data, these could potentially be interpreted as additional conductivity contrasts that do not actually exist. To better understand the physics of these more complex responses, four transmitter positions along the $HD_{transverse}$ of model B1 are highlighted to illustrate how the current density distribution changes as the EM31 approaches and crosses a conductivity boundary (Figure 3.9). The current is concentrated in the conductive lens and accumulates along the lens edge as the source is incrementally shifted along the profile. Even as overall current amplitude decreases as the source moves past the lens edge. This concentration of current contributes at the edge to an increased response 2 m before the instrument crosses over the conductive lens edge. Though not shown, similar behavior is seen in inline profiles where responses after the lens edge increase because the transmitter is still over or close to the conductive lens (i.e. $VD_{inline}$ profile on model A1 and B3). Additionally, the sharp corners in $HD_{transverse}$ profiles are wider than those observed in models with a lens of constant thickness. This suggests the amount of current concentrating in conductive layers is, at least partially, a function of layer thickness.
3.4.3 1D inversion across the lens edge

Collecting four measurements per profile style is a common practice in the field for the purpose of attaining four different effective depths of exploration and is critical to obtaining any information about variations in conductivity with depth. Inversion. Sets of soundings taken at a common point are typically inverted using 1D inversion. The objective here is to examine the errors arising from 1D inversions of the 2D data synthesized here. It is important to note, however, that 1D inversion is best suited in conjunction with geostatistical techniques (Brosten et al., 2011) and multiple frequency data (e.g. Brosten et al., 2011; Viganotti et al., 2013; Farquharson et al., 2003) though the latter are not a consideration for the single frequency EM31 (McNeill and Bosnar, 1999).

The 1D inversion results for the synthetic EM response of model A1— a discontinuous clay layer in sand— imperfectly resolve the 2D conductivity structures. The top of clay is well resolved at 2 m depth from the left extent of the profile to the clay’s edge at the conductivity boundary (Figure 3.8). The inversion solution can recover the bulk apparent conductivity of the forward model and an approximate vertical position of the clay layer if given parameter constraints (the conductivities of the first and third layer was fixed at 1 mS/m). Not surprisingly it is unable to definitely capture the lateral boundary defined by the clay lens edge because the solver is only fitting data to vertical variability. Instead, the required intermediate layer thins and becomes more similar to the background conductivity. Interestingly, the area of highest apparent conductivity (>30 mS/m) is about 2-5 m deep but 4-10 m away from the clay’s edge. A more interesting artifact of the 1D inversion is that the lens is inferred to shoal as it pinches out near the boundary. This feature is most pronounced for transverse coil orientation. This may be attributed to the aforementioned current density increase as the transmitter approaches the clay edge.

1D inversion results across a much more conductive lens edge— a lens of saltwater in sand— show different artifacts. Given the same inversion constraints and starting models described for model group A, inversion resolves the overall thickness and depth to the lens,
but a wide range of conductivities are assigned to the saltwater lens, which does not extend as far to the right as it should (Figure 3.8). For inline soundings, the inverted lens is close to the appropriate location, but the solution assigns a erroneously high conductivity to the lens end. The transverse solution assigns high conductivity towards the end of the edge, and yields an intermediate layer slightly more resistive and very shallow at the conductivity boundary. The inline inversion solution also fits the data with an unrealistically thickness to a very deep conductive layer at the lens edge. Results in the appendix show the EM31 instrument would be insensitive to this layer boundary at such depths.

3.5 Discussion

3.5.1 Implications for Field Data Interpretation

In the LIN model group A most of the signal drop (60% or more) occurs near the end of the conductive lens, so if the user mistakenly interpreted the boundary as being at the midpoint of the signal change, they would misinterpret that boundary location, and interpret the conductive lens as not extending as far into the sand as it actually does. This is true for both shallow and deep lens models (Figure 3.8).

In this LIN case, the lateral location of a shallow vertical conductivity boundary can be estimated by the lower slope break in vertical dipole responses, the midpoint along gradients in inline profiles with a horizontal dipole, and the maximum along a gradient in transverse profiles with a horizontal dipole.

A relatively deeper clay lens creates some interesting contrasts in terms of the shape of the response crossing over the edge of the clay lens. The wavelength of the edge effect of the deeper lens is longer than that of the shallow lens, as expected (Figure 3.8). However, the difference is not as dramatic as might have been expected for the increase in depth to the top of the lens (2m to 4m).

A shallow clay lens with a gradational edge produces a very similar curve as a deep clay lens in $HD^{\text{transverse}}$ mode. However, other modes show that the wavelength of the edge effect of the deeper lens is slightly shorter than that of the gradational lens with a 6 meter
width transition zone. This indicates that a shallow gradational boundary may produce a longer edge effect wavelength than a deeper sharp boundary.

There are subtle edge effects that appear in the model with a shallow clay lens (2-4.5 m depth) that do not appear in the deeper clay model (4-6.5 m depth) or the gradational clay model, namely in the $V_D^{\text{inline}_{\text{ground}}}$ and $H_D^{\text{transverse}_{\text{ground}}}$ profiles (Figure 3.8). Such complexities may be indicative of a shallow sharp boundary, but, again, might be hard to distinguish in a more realistic environment.

The major difference in response between model A1 profile styles is the transition zone surrounding the conductivity boundary. A horizontal dipole source along a transverse profile appears to best capture the sharp edge of a clay lens via a narrow negative gradient as the instrument crosses from conductive to resistive material. The top of this gradient corresponds with the location of the boundary as well. The same gradient from a horizontal dipole along an inline profile is wider and centered around the actual boundary. Conversely, transition zones generated by vertical dipoles are not only much wider, but are not centered around the boundary. Rather a slope break indicates the approximate location of the clay’s edge.

Figure 3.14 compares similar forward solutions of the $V_D_{\text{ground}}$ orientation where only one parameter is different. The width of the transition zone from a layered to non-layered subsurface is more influenced by a sharp conductivity contrast and change in thickness of an intermediate conductive layer than depth of that layer or a gradational conductivity change. A highly conductive layer (above the LIN threshold) shows a much sharper transition than that of a LIN valid layer. Local inflections on the curves off the ends of the lens are associated with sharp boundaries.

3.6 Conclusions

A series of geo-electric models are evaluated to assess the utility of the EMI method for mapping lateral variations in wetland hydrostratigraphy and in particular to understand the signatures of lateral transitions in very high conductivity terrain.
The focus on forward models allows us to efficiently compare and contrast the expected variabilities in a layered subsurface with variable conductivity contrasts. Although a fully three-dimensional, broadband forward solver was utilized, the models are restricted to 2D structures. The influence of lateral conductivity contrasts normal to the profile direction has still not been considered.

In general,

- The wavelength of a transition in the EM signal across a lateral boundary is not much greater for a 4 meter deep layer than a 2 meter deep layers.

- A highly conductive layer (above the LIN threshold) shows a much sharper transition than that of a LIN valid layer.

- Highly conductive discontinuous layers can produce kinks in an EMI profile across a transition that could potentially be interpreted as multiple boundaries. This is more pronounced in the presence of sharp boundaries rather than a gradational change in properties.

- Differences between a transverse and inline profile could be useful for discriminating the structure of a lateral boundary.

Differences in response curves suggest that 2D inversion methods could be useful for constraining the nature of lateral boundaries.

3.7 References


### 3.8 Figures & Tables

![Simulation schematic for an EM31 in a horizontal dipole mode.](image)

Figure 3.1: Simulation schematic for an EM31 in a horizontal dipole mode. A profile is simulated by running the same model multiple times and shifting the 9800 kHz source each time. The response is a trilinear interpolation at a point 3.66 m away from the source. The magnetic field component used in the results is controlled by the orientation of the source dipole. For a vertical dipole in either an inline or transverse survey, the $z$ component of the magnetic field is interpolated, whereas the $y$ component is interpolated for a horizontal dipole in an inline profile; $x$ component in transverse.
Figure 3.2: EM31 profiles over a clay lens with a sharp edge at $x = 0$ m in a sand surrounding. Each line is a different coil orientation: $V D_{\text{ground}}$, blue; $H D_{\text{ground}}$, orange; $V D_{\text{hip}}$, green; $H D_{\text{hip}}$, red.
Figure 3.3: EM31 profiles over a clay lens with a gradational edge from $x = -3$ m to $x = 3$ m in a sand surrounding. Each line is a different coil orientation: $VD_{ground}$, blue; $HD_{ground}$, orange; $VD_{hip}$, green; $HD_{hip}$, red.
Figure 3.4: EM31 profiles over a deeper clay lens with a sharp edge at $x = 0$ m in a sand surrounding. Each line is a different coil orientation: $VD_{ground}$, blue; $HD_{ground}$, orange; $VD_{hip}$, green; $HD_{hip}$, red. Vertical scale is the same as that in figure 3.2 to comparison purposes.
Figure 3.5: EM31 profiles over a saltwater lens with a sharp edge at $x = 0$ m in a sand. Each line is a different coil orientation: $V_D_{ground}$, blue; $H_D_{ground}$, orange; $V_D_{hip}$, green; $H_D_{hip}$, red.
Figure 3.6: EM31 profiles over a saltwater lens with a gradational edge at x = 5 m in a sand background. Each line is a different coil orientation: $V_D\text{ground}$, blue; $H_D\text{ground}$, orange; $V_D\text{hip}$, green; $H_D\text{hip}$, red. Dashed lines correspond to the same coil geometries, but are the results from a saltwater lens with a sharp edge.
Figure 3.7: EM31 profiles over a saltwater lens with an edge that pinches out at x = 5 m in a sand background. Each line is a different coil orientation: $VD_{\text{ground}}$, blue; $HD_{\text{ground}}$, orange; $VD_{\text{hip}}$, green; $HD_{\text{hip}}$, red. Dashed lines correspond to the same coil geometries, but are results of a saltwater lens with a sharp edge.
Figure 3.8: EM31 profiles over a clay lens with a sharp edge at x = 0 m in a saltwater saturated sand. Each line is a different coil orientation: $VD_{ground}$, blue; $HD_{ground}$, orange; $VD_{hip}$, green; $HD_{hip}$, red.
Figure 3.9: Selected current density plots of saltwater-saturated sand lens models at four positions along the $H D^{\text{transverse}}_{\text{ground}}$ transverse profile. Plots are a series of curves tangential to every point in the current density vector field. Curves are colors by vector magnitude at a given point. Only the real component of the current density is considered since it is the real part of the electric field that is in-phase with the imaginary part of the magnetic field. Inset image shows a schematic of the coil configuration considered for this suite of models. The current is concentrated in the conductive lens (gray zone) and remains concentrated at the lens edge, but diminishes in amplitude as the source moves in the profile direction.
Figure 3.10: 1D inversion solution for synthetic EM profiles of sand with a discontinuous clay layer. The black box delineates the position of the modeled clay layer and a black line on the color scale indicates its true conductivity. The inversion algorithm can recover the bulk conductivity of the forward model and a somewhat coarse distribution of a clay layer if given parameter constraints. The starting model is the same for each sounding—a three layer system where the first and third layer have fixed conductivities. The unknowns are first and second layers thicknesses and the second layer’s conductivity. It is unable to capture the abrupt boundary defined by the clay lens edge, but does resolve the top of the clay layer up to its edge. The inversion also yields an area of particularly high conductivity (dark red) within the second layer and 4-10 m before the clay edge that is not in the synthetic model.
Figure 3.11: 1D inversion solution for synthetic EM profiles of sand with a discontinuous saltwater layer. The black box delineates the position of the lens and the black line indicates its true conductivity. The inversion algorithm can recover the bulk conductivity of the forward model and a coarse distribution of a hypersaline layer if given parameter constraints. The starting model is the same for each sounding—a three layer system where the first and third layer have fixed conductivities. The unknowns are first and second layers thicknesses and the second layer’s conductivity. However, it is unable to capture the lateral boundary defined by the lens edge. Note how the most conduction portion in the inverted profile is just before the lens edge. This is similar to the clay lens inversion results and further reinforces the idea that a discontinuous conductive layer will cause current to concentration at its edge and yield a higher instrument reading at the conductivity boundary.
Figure 3.12: EM31 profiles over a deeper clay lens with a sharp edge at x = 0 m in a sand surrounding. Each line is a different coil orientation: \( VD_{\text{ground}} \), blue; \( HD_{\text{ground}} \), orange; \( VD_{\text{hip}} \), green; \( HD_{\text{hip}} \), red. Response scaled by a factor of two and shifted down to compare wavelengths. Dashed lines correspond to the same coil geometries, but are the not scaled results from a clay lens at 2-4.5 meters depth.
Figure 3.13: EM31 profiles over a deeper clay lens with a sharp edge at $x = 0$ m in a sand surrounding. Each line is a different coil orientation: $VD_{ground}$, blue; $HD_{ground}$, orange; $VD_{hip}$, green; $HD_{hip}$, red. Response scaled by a factor of two and shifted down to compare wavelengths. Dashed lines correspond to the same coil geometries, but are the not scaled results from a shallow gradational lens at 2-4.5 meters depth.
Figure 3.14: All $V_D^{ground}$ profiles normalized to their respective range of values. Solids lines are inline profiles; dashed lines are transverse profiles. All models except for a clay lens in a saltwater-saturated sand background have a higher bulk conductivity on the left hand half of the profile than the right.
Table 3.1: Comparison of 1D forward calculations by IX1D and 3D solution (S/m) at profile endpoints.

<table>
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<td>3D</td>
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<tr>
<td>HD ground</td>
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<td>0.0054</td>
</tr>
<tr>
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<td>0.0079</td>
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<tr>
<td>HD 1 m</td>
<td>0.0040</td>
<td>0.0036</td>
</tr>
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</table>

3.9 Appendix

Figure 3.15: Schematics of various transmitter-receiver configurations for EM31 used in field surveys. Transverse with instrument on the ground (top left), Transverse with instrument at ‘hip’ height (top right), Inline with instrument on the ground (bottom left), and Inline with instrument at ‘hip’ height (bottom right). Not shown are vertical dipole versus horizontal dipole orientations, which is simply a matter of rotating the instrument along its long axis.
Figure 3.16: $V_{D_{ground}}$ responses for a complex conductivity model as a function of angular frequency. Open circles are the $V_{D_{ground}}$ response for a real conductivity model at 9.8 kHz. For the purposes of simulating an EM31, which operates at 9.8 kHz ($\omega = 6.16e4rad/s$), the imaginary component of the complex conductivity, $i\omega\epsilon_r\epsilon_0$, does not to be considered.
Figure 3.17: Plot of modeled apparent conductivity for VD mode (open circles) and HD mode (dots) for EM31 in two layered systems using representative conductivity values. Plots illustrate the depth of resolution of conductivity contrasts. For the non-LIN case (left), the presence of a lower conductivity beneath a higher conductivity is only detectable to 3 m depth. For the LIN case (right), the same relatively resistive layer is detectable up to 10 m depth. Although a saltwater over freshwater (bottom left) is physically unstable, it can occur during times of pumping.
Figure 3.18: Results from a series of models where the EM source (EM31 transmitter) was positioned along x-directed profile. Values plotted are the EM response 3.66 m from the transmitter in the +x direction and represent that measured by an EM31. Note the wide zone where edge effects are present. Results reports here are profiles that do not extend beyond 35 m from the center of the model.
Figure 3.19: The full analytical solution for $\mathbf{B}$ as a function of conductivity (squares) compared with the LIN approximation (triangle) and APhiD solution (circle) ($f = 9800$ Hz, coil separation = 3.66 m). The linear relationship between terrain conductivity and the total magnetic field breaks down around 100 mS/m. The numerical results from APhiD mostly agree with the full analytical solution. The slight discrepancy is likely due to the near-field energy from a square transmitter loop defined in the numerical model.
4. Investigation of the Hydrogeologic Controls on Selected Wetlands in Covered Karst with 2D Resistivity and Ground Penetrating Radar

4.1 Abstract

Geophysical surveys were conducted at selected xeric wetlands to image the underlying lithology in west-central Florida. ER was used to image the geometry to depths including the top of limestone; ground penetrating radar (GPR) was used to image stratigraphy beneath and surrounding wetlands. The wetlands can be grouped into two models. Topographic highs surrounding wetlands are either controlled by limestone pinnacles or thick sand packages. Areas with limestone pinnacles defining uplands are observed in the Brooksville Ridge physiographic province, which is characterized by limestone ridges. Areas with thick sands defining uplands are observed in the Gulf Coastal Lowlands, which is characterized by sand dunes deposited over scoured limestone.

4.2 Introduction

West-central Florida is characterized by a mantled karst terrain composed of Cenozoic sediments and sedimentary rocks and wetlands are very common throughout the region. Fine sands and clays overlie highly-weathered limestone with a regionally-discontinuous clay unit that is sometimes present between them. The top of limestone is extremely irregular and can outcrop at locations with little to no overburden.

This overall description of the geology is a good summary of regional trends, but may or may not account for local variations beneath and surrounding wetlands, in particular, a subset of wetlands in xeric environments surrounded by uplands. The underlying lithology, and thus how it relates to hydrogeology, is not well understood.
In an effort to better contrain the underlying geology beneath these wetlands, a series of ground penetrating radar (GPR) and electric resistivity (ER) surveys were conducted at five wetland sites (Figure 4.1). Geophysical data and borehole logs were used to identify the geometry of strata and locate the top of weathered limestone beneath the wetlands. Interpretations were also used to compare and contrast these wetlands to cypress dome wetlands also found in the region as described by Watson et al. (1990); Tihansky (1999); Lee et al. (2009).

4.3 Study Sites

Mid-Pliocene and Holocene undifferentiated sands and clays define the region’s surficial aquifer (Miller, 1997). These are fine to medium quartz sands (some spodic) containing lenses of silty and sandy clay (Scott et al., 1988; Healy, 1982). The surficial aquifer ranges from 3 to 30 meters thick. The water table is typically relatively close to the surface (0-10 m). Miocene to mid-Pliocene interbedded fine sands, clays, and carbonates make up the Intermediate Confining Unit, which is identified as the Intermediate Aquifer where the clay content is low. Thickness is extremely variable, but generally thins to the northeast. In the study area, relatively dense clays are observed the the intermediate unit. Where these clay layers are very thin, or at breaches due to sinkhole activity, vertical flow may occur between the surficial and Floridan aquifers. To the south and west, where the intermediate unit is thickest, more sands and carbonates increase the overall permeability, however there is still limited vertical flow to or from the surficial or Upper Floridan aquifer (Miller, 1997; Scott et al., 1988).

The Upper Floridan aquifer is defined by a thick sequence of carbonates which dips to the southwest where overlying intermediate units thicken. Dissolution along pre-existing fractures and karst features are sites of large, localized permeabilities (Miller, 1997). As stated above, in areas with little vertical flow through the intermediate units, the Upper Floridan is confined. Otherwise, the surficial and Upper Floridan aquifers are hydraulically connected.
4.4 Methods

Ground penetrating radar (GPR) and resistivity (ER) data were collected at a total of five wetlands in west-central Florida (Figure 4.1). Each method offers distinct advantages for identifying subsurface lithology. Because the depth of ER penetration is much greater than that of GPR (at the frequencies used in this study) it is better suited to identify limestone surfaces and karst features that lie, in these sites, between 0 m and 30 m below ground surface. GPR is better suited to resolve shallower features, such as clay layer(s) and the water table, within the uppermost 10 m Singh (2006); Tronicke et al. (1999). Used together with borehole logs and water level data these data can provide important clues to lithology underlying the wetlands.

GPR surveys were designed, in particular, to image stratigraphy on the banks of depressions. Data were collected with MALÅ250 MHz or 500 MHz shielded antennae and post-processed with the Reflex-W software package (Sandmeier). Profiles cut across the long and short axes of the wetlands, and in some cases, circumnavigate the depression.

ER surveys were designed to maximize the depth of signal penetration in an effort to image bedrock. Data were collected with AGI Super Sting R8 resistivity meter with 56 electrode spread using a combination of a dipole-dipole and inverse Schlumberger array. This array was chosen for its balance between high sensitivity to lateral variations, moderate sensitivity to horizontal structures, and acquisition time Loke (1999). ER transects were located as close as possible to the long and short axes of wetland without inundating the electrodes in surface water. For wetlands with surface water, the transects were shifted upslope and tangential to the wetlands edge. Data were post-processed and inverted with the RES2DINV software package from Geotomo, Inc.

Borehole logs, provided by the Southwest Water Management District, were used to characterize the substrate within and/or near the wetlands and were especially helpful in distinguishing different sediment types. Bedrock is seldom recorded in borehole data so as
to avoid puncturing the confining clay unit, though limestone is presumed to exist beneath the clay.

4.5 Results and Interpretations

All wetlands from this subset have a long and short axis and are herein referred to as a pool if they were inundated at the time of the surveys, or as a depression if not.

4.5.1 Chassahowitzka

The 4.9-acre oval-shaped pool at Chassahowitzka is elongated in the NE-SW direction and situated at the bottom of a relatively steep basin (Figure 4.2). At a deep coring site 2.5 km to the north the Upper Floridan is unconfined.

In both ER profiles, there is a laterally continuous high resistivity layer that corresponds to sand/silty sand seen in boreholes. (Figures 4.3 and 4.4). NW slopes are slightly steeper than other slopes and show a zone of locally highest resistivity, interpreted as dry sands that thicken upslope. There is a steep decrease in resistivity at depths coinciding with lithologic change (sand/silty sands to sandy, silty clay). Correlations between borehole logs and ER support the interpretation of the lower resistivity zone as clay underling sand. The variable thickness in both profiles suggests a hummocky contact between sand and clay. Considering the resistivity’s depth of resolution much of the low resistive material is inferred to be saturated limestone, though there is no local lithological data to corroborate.

The transition to more clay-rich sediment is seen as a strong GPR reflector with significantly more noise beneath it. Although topography surrounding the pool is subtle, a strong horizon parallels the ground surface; with the sand package on profile G1 (Figure 4.5). Two notable gaps at ~100 and ~450 m are positioned beneath ‘sags’ in overlying strata. These zones, each 20 m wide, are interpreted as zones of subsidence, where soil has filled in above the sagging layer, and the underlying 2 m deep layer has been breached. These areas of subsidence have been covered with younger sediment. Overall the GPR signal below -10 m s.l. is noisy and may indicate complex returns from weathered limestone and noise
from below clay. The depth of this transition in the character of the GPR signal generally corresponds with the depth of the high to low resistivity transition. Thus both the GPR and ER indicate an undulating contact between silty sand and clay or limestone.

One transect, G3, run radially out from the pool and continues upslope (Figure 4.6). Using the interpretation from G1, it appears limestone is at its lowest elevation slightly offset from the topographic high. The topography is a function of sand thickness instead. Spodic sands recorded in the upland borehole (B1) are not found at the wetland’s edge (B2) (Figure 4.5). This, along with the GPR data, suggests a lateral change in material between the upland and the pool. Plastic clay recorded in B2 sits at the same elevation as silty clay in B1, indicating the top of limestone is at a potentially higher elevation beneath the wetland than it is beneath the topographic highs.

The remaining GPR transects image the uplands surrounding the pool. The highest degree of undulation in limestone/clay, inferred from both GPR and ER, occurs beneath topographic highs or the slopes adjacent to the highs. Sand is thickest at topographic highs and in places pinches out, including at the wetland edge. Based on correlations between borehole data and the GPR transects that intersect them, three layers of sediment are inferred to overlie undulating limestone-sand, spodic sand, and sandy, silty clay. Only the borehole along the wetland edge showed dense, plastic clay.

In summary, at Chassahowitzka, the spodic sand and underlying clay appear to blanket the undulating limestone while the sand partly controls topography via variable thicknesses. Note, profile G5 has very little relief and all three layers of sediment have a consistent thickness over a relatively flat top of limestone.

4.5.2 Starkey E

The Starkey E site is also a oval-shaped wetland and is surrounded by uplands (Figure 4.7). The strongest GPR reflection is the contact between sand or spodic sand and clay (Figure 4.8). Sand appears as lens that define local topographic highs. The underlying clay may or may not pinch out at the edges of local depressions. These factors are observed
across all three GPR transects, two of which have corroborating borehole data. Sandy clay often underlies clay. Clay is seen in the GPR as a discontinuous reflection. Where clay is discontinuous the clay reflector above and deeper reflections below are also discontinuous or less prominent.

Gravelly clay is recorded in one borehole (B4). This is interpreted as the top of highly weathered limestone since clayey weathered limestone with gravel sized clasts are recorded in borehole at the nearby Starkey A site (Figure 4.7). The gravelly clay and the underlying limestone appear to undulate, but appear to have little control on topography. Rather, limestone is closest to the surface at topographic lows and depressed beneath topographic highs (Figures 4.9 and 4.10).

4.5.3 Starkey A

The Starkey A site includes two round pools situated between moderately low and high relief to the north and south, respectively (Figure 4.7). There are only ER and borehole data from this site.

Geophysical data are consistent with borehole data (Figure 4.11) in that a highly resistive surface layer agrees in thickness with sand observed at two adjacent boreholes. This resistive layer ranges from 0 m to 6 m and covers most of the surface on the profile in Figure 4.11. The inferred laterally discontinuous sands are most resistive (interpreted as driest) on the steeper slopes. On the NW end, the sand pinches out toward the wetlands. The sharp negative downward gradient in ER corresponds to a lithological change from sand to silty or sandy clay. Where the sand is discontinuous the clay layer crops out at the surface, which, at this location, is also a topographic high.

Starkey A is the only site where limestone is recorded in borehole data. Beneath the sandy clay is 2-3 meters of highly weathered limestone (clayey and with gravel size clasts) over intact limestone. These depths correspond well with the large area of low resistivity imaged in E1. Thus variations in the sharp negative downward resistivity gradient are interpreted to be the absence of a clay layer where sands and silts directly overlie weathered limestone.
Similar to Starkey E, the top of limestone is closest to the surface at topographic lows and deepest beneath highs.

### 4.5.4 String of Pearls

String of Pearls is an irregularly-shaped wetland located at the center of a basin (Figure 4.12).

Two separate but overlapping ER transects were collected and later merged together, resulting in a section of subsurface between the two profiles not being imaged (Figure 4.13). As for previous sites, highly resistive material corresponds with sand that appears driest in topographic highs. Underlying clayey sands correlate to a sharp negative downward resistive gradient. The depth of this sand to clayey sand contact is highly variable. The contact is closest to the surface beneath topographic highs and lowest beneath lows, which contrasts findings at the previous sites. It is presumed that limestone underlies clay and thus the top of limestone may also be undulating.

The increase in clay content from sand and silt to clayey sand or to dense clay is imaged as a prominent reflector in the GPR data. Areas in the GPR profiles where no data were returned from beneath this reflector are likely due to electrically conductive dense clay, in which the GPR signal strongly attenuates (Figure 4.14, for example). Limestone is inferred to underlie dense, plastic clay. However, areas where the signal does penetrate are interpreted to be a clay and sand/silt mix or clayey limestone (Figure 4.15). For either case, clay-rich layers are discontinuous and the available borehole data show thick packages of sand to reside where clay-rich layers are absent.

In summary, combining boreholes with GPR and ER, sands transition downward to stiff clayey sands within the depression. At higher elevations surrounding the wetland, sand transitions to dense clay, which is presumed to overlie weathered limestone based. Although there appears to be little connection between the depth to clayey sand and topography, upland elevations appear to be a function of depth to limestone as inferred by the position of dense clay that presumable overlies limestone. Where the dense clay layer is close to the
surface, limestone is interpreted to also be close to the surface. These areas are situated beneath topographic highs.

4.5.5 Croom

This small two-pool wetland is 4 km east of String of Pearls (Figure 4.16). Borehole data are not available, however, water district records show approximately 3 meters of sand over 5 meters of clay over weathered limestone about 1 km to the east (District, 1998).

An ER profile images a thin resistive zone immediately below the surface. It is thickest (4 meters) at higher elevations and almost absent at the south end of the transect (Figure 4.17). In the GPR data, this resistive layer appears as a scattering of high amplitudes with possible internal structures imaged where the zone is thickest (Figure 4.18). This is interpreted to be the sands recorded in the nearby well.

As in the ER data at other sites, a sharp negative downward gradient occurs beneath the resistive zone. The undulation of this gradient is also observed in GPR as an abrupt decrease in signal amplitude. This is interpreted as a clayey sand or sandy clay layer, the bottom of which produces a bright, but discontinuous reflection.

4.6 Discussion

Dry sands thin or completely pinch out approaching wetland pools or depressions. Thicknesses vary laterally and loosely control topography at Chassahowitzka (figures 4.6), Starkey A, and Starkey E (figures 4.8, 4.9, and 4.10). Underlying clay layers are often discontinuous both beneath wetland slopes and the wetlands themselves (i.e. figure 4.9). Gaps in clay, as imaged with GPR in figures 4.5 and 4.15, coincide with possible slumping of overlaying sands. Relatively low resistivity material at depth is confirmed as saturated limestone in one borehole, B8, at Starkey A (Figure 4.11) and inferred elsewhere at depth.

Possible breaches in confining clay layers beneath topographic highs between depressions are seen in resistivity profiles as conductive anomalies and in GPR as interruptions in otherwise continuous horizons.
Two conceptual models (Figure 4.19) are offered for the wetlands of west-central Florida. Topographic highs surrounding wetlands are either controlled by limestone pinnacles or thick sand packages. Areas with limestone pinnacles defining uplands are observed in the Brooksville Ridge physiographic province, which is characterized by limestone ridges. Area with thick sands defining uplands are observed in the Gulf Coastal Lowlands, which is characterized by sand dunes deposited over scoured limestone.

4.7 Conclusions

The wetlands situated here from the current model of cypress dome wetlands in that surrounding topography is not always controlled by the undulating top of limestone at depth. Topographic highs surrounding wetlands are either controlled by limestone pinnacles or thick sand packages. Areas with limestone pinnacles defining uplands are observed in the Brooksville Ridge physiographic province, which is characterized by limestone ridges. Area with thick sands defining uplands are observed in the Gulf Coastal Lowlands, which is characterized by sand dunes deposited over scoured limestone.

4.8 References


Figure 4.1: Map of the selected wetland sites and physiographic provinces
Figure 4.2: Position of ER, GPR, and boreholes at Chassahowitzka overlaying bare Earth LiDAR collected by EarthData Int’l LLC 2003-2011. Blue polygon delineates historic wetland boundary.
Figure 4.3: ER transect, E1, with borehole data and hydrogeologic interpretations. See figure 4.2 for its location.
Figure 4.4: ER transect, E2, with borehole data and hydrogeologic interpretations. See figure 4.2 for its location.

Figure 4.5: GPR transect, G1, with borehole data and hydrogeologic interpretations. See figure 4.2 for its location.
Figure 4.6: GPR transect, G3, with hydrogeologic interpretations. See figure 4.2 for its location.
Figure 4.7: Position of ER, GPR, and boreholes at Starkey overlaying bare Earth LiDAR collected by EarthData Int’l LLC 2003-2011. Blue polygon delineates historic wetland boundaries.
Figure 4.8: GPR transect, G1, with borehole data and hydrogeologic interpretations. See figure 4.7 for its location.

Figure 4.9: GPR transect, G2, with borehole data and hydrogeologic interpretations. See figure 4.7 for its location.
Figure 4.10: GPR transect, G3, with borehole data and hydrogeologic interpretations. See figure 4.7 for its location.

Figure 4.11: ER transect, E1, with borehole data and hydrogeologic interpretations. See figure 4.7 for its location.
Figure 4.12: Position of ER, GPR, and boreholes at String of Pearls overlaying bare Earth LiDAR collected by EarthData Int’l LLC 2003-2011. Blue polygon delineates historic wetland boundaries.
Figure 4.13: ER transects, E1 and E2, merged together with borehole data and hydrogeologic interpretations. See figure 4.12 for its location.

Figure 4.14: GPR transect, G4, with borehole data and hydrogeologic interpretations. See figure 4.12 for its location.
Figure 4.15: GPR transect, G1, with borehole data and hydrogeologic interpretations. See figure 4.12 for its location.
Figure 4.16: Position of ER and GPR at String of Pearls overlaying bare Earth LiDAR collected by EarthData Int’l LLC 2003-2011. Blue polygon delineates historic wetland boundary.
Figure 4.17: GPR transect, E1, with hydrogeologic interpretations. See figure 4.16 for its location.

Figure 4.18: GPR transect, G1, with hydrogeologic interpretations. See figure 4.16 for its location.
Figure 4.19: Conceptual models based on geophysical interpretations as a function of physiographic province.

4.10 Appendix

Figure 4.20: GPR transect, G2, with borehole data and hydrogeologic interpretations. See figure 4.2 for its location.
Figure 4.21: GPR transect, G4, with hydrogeologic interpretations. See figure 4.2 for its location.

Figure 4.22: GPR transect, G5, with hydrogeologic interpretations. See figure 4.2 for its location.
Figure 4.23: GPR transect, G2, with borehole data and hydrogeologic interpretations. See figure 4.12 for its location.
Figure 4.24: GPR transect, G3, with borehole data and hydrogeologic interpretations. See figure 4.12 for its location.

Figure 4.25: GPR transect, G2, with hydrogeologic interpretations. See figure 4.16 for its location.
5. Conclusions

This dissertation illustrates the usefulness of shallow geophysics, particularly, electric and electromagnetic methods, to address research questions from critical zone hydrogeology. The research presented contributes the mangrove tree physiology, field interpretation, and wetland science literature. A time-lapse electrical resistivity (ER) survey was conducted in section of mangrove forest on a barrier island in southeast Florida. The objective was to image changes in pore water salinity due to mangrove tree physiology. ER data show that over a 24 hour period there is more variability in the root zone than elsewhere along the transect. Around the roots, a decrease in resistivity observed two hours after sunset suggests water and salts redistributed to the soil via roots. An increase in resistivity at sunrise suggests roots resume water uptake and pulls fresher water up from below the root zone. In the presence of brackish or saltwater there is a non-linear relationship between an electromagnetic induction (EMI) response and terrain conductivity. A suite of forward models that simulate the response of a commercial EMI instrument in a series of 3D conductivity models with varying lateral boundaries and conductivities typical of a coastal wetland (i.e. clay lens or saltwater lens in sand). While the response crossing over the edge of a clay lens versus a saltwater lens is different in amplitude, the normalized profiles show the transition is much sharper in a hypersaline regime. There is also enough variability in hypersaline regimes to justify collecting profile measurements in inline, transverse, vertical dipole, and horizontal dipole coil configurations to constrain the nature of a lateral boundary. Also under certain circumstances there a kinks in the EMI response even across abrupt boundaries due to concentrated current density at a layer’s edge. In west-central Florida, geophysical surveys were conducted at selected xeric wetlands to image the underlying lithology. ER was used to
image the geometry of the top of limestone; ground penetrating radar (GPR) was used to image stratigraphy beneath and surrounding wetlands. These wetlands can be grouped into two models. Topographic highs surrounding wetlands are either controlled by limestone pinnacles or thick sand packages. Areas with limestone pinnacles defining uplands are observed in the Brooksville Ridge physiographic province, which is characterized by limestone ridges. Area with thick sands defining uplands are observed in the Gulf Coastal Lowlands, which is characterized by sand dunes deposited over scoured limestone.