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Mapping porewater salinity with electromagnetic and electrical methods in shallow coastal environments, Terra Ceia, Florida

Wm Jason Greenwood
University of South Florida

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Mapping Porewater Salinity with Electromagnetic and Electrical Methods in
Shallow Coastal Environments: Terra Ceia, Florida

by

Wm. Jason Greenwood

A thesis submitted in partial fulfillment
of the requirements for the degree of
Masters of Science
Department of Geology
College of Arts and Sciences
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DEDICATION

I dedicate this thesis to my parents:
Dr. William R. Greenwood (1938-1992) and Ellen M. Greenwood
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Mapping Porewater Salinity with Electromagnetic and Electrical Methods in Shallow Coastal Environments: Terra Ceia, Florida

Wm. Jason Greenwood

ABSTRACT

The feasibility of predicting porewater salinity based on calibrated surface electromagnetic methods is discussed in a coastal wetland on the southern banks of Tampa Bay in west-central Florida. This study utilizes a new method to float commercial land based electromagnetic (EM) instruments in shallow marine waters of less than 1.5 meters. The floating EM-31 (Geonics, Ltd.) effectively sensed the magnitude and lateral extent of high and low salinity porewaters within mangrove lined ditches and ponds. Resistivity and EM geophysical methods are merged with direct sampling data to calibrate layers in electromagnetic models to infer shallow (<30m) groundwater salinity patterns. Initial marine resistivity surveys are necessary to discriminate between equivalent EM model solutions for seafloor conductivities beneath shallow (0.1-1.5m) marine (~30 ppt) waters. Using formation factors computed from nearby resistivity surveys, porewater conductivity predictions based on surface EM-31 and EM-34 measurements are successful at distinguishing overall porewater salinity trends.

At the Tampa Bay study site, the most distinctive terrain conductivity anomalies are associated with mangroves bordering marine waters. Highly elevated porewater conductivities are found within 5m of the mangrove trunks, falling sharply off within 10m, presumably due to saltwater exclusion by mangrove roots.
Modeling indicates the shallow water EM-31 measurements probably lack the resolution necessary to image more subtle porewater conductivity variations, such as those expected in association with diffuse submarine groundwater discharge. However, the technique has potential application for locating high contrast zones of freshwater discharge and other salinity anomalies in shallow and nearshore areas not accessible to conventional marine resistivity or land-based arrays, and hence may be useful for interdisciplinary studies of coastal wetland ecosystems.
Introduction

Coastal hydrologists, oceanographers, biologists and land managers all seek an understanding of the patterns of shallow groundwater salinity. Salinity strongly influences the health, productivity and species composition of essentially all coastal life (Morss, 1927; Chapman, 1960; Mitsch and Gosselink, 1993). Knowledge of groundwater salinity patterns improves and gauges the effects of wetland restoration planning, which is complicated by inaccessible terrain that exhibits large lateral and vertical salinity variations over small distances. Increased resolution is afforded when salinity data extends beyond available wells in dual density numerical groundwater flow models (Voss, 1984; SUTRA, Souza, 1987; SEAWAT, Guo and Langevin, 2003).

Closer to shore, anomalous zones of low salinity groundwater have been associated with submarine groundwater discharge (SGD), the upward flux of groundwater across the sediment-water interface (Johannes, 1980; Vanek, 1991; Hoefel and Evans, 2001; Manheim et al., 2001). SGD has significant ecological consequences and may be an important public health risk, as it is a potential source of excess nutrients, pollutants and human pathogens into coastal waters (Johannes, 1980; Capone and Bautista, 1985; Paul et al., 1997).

Effective delineation of salinity patterns in coastal zones, particularly wetlands, often requires numerous wells, which are prohibitively expensive in comparison to widely used geophysical methods that are sensitive to the conductivity contrast between
fresh and saline saturated terrain (terrain conductivity) (Cameron et al., 1981; Barker, 1990; McNeill, 1990; Fitterman and Deszcz-Pan, 1999; Stewart, 1999; Hoefel and Evans, 2001; Manheim et al., 2001). In this study, the feasibility of mapping groundwater salinity is assessed in the Terra Ceia Study Area (TCSA), which encompasses 7.8 km² of primarily tidal marsh interspersed with coastal uplands and freshwater ponds on the southern bank of Tampa Bay, 10 kilometers (km) north of Palmetto and 15 km west of Parish, in Manatee County, Florida (Figures 4 and 5). Investigations were conducted cooperatively between the US Geological Survey Tampa Bay Integrated Science Study, the University of South Florida Geology Department and the State of Florida Department of Environmental Protection.

Groundwater salinity patterns in the TCSA are strongly influenced by topography, precipitation, evaporation, transpiration, mangrove soil salinization, tides, and surface water flow in ditches and ponds. These influences are often highly variable. For example, such as the case where groundwater salinity was found to vary from 2 to 27 parts per thousand (ppt) in the uppermost 15 meters (m) of a 50 m² area of densely vegetated upland and wetland modified by dredge and fill structures including mosquito control ditches and berms (Figure 5, Area 4). Increased coverage may be possible by aerial electromagnetic methods of this relatively inaccessible terrain, but these methods are expensive and may lack resolution necessary to identify small-scale features (Fitterman and Deszcz-Pan, 1999; Fitterman and Deszcz-Pan, 2001; Stewart et al., 2002).

Locating general areas of high and low salinity groundwater is possible based on surface based geophysical methods, however quantifying these areas requires additional knowledge of the factors that influence terrain conductivity. These factors include
porewater conductivity, temperature, conductive clay content, porosity, pore space shape and connection and degree of saturation (Keller and Frishknecht, 1970; McNeill, 1990). Mapping groundwater salinity via geophysical methods requires the following three steps: (1) reconnaissance mapping by geophysical methods to assess horizontal and vertical variability in terrain conductivity, (2) direct sampling of areas of interest to determine the local relationships between terrain conductivity and porewater conductivity and (3) application of the widely recognized standard for the relationship between seawater salinity and porewater conductivity (IES 80 method in Appendix 1).

Results from this study incorporate methods to measure terrain conductivity, relate terrain conductivity to groundwater conductivity based on local direct samples and the adaptation to shallow water (<1.5 m depth) of commercial electromagnetic (EM) and resistivity (DC) systems. Discussion of results and their relevance to other sites includes the strengths and limitations of EM and DC instruments in coastal settings, influences of mangroves on groundwater salinity and the potential imaging of submarine groundwater discharge.

Electromagnetic data is typically expressed in the units of conductivity or Siemen per meter (S/m); direct current resistivity data is typically reported as resistivity or Ohm-meters (Ohm-m). An ohm-meter is the reciprocal of a Siemen per meter. Comparison is facilitated in this text by consistently expressing all EM and DC data in units of conductivity in milli-Siemens per meter (mS/m).

Development of shallow-water geophysical techniques in this study have the potential for imaging submarine groundwater discharge (SGD) which occurs when groundwater flows upward across the sediment-seawater interface into near shore.
environments when an aquifer is hydraulically connected with the sea through permeable bottom sediments and the hydraulic head is above sea level (Johannes, 1980; Hutchinson, 1983). The presence of SGD has been documented in most coastal environments, including bays, coves and coral reefs (Lewis, 1987; Giblin and Gaines, 1990; Vanek, 1991; Simmons, 1992; Simmons et al., 1992; Schneider, 2003). In previous studies, SGD has been shown to contribute up to 20% of all freshwater and ≥20% of the total dissolved nitrogen to Great South Bay, New York (Capone and Bautista, 1985) and 50% of the total dissolved nitrogen input near Perth, Australia (Johannes, 1980) as well as being a potential vehicle for the dispersal of human pathogens to coastal waters, especially in regions with waste water injection wells (Paul et al., 1995; Paul et al., 1997).

A finite difference numerical groundwater model of Tampa Bay estimates SGD as 5% of the total fresh water input (Hutchinson, 1983). This model did not account for dual density water and was run under steady state conditions, so this value may rise as high as 10-20%, during peak months, using current groundwater models (Swarzenski pers. comm.).

Submarine groundwater discharge may be found either in the form of diffusive seeps or more localized springs, both of which have been clearly delineated with marine resistivity and electromagnetic methods (Hoefel and Evans, 2001; Manheim et al., 2001). Locating diffuse SGD from surficial aquifers is more difficult because anomalies are subtle and analytical models, seepage meter data, and tracer studies all indicate that overall flux rates will most likely be greatest close to the shoreline where interference with mangrove soil salinization may occur and shallow depths may limit the use of marine systems (Vanek, 1991; Passioura et al., 1992; Banks et al., 1996; Corbett et al.,
One focus of this thesis is to present adaptations of commercially available land based electromagnetic and resistivity devices to sense shallow porewater conductivity that may lead to improved imaging of spatial patterns of SGD in near shore environments.

**Electromagnetic Methods**

Electromagnetic instruments generate alternating currents in a transmitting coil at the surface, which induce eddy currents in the sub-surface. The ratio of the secondary magnetic field induced by the eddy currents to the primary magnetic field is measured by a receiving coil, and can be related to the terrain conductivity, or bulk electrical conductivity of the material beneath the instrument (McNeill, 1980a).

*Use of EM in Groundwater Studies*

EM methods are widely used in hydrogeologic studies, exploiting the terrain conductivity variations associated with freshwater/saltwater interfaces, highly conductive clay confining units, high conductivity contaminant plumes, and low conductivity aquifer units (McNeill, 1990; Cherkauer et al., 1991; Woldt et al., 1998; Ayotte et al., 1999; Fitterman and Deszcz-Pan, 1999; Fitterman and Deszcz-Pan, 2001; e.g., Bendjoudi et al., 1999; Uchiyama et al., 2000). Wetlands and shallow (<1 m) water depths coincide with a spatial gap between existing land-based and marine EM and DC methods. Recirculated seawater pumped by tides and bioturbation mixes with fresh groundwater to form fresher porewaters near the sediment seawater interface in areas of shallow SGD (Moore, 1999).
2002; Stewart et al., 2002). In coastal environments, these methods have also been used to map freshwater lens morphology and seasonal variation on siliciclastic barrier islands (Stewart, 1990; Anthony, 1992; Caballero, 1993; Ruppel et al., 2000; Schneider and Kruse, 2001).

Use of electromagnetic methods offshore or in lakes has not been extensive. Time domain EM was used in a freshwater lake in order to estimate the depth of a saline body of water underlying the lake bed (Goldman et al., 1995; Goldman et al., 1998). In other studies, a marine EM transmitter-receiver array, with multiple frequency and coil spacing capability, similar to the Geonics, Ltd. EM-34, was towed along the seabed in order to delineate paleo-channels by changes in associated porosity and to locate prospective zones of submarine groundwater discharge (Evans et al., 2000; Hoefel and Evans, 2001). Nadeau et al. (2003) used a streaming digital EM-34 with the receiver and transmitter coils mounted in small non-conductive boats. This system was used to map gravel deposits associated with a municipal well field recharge area beneath a freshwater river. A simple numerical correction for the effect of the river water was feasible in this relatively low-conductivity environment (McNeill, 1980a; Nadeau et al., 2003).

The TCSA site differs from most sites discussed in the literature in that terrain conductivities in the uppermost few meters are an order of magnitude or more higher. Most previous studies show less complex and spatially variable terrain conductivity structures. In addition, the water-born data acquisition and interpretation techniques described in preceding studies were not directly transferable to the TCSA, where water depths in areas of interest are shallow (<1.5 m) and surface water conductivities are very
EM data acquisition methods and interpretation that are feasible in shallow marine environments is the focus of discussion below.

**EM-31 and EM-34 Operation at the TCSA**

The electromagnetic instruments used in this study are the EM-31 and EM-34 of Geonics, Ltd. The EM-34 consists of a pair of transmitter and receiver loop type antennas with corresponding control boxes that are connected by coaxial cables. The EM-34 operates at three frequencies designed to work with transmitter and receiver coil separations of 10, 20 and 40m. The two antenna coils can be placed in either the vertical co-planar orientation (horizontal magnetic dipole - HMD), or in the horizontal co-planar orientation (vertical magnetic dipole - VMD). The HMD mode is significantly more sensitive to near surface materials when compared to the VMD mode (McNeill, 1980a; Kaufman and Keller, 1983; Kaufman and Hoekstra, 2001). Ideally, all three coil separations and two magnetic dipole orientations may be used over the same location for a total of six unique effective exploration depths. Practical limitations in the highly conductive environment at the TCSA are discussed in the results below.

The EM-31 operates at one frequency and has a fixed length boom type antenna with a coil spacing of 3.67 m, so exploration depth is a function of magnetic dipole orientation and instrument height. The EM-31 used in this study had the capability of logging data at timed intervals, allowing the operator to move the instrument in a streaming mode by carrying the instrument at hip height (0.9m) or towing the instrument in a boat (floating 0.1m above the water surface). The boat used to hold the EM-31 in this study was constructed of polyethylene and fitted with wooden supports, plastic
splash shields and a foam and plastic outrigger (Figure 1). EM-31 data were later merged with global positioning satellite fixes by synchronizing the time of data acquisition.

![Figure 1 - EM-31 mounted in non-conductive canoe.](image)

The effective depth of exploration for electromagnetic methods has been defined as the depth where 70% of instrument response is from the overlying material, and is controlled by variations in instrument design and acquisition parameters such as coil orientation (Stewart, 1982; Stewart and Bretnall, 1986). Effective depth of exploration on the TCSA with the EM-31 and EM-34 ranges between approximately 1 and 30 m.

The coil spacings and frequencies of the EM-31 and EM-34 are designed such that, where terrain conductivities are less than 80-100 mS/m, the ratio of the secondary magnetic field induced by eddy currents to the primary magnetic field is linearly proportional to the terrain conductivity over a homogenous sub-surface (McNeill, 1980a). Thus, EM instrument readings are expressed as apparent conductivity: the conductivity of a homogenous half-space that will produce the same response as that measured over the
real heterogeneous sub-surface when using the same acquisition parameters (Spies and Eggers, 1986). Throughout most of the TCSA terrain conductivities are greater than 100 mS/m; therefore, the raw instrument readings do not represent an apparent conductivity, or equivalent conductivity of a homogeneous subsurface. Nevertheless, following convention, the instrument readout here is referred to as the raw apparent conductivity ($\sigma_{\text{raw}}$).

To infer terrain conductivity structure based on raw apparent conductivity data in this high-conductivity environment, the raw data must be compared to layered models that incorporate EM instrument design and data acquisition parameters. In this study, models are restricted to simple horizontal layers with homogeneous conductivities. The conductivities of individual layers in these models are referred to as terrain conductivities or model layer conductivities ($\sigma_{t,1,2,\text{etc}}$). The instrument response predicted from the layer models is designated as predicted apparent conductivity ($\sigma_p$).

**Modeling of EM Data**

The forward modeling program PCLOOP was used to calculate predicted apparent conductivity ($\sigma_p$) over layered earth models (Geonics, 1994). PCLOOP calculates instrument response with an algorithm by Anderson (1979) that incorporates theoretical solutions by (Frishknecht, 1967; Kaufman, 1969; McNeill, 1980a). The testing of instrument sensitivity to various model parameters can be done with forward models. For example, forward models can predict the maximum practical exploration depth and conductivity sensitivity for a particular EM instrument in a given environment. Portions of the data were also interpreted using the EMIX 34 program (Interpex, 1994).
The EMIX program can perform forward calculations similar to those of PCLOOP using theoretical solutions published by (McNeill, 1980a; Patra and Mallick, 1980). However for this study, the program was used in an inversion mode. Given an initial model conductivity structure, EMIX can invert a set of raw apparent conductivity readings to find the best-fitting values of one or more layer conductivities or layer thicknesses (McNeill, 1980a; Patra and Mallick, 1980; Kaufman and Hoekstra, 2001) using a ridge regression estimation algorithm (Inman, 1975).

Appendices 3 to 7 contain the full set of EM and resistivity data collected in this study. Portions of these data were incorporated in EM models with depths ranging from less than a few meters for shoreline models to a maximum of 15m at upland sites. Model complexity was minimized by representing ground and surface water layers by just two or three conductively uniform and horizontal layers. Upper layer thickness and/or conductivity \( (σ_1) \) were constrained by other measurements. For example, for measurements made on land, an upper layer conductivity was set to a value determined by resistivity soundings to a local site with similar lithology. For readings over water, the water depth and conductivity were both measured; hence the properties of the surface water layer were known and fixed in the model. Because EMIX 34 takes into account the instrument height and dipole orientation (as well as coil spacing and operating frequency) the program could be used for data collected over shallow water with floating coils.
Equivalence in EM inversions

The very high terrain conductivities present on the TCSA produce a non-linear instrument response in the EM-31 and EM-34 that changes from a positive to a negative slope with increasing terrain conductivity (Figure 2, Appendix 2). Due to this response, there is a non-unique relationship between instrument readings ($\sigma_{raw}$) and terrain conductivities ($\sigma_t$), even for a simple condition, such as the one layer model consisting of a homogeneous half-space shown in Figure 2. Note that for ($\sigma_t$) of both 400 and 2100 mS/m, a ($\sigma_{raw}$) of 200 mS/m is produced (black dots in Figure 2). HMD response slope is negative beyond a ($\sigma_t$) of 9000 mS/m, which yields a ($\sigma_{raw}$) of 1730 mS/m which is beyond the ±1000 mS/m range of the EM-31 MK II used in this study. Similar response curves for the EM-34 are in Appendix 2.

Figure 2 - PCLOOP forward model of EM-31 response over a homogenous half-space with infinite depth (Geonics, 1994).
In settings such as the example in the preceding paragraph (EM-31 VMD response of 200 mS/m expected for terrain conductivities of either 400 mS/m or 2000 mS/m), particular care must be taken when using the EMIX 34 inversion routine to solve for the terrain conductivity. The inversion routine requires that an initial estimate for terrain conductivity be input by the user. If, for example, the true terrain conductivity is 400 mS/m, then the initial estimate given to inversion routine must be reasonably close to this true value. If the initial estimate terrain conductivity specified is closer to 2000 mS/m, the inversion routine will converge on 2000 mS/m rather than 400 mS/m. This need to have a reasonably good idea of which of the equivalently possible terrain conductivities is valid can be solved by either collection of more detailed EM data afforded by multiple dipole orientations, coil spacings and heights, or with resistivity measurements. Additional EM modes were not practical while using the EM-31 VMD over shallow high conductivity water because they lacked resolution, exceeded the instruments scale or were not possible when the instrument was logging while moving. Limitations of the use of EM in very high conductivity environments are discussed further in the results section below. In these cases, ambiguity was resolved by running resistivity soundings at representative sites. The terrain conductivities derived from a resistivity sounding were then used as the starting structure for inversions of EM readings in the vicinity of the resistivity sounding. In this way, local variations in terrain conductivity between sites of resistivity surveys could be mapped with the more rapid EM methods.
Resistivity Methods

Resistivity methods use arrays of electrodes that are driven into the ground, towed on a floating streamer or positioned on a stationary floating tube. Direct current is then introduced into the ground (or surface water) from a pair of current electrodes and the resulting potential differences at another pair or pairs of electrodes are measured. The circuit that is completed by these arrays includes the earth, groundwater and any surface water as a resistor whose resistance is empirically related to the source current and measured voltage by Ohms law (Koefoed, 1979). Depth and degree of spatial resolution are controlled by electrode spacing and the conductivity of the sub-surface (Koefoed, 1979). When compared to EM methods over the same target, resistivity methods are generally regarded as a more accurate and reliable estimate of apparent conductivity (Koefoed, 1979; Patra and Mallick, 1980; Kaufman and Keller, 1983; Kaufman and Hoekstra, 2001).

Use of Resistivity Data in Groundwater Studies

DC resistivity methods have been used extensively to estimate water quality, locate salt/freshwater interfaces, monitor contaminant plumes, and to locate aquifers e.g. (Cameron et al., 1981; Barker, 1990; Griffiths and Barker, 1993; Sharma, 1997; e.g. Aristodemou and Thomas-Betts, 2000; Fetter, 2001). The more cumbersome resistivity
methods are often combined with the faster but less accurate EM methods (McNeill, 1990).

Marine resistivity methods have been used to map zones of low terrain conductivity or sea bed conductivity that have been associated with submarine groundwater discharge (Vanek, 1991; Hoefel and Evans, 2001; Manheim et al., 2001). Towed dipole-dipole resistivity streamers built by Zonge, Inc. and Advanced Geosciences, Inc., were successful at locating prospective zones of submarine groundwater discharge which were subsequently confirmed by direct sampling (Swarzenski pers. comm., Manheim et al., 2001). However, these marine resistivity systems are limited to open water applications compatible with the draft of the boat and the turning radius of the typically long (~100m) towed streamer. Further, these commercial systems typically use a dipole-dipole array geometry which has lower vertical resolution than other array geometries such as the Wenner or Schlumberger.

**Resistivity Data Acquisition at the TCSA**

In this study, land based profiles were run with a 50-electrode Campus Geopulse resistivity system using the Wenner traverse geometry with electrodes spaced between 1 and 6 m. Resulting profiles were 50-300m long with effective depths of exploration of 0.2 to 50 m. For resistivity surveys over shallow (< 1m) water, a novel floating electrode array with Schlumberger geometry was constructed at the University of South Florida Geology Department, with electrodes spaced between 0.5 and 4m for an effective depth of exploration of approximately 1.5 m (Figure 3 and 14a, Edwards, 1977). Resistivity
measurements for the floating electrode array were made manually with a Terrameter SAS 300C resistivity system.

Figure 3 - Floating Schlumberger array connected to Terrameter SAS 300C operated by Arnell Harrison of the USF Geology Department Geophysics Lab.

Resistivity Data Interpretation

Land-based Wenner traverse resistivity surveys were inverted for apparent conductivity using the two-dimensional RES2DINV inversion program (Loke, 2002, Appendix 8,9,10). Marine Schlumberger sounding data were inverted for apparent conductivity using the one-dimensional 1IXD inversion program (Interpex, 2002, Appendix 8). Both of these programs assign each sub-surface grid node an initial terrain conductivity and then calculate the apparent conductivity that would result and iteratively adjusts the model until the RMS error is minimized to less than 5%.
Estimation of Porewater Conductivity from Terrain Conductivity

As discussed above, terrain conductivity is a function of porewater conductivity, temperature, conductive clay content, porosity, pore space shape and connection and degree of saturation (Keller and Frishknecht, 1970; McNeill, 1990). This relationship is summarized with Archie's law, used extensively in the oil exploration industry to calculate the porosity of oil reservoirs. Archie’s Law relates the formation conductivity $\sigma_t$ (equivalent to terrain conductivity) to the porewater conductivity ($\sigma_w$), in fully saturated media, by $\sigma_t = \frac{\sigma_w \phi^m}{a} + \sigma_c$. The $a$ and $m$ symbols are empirically determined constants, $\phi$ is the porosity and $\sigma_c$ is the grain surface conductivity attributed to clay (Archie, 1942; Keller and Frishknecht, 1970; Robinson and Coruh, 1988; McNeill, 1990; Sharma, 1997; Hearst et al., 2000). In a common alternative formulation, the relationship between terrain and water conductivity is described as the formation factor $F = \frac{\sigma_w}{\sigma_t}$ (Keller and Frishknecht, 1970; Fitterman and Deszcz-Pan, 1999; Hearst et al., 2000; Fitterman and Deszcz-Pan, 2001; Manheim et al., 2001; Stewart et al., 2002). This alternative formulation is commonly used in groundwater studies to find the relationship between terrain and porewater conductivity and to estimate a formation factor (F) for lithologic units of interest (Fitterman and Deszcz-Pan, 1999; Fitterman and Deszcz-Pan, 2001; Manheim et al., 2001; Stewart et al., 2002). Implicit in the alternative formation factor expression are that clay conductivity effects are small compared to those of
porewater conductivities, units are saturated and porosity variations are small within the units defined.

It was initially unclear what effect the variability in clay content, porosity and saturation on the TCSA would have on the reliability of formation factor calculations for different lithologic units of interest and the subsequent predictions of porewater conductivity from terrain conductivity where these units were defined. To determine a formation factor requires measurements of terrain conductivity and porewater conductivity at the same location and depth. The TCSA has 3-D spatial variability in terrain conductivity, so uncertainties in terrain conductivity estimates are expected when 2D models are used for computing terrain conductivities from resistivity data and, most importantly, 1D models are used to compute terrain conductivities from EM data.

To minimize uncertainties, formation factors were only computed where both resistivity surveys were made and water samples were collected. With these data, formation factors were computed in highly porous organic rich mangrove soils and at three depths within the clay rich Hawthorn Formation. Once a formation factor was determined using the more reliable resistivity methods on a particular lithology, then terrain conductivities derived from the more rapid EM methods were used to extend groundwater conductivity predictions out laterally until new lithologies were encountered. The efficiency of this method was then tested by comparing predicted porewater conductivities against directly measured porewater samples.
Terra Ceia Study Area

The TCSA is described as a nearly level coastal lowland with progressively rolling terrain to the east (Hyde and Huckle, 1983). Maximum relief in the study area is approximately 2 m with low-lying ridges and hammocks having slopes generally less than 2% (Hyde and Huckle, 1983; Carter et al., 2003; UF, 2003). Upland areas are comprised of maritime hammocks or fallow agricultural lands overgrown with invasive exotic plants. The lowlands are comprised of mangrove fringe forests, interior salt barrens and the following wetlands: freshwater creek, freshwater marsh, karst tidal ponds, karst freshwater ponds, high and low estuarine marshes, and transitional marshes (Hyde and Huckle, 1983).

Almost the entire upland area of the TCSA was cleared and farmed between 1890 and 1967. Numerous dredge and fill structures changed the shallow groundwater salinity (Figure 5). Even though the TCSA has been significantly altered from a natural state, it provides habitat for a wide variety of flora and fauna, including endangered species and economically important game fish. The State of Florida plans on restoring the TCSA to a more natural state, which will improve the wetland functions of flood water dampening and denitrification, as well as improve habitat for native species and mitigate invasion by exotic species (Mitsch and Gosselink, 1993; Bendjoudi et al., 2002).
Figure 4 - Location of the TCSA within Tampa Bay and Florida. USGS 1:24K scale shoreline basemap with UTM NAD83 Zone 17 datum.
**Surficial Geology**

Exploration depths of the geophysical methods employed in this study are limited to the upper 50 m, which consist of poorly drained, moderately permeable Pliocene to recent surficial sediment underlain by the Miocene Hawthorn Group phosphatic sand, clay, marl, and intermittent beds of fossiliferous limestone that form the upper confining unit of the Floridan Aquifer (Miller, 1997). This lithology contains a high content (22-40%) of electrically conductive clays such as illite, kaolinite, palygorskite, sepiolite, and smectite (Hyde and Huckle, 1983; Compton, 1997).

**Core Samples**

Hyde and Huckle (1983) mapped virtually the entire upland area soil type of the TCSA as Bradenton fine sand with minor occurrences of Wabasso fine sand, both of which formed from the underlying Hawthorn Group (1983). This classification scheme is limited to the upper 2m of sediment and was based on shallow hand auger type core samples. The frequently flooded portions of the study area consist of Wulfert-Kesson type soil, which also formed from reworked Hawthorn Group sediment (Hyde and Huckle, 1983).

The USGS collected a 15m hydraulic rotary core and installed the TC1 multi-port well (Figure 5 and 6) in February 2002. The State of Florida Department of Environmental Protection took seven 3-m depth percussion driven split-spoon cores and gave sample splits to the author for this study in October of 2002 (Figure 7). Penetration of the split-spoon cores was limited to 3m by a thin limestone layer that was sampled with the 15m USGS rotary core. All cores on the TCSA showed a surficial 0.5 to 1 m
layer of organic-rich quartz sand grading into underlying iron-stained clay and marl with the deeper rotary core showing clay and marl with intermittent thin (<10cm) limestone layers at 3, 10 and 15 m. These core samples resemble descriptions of the Hawthorn Group sediment in Sarasota and Manatee County (Barr, 1996). Visual inspection of grain size, texture and mineralogy from these 8 cores suggests that to a depth of 15m, the upland portions of the TCSA may have a fairly uniform lithology comprised of Hawthorn formation clay and marl with intermittent limestone overlain by Bradenton Fine Sand soil.

Three vibra-core samples were taken by the USGS, one in the center and two in the adjacent mangrove wetlands of Moses Hole pond (Figure 5, Area 2). The center of Moses Hole is characterized by 2 meters of bioturbated phosphatic quartz sand with occasional 1-2 cm clay and mud lenses and small <2cm shell fragments which then terminates in 60cm of cohesive clay with abundant semi-lithified limestone clasts that resemble those found in upland cores on the TCSA. Two adjacent mangrove wetland cores consisted of approximately 60cm of spongy organic rich mangrove peat mixed with sand grading into 70cm of cohesive clay with mud lenses similar to the center of the pond, but lacking limestone clasts and appearing to be considerably more porous.
Figure 6 – Schematic of the USGS TC1 multi-port well plotted on a RES2DNV (Loke, 2002) inversion profile of Wenner array resistivity data (Appendix 9-10).
Figure 7 - Location of split-spoon cores and predominate soil types on the TCSA (boxed area). Bradenton and Wabasso soils predominate the upland areas and Wulfert-Kesson forms the wetlands (Hyde and Huckle, 1983).
Climate

Manatee County receives an average of approximately 127 cm of rainfall a year, with 66% occurring in the wet season (May to September). The mean temperature is 21.1 degrees Celsius (°C). Tides are 80% semi-diurnal and 20% diurnal and average 82.7 cm (Hyde and Huckle, 1983). Data collection on the TCSA began in April of 2001 and continued into Fall of 2002, with both years having greater than normal rainfall of 144 and 164 cm respectively.

Hydrology

The TCSA is bordered by the saline waters of the Tampa Bay estuary to the north and the fresh to saline Frog Creek to the south (Figure 4 and 5). Before portions of Frog creek and related wetlands were drained or filled with causeways, the TCSA was an island bounded by wetlands and restricted marine waters. The slow-moving and meandering Frog Creek headwaters begin in a fresh water wetland complex 12.5 km to the east of the TCSA. Salinity becomes stratified in Frog Creek as the slow moving, fresher and less dense waters from the east mix with the tidal, more saline and dense waters of Tampa Bay to the west.

Numerous round ponds dot the landscape of the TCSA, however no research was found that classified these ponds as active karst features or conduits between surface waters and the Floridan Aquifer (Figure 5). Pond salinity is controlled by marine waters flooding through mosquito control ditches or natural creeks, rainfall, and mixing with the surficial aquifer. Ponds in lower elevation terrain are more frequently flooded and tend
to have higher salinities. Heavy rainfall during the wet season can cause short term salinity stratification in surface water bodies with fresher water temporarily overlying more saline water.

Hyde and Huckle (1983) report that if undrained, the TCSA soil types will have a water table within 20 cm of the surface for 2 to 6 months of the year and a depth of 25 to 100 cm for much of the rest of the year (1983). The TCSA has been extensively modified by dredge and fill structures, which may lower ground and surface water levels. Higher than normal rainfall during 2001 and 2002 probably raised the water table above normal ranges, nevertheless, SWFWMD well data show the water table to be closer to 50 to 100 cm during the wet seasons of 2001-2 (Figure 8). Slow drainage and standing water were observed after precipitation. The USGS TC1 well and 11 SWFWMD wells have hydraulic heads that are above high tide during the wet season (Figure 9,10). Thus, favorable conditions exist for submarine groundwater discharge into the shallow waters of Tampa Bay.

![Figure 8 - SWFWMD shallow water table levels near EM collections sites.](image-url)
Figure 9 - Hydraulic head distribution at the USGS TC1 multi-port well using North American Vertical Datum of 1988. The TC1 well has greater pressure with depth and positive hydraulic head for all ports accept the 2 shallow ports during the dry season in May of 2003 when compared to local mean sea level in Tampa Bay.

Figure 10 - Hydraulic head distribution for 11 SWFWMD wells within the TCSA during geophysical data collection period using the North American Vertical Datum of 1988 and the local mean sea level of Tampa Bay.
Results

**EM-34 and EM-31 Data Coverage**

EM-34 and EM-31 readings over both land and water were acquired in various modes on the TCSA (Appendices 3-7). Both lowland and upland sites contain EM-34 HMD mode data, while the analog EM-34 VMD mode was in most cases limited to upland regions where terrain conductivities were less than 600 mS/m (Appendix 2 and 6). EM-34 readings over water were not compatible with values predicted from reasonable models. The EM-31 model MK2 was successfully used in VMD and HMD mode at ground level (0 m) and hip height (0.9m) over land, and floating (0.1m) over shallow (<1.5m) marine salinity water (4000-5000mS/m) in the study area.

**Shallow Marine EM-34**

Nadeau et al. (2003) showed a floating EM-34 in VMD mode could be used successfully in freshwater (10 mS/m) to image lake floor conductivity between 0.3 and 42 mS/m allowing for a minor numerical correction for water conductivity and depth. No studies were found, however, for saline environments that require the application of models such as PCLOOP and EMIX to correct for water depth. The concept of floating an EM-34 over saline water and attaining useful information on seabed conductivity was tested by a suite of PCLOOP forward two-layer models (Figure 11,12,14c) with upper layers run using a value of 4550 mS/m (common in surface waters on the TCSA) and at
water depths from 0.2 to 1.5m. These models incorporated a lower layer seabed conductivities ranging from 10 to 3000 mS/m and spanning the value of ~1000 mS/m expected on the TCSA based on resistivity data.

These models show that negative apparent conductivity readings will likely occur in VMD mode unless an unlikely <200 mS/m seabed is encountered, which limits the analog EM-34 available in this study to HMD mode because only positive apparent conductivity readings up to 300 mS/m can be measured (Appendix 2). A gauge replacement or rewiring may solve this problem (Stewart pers. comm.). Figures 11 and 12 show that VMD and HMD mode could in theory provide useful information on seabed conductivity in shallow water under a variety of conditions.

Both VMD and HMD mode were tested, and as expected, no readings were attained in VMD mode. HMD mode readings at 10 and 20 m coil spacings contained noise that was associated with small movements of the floats in waves and wind. The 40 m coil spacing HMD data was the least affected by this surface noise, as expected from response curves (McNeill, 1980b). EMIX two-layer inversion models were created for each of the 27 floating stations (see Appendix 3,4,5 for locations over water) which included an upper layer set to the surface water depth and conductivity and included HMD apparent conductivity data at combined 10, 20 and 40 m coil spacings. EMIX inversions converged at the same solution when using starting lower layer seabed conductivity values of both 200 mS/m and 3000 mS/m. These starting values were chosen based on EM-31 data showing a seabed conductivity of 3000-5000 mS/m at the edges of Moses Hole pond and between 340 and 1880 mS/m in the middle (discussed in the mangrove salinization section below).
Unfortunately, EMIX models of floating EM-34 data, including individual runs of the relatively noise free 40 m coil spacing data, produced implausibly high or low seabed layer conductivities. Simple 2-layer PCLOOP forward models with seabed conductivities between 100 and 5000 mS/m also failed to fit the observations. The misfit between the EM-34 observations and any reasonable 2-layer model may be linked to the following factors: three-dimensional conductivity variation at the scales imaged, water depth error, surface water conductivity variation not accounted for, EM coil misalignment or movement, or instrument calibration. Further study would be needed to determine the importance of the various factors mentioned above.

In summary, forward models predict that the floating EM-34 VMD and HMD modes have potential for measuring useful information on seabed conductivity. The field experiments on the TCSA, however, were unsuccessful at reproducing these theoretical results. Given the space needed for EM-34 measurements (10-40 meters between coils), instrument development efforts targeting these settings may be better focused on short marine resistivity streamers (~10-50m).
Figure 11 - PCLOOP two-layer forward models of EM-34 VMD response over shallow marine water.
Figure 12 - PCLOOP two-layer forward models of EM-34 HMD response over shallow marine water.
Shallow Marine EM-31 Data

The relatively small size of the EM-31, relative to resistivity streamers or the space needed between EM-34 coils, may offer the possibility of profiling in otherwise inaccessible coastal terrains. Application of the floating EM-31 method to discriminating seafloor conductivities and the conditions under which it might be successful are discussed next. HMD mode data were not used because two-layer PCLOOP models predict that saline water depths as shallow as 0.1m, even when combined with seabed conductivities as low as 200 mS/m, produced out of range readings (>1000mS/m). Field trials proved (1) the HMD mode has significant noise problems associated with sensitivity to near surface materials and movement of the floating coils (McNeill, 1980a) and (2) rotating the instrument coils between VMD and HMD mode while streaming data was impractical.

Floating EM-31 in VMD mode shows greater promise, with limitations (Figure 13). Conclusions from analysis of PCLOOP two-layer forward models of EM-31 VMD data (Figure 13) include (1) at depths greater than 0.70 m, seafloor conductivities less than ~1000 mS/m are distinguishable from one another and (2) equivalence issues exist at shallower water depths, where low \( \sigma_f \) (~100 mS/m) and a high \( \sigma_t \) (~2000 mS/m) may yield similar data.
Figure 13 - PCLOOP two-layer forward models of EM-31 VMD response for changing water column thickness and lower layer conductivity. Note the relative lack of sensitivity to 100 mS/m changes in lower layer terrain conductivity at water columns greater than 0.75 m. Also note the approximately equivalent readings for lower layers of 10 to 1000 mS/m at a water column of 0.75m.

To test whether the EM-31 actually performs as predicted by these models, an experiment was conducted floating the EM-31 in shallow seawater at the location shown in Figure 5, Area 3. Stationary time series EM-31 VMD readings were taken during a rising tide, with the assumption that changes in subseafloor terrain conductivity during this period were small. Data were compared against PCLOOP forward models, with the expectation that all readings should be compatible with approximately the same subseafloor conductivity.
Figure 14 – Floating Schlumberger Array (A), Floating EM-31 (B), Floating EM-34 (C) and a cross-section of the floating EM-31 calibration model (D).
A non-conductive canoe held the EM-31 instrument 0.1 m above the water surface and was laterally fixed, but allowed to rise with the tide along plastic poles driven into the sediment. The EM-31 was programmed to log readings every 3 minutes for 14.75 hours over one half of a Tampa Bay tidal cycle (Figure 14b and 15a). The VMD mode was chosen in order to limit the effect of the highly conductive surface layer of seawater and because rotating the EM-31 to HMD mode inside the canoe while logging was not practical. A site shielded from wind and waves was chosen in Bishop Harbor (Figure 5, Area 3). A Van Essen conductivity, temperature, depth sensor (CTD) logged readings every 10 minutes at the sediment seawater interface directly beneath the EM-31 while manual readings of the upper water column were measured with a YSI-30 probe (Figure 15 and 17). Field trials found no conductivity effect from placing the small stainless steel CTD (2 cm diameter by 26 cm length) directly beneath the EM-31 (the in phase component of the EM-31 signal may have been able to detect the CTD, but was out of range in this high conductivity environment).

For interpretation of the EM results, a resistivity sounding was run and porewater samples were then collected at the site of the EM-31 experiment. The resistivity sounding was conducted with the floating Schlumberger array and inverted for terrain conductivity using a two-layer IX1D model with the upper layer fixed to the water column measurements of 4702 mS/m and 0.87m. A lower layer conductivity of 1170 mS/m provides the best fit to the observations, with an RMS error of 7.8% (Figure 16).
Figure 15 - (A) Raw EM-31 VMD readings for over 14 hr of a rising tide. Note the inverse response of water column thickness to raw apparent conductivity and the raw apparent conductivity shifts between 09:00 and 13:00hrs. (B) Correlation of raw EM-31 VMD and water column thickness readings for over 14 hr of a rising tide.
Figure 16 - IX1D two-layer model of floating Schlumberger array resistivity data over 0.87 m of 4702 mS/m marine salinity (28.7 ppt) water with a theoretical best fitting lower layer (sea-bed) conductivity of 1170 mS/m.
Figure 17 - Upper and lower water column conductivity and temperature beneath the floating EM-31 (water sampling method in Appendix 1). Noise centered around 07:00hr may be due to the author walking near the CTD on the sea floor, which may have stirred up conductive clays or released more saline porewater.
A porewater sample was obtained from 1 meter beneath the sediment seawater interface beneath the canoe (Appendix 1) and was slightly lower in conductivity ($\sigma_w = 4270$) than the overlying surface water (Figure 17). Combining this porewater value with the resistivity sounding result yields a formation factor $F = \sigma_w / \sigma_t = 3.65$. A formation factor of 3.65 at a depth of 1 m below the sediment seawater interface is consistent with data from resistivity probes of core samples in Hawthorne Group clays in other shallow marine sites in Tampa Bay (Manheim, pers. comm.), thus increasing confidence in the floating Schlumberger array resistivity-derived terrain conductivity.

To determine whether the EM readings are in agreement with the resistivity-derived subseafloor conductivity of 1170 mS/m, a set of two-layer forward EM models were run using this lower layer value. The upper layer thickness was set to the water column measurement at the corresponding time (blue dots in Figure 15). For each model, the upper layer (water column) was set to a uniform conductivity equal to the average of the upper and lower water conductivities measured at that time (Figure 17). The orange triangles in Figure 18 show the forward model results simulating eight different times during the experiment.

Lower layers of 10 and 2000 mS/m were run for comparison purposes (Figure 18). Clearly, predicted apparent conductivities calculated with lower model layers of 10 and 2000 mS/m do not match measured values as well as the 1170 mS/m lower model layer (Figure 18). Readings between 9:15 and 13:00 hrs show the poorest fit in the 1170 mS/m model, which corresponds with a time window that begins and ends with shifts in
raw apparent conductivity that seem unrelated to water column measurements (Figure 15, 16 and 18).

![Graph showing comparison of predicted to measured apparent conductivity]

Figure 18 - Comparison of predicted to measured apparent conductivity for 8 two layer PCLOOP models with upper layers fixed to surface water data and lower layers set at 100, 1170 and 2000 mS/m (Geonics, 1994).

The strengths and limitations of this use of the EM-31 are highlighted in the experimental and model results in Figure 18. The primary limitation is equivalent solutions, which are most severe for water depths of 0.65-0.75 meters (for surface water of ~4550 mS/m), as seen in the model suite in Figure 13. At this depth range, all lower layer conductivities of ≤1000 mS/m yield equivalent predicted apparent conductivity readings. In practice, similar equivalent results occur during the 9:15-11:00 hr range in Figure 18 when the 1170 mS/m and 100 mS/m model predictions and observed raw
apparent conductivities converge. It is clearly difficult at these water depths to resolve the lower layer conductivity based on models of raw apparent conductivity readings.

The growing and abruptly terminating discrepancies between the 1170 mS/m model and the readings between 8:00 and 10:00 hr further illustrate the uncertainties in this method and the need for good calibration against other data. The cause of this discrepancy is unresolved as it coincides only with a slowing in the rate of water rise into the bay and not with detectable changes in the water column or any changes to the instrument set-up.

At the shallowest and deepest water depths encountered (< 50 cm and > 1.0 m) there is remarkably good agreement between the resistivity results and the EM-readings. Within these depth ranges, EMIX inversions of the EM readings for lower layer conductivity would yield values close to the “observed” resistivity value (Figure 18). Further tests of the EM-31 in shallow coastal waters are described below in the context of comparing observed and EM-predicted porewater conductivities.

**Correlation of Terrain Conductivity and Porewater Conductivity**

Formation factors in Table 1 were calculated from resistivity surveys coincident with porewater sampling (Figures 19 and 20). Formation factors are lower in the Hawthorn Group (2.5-2.9) than in the mangrove soils (3.65) which is expected as there are conductive clays present in the Hawthorn Group (see lithologic descriptions in the Introduction). These values are similar to results obtained for sediment resembling the Hawthorn Group 50 km to the north in Tampa Bay (Manheim pers. comm.) Results of porewater conductivity predictions using this formulation are discussed below.
Resistivity-derived formation factors were applied to 12 unique EM models from 8 sites with directly measured porewater samples (yellow dots in Figure 20 and Table 2). A reasonable degree of correlation exists between the measured and predicted porewater conductivity for 12 samples (Figure 21 and Table 2). A similar correlation is plotted for aerial electromagnetic data over relatively clay free sediment in a study area ~330 km
south in the Everglades National Park, Florida (Fitterman and Deszcz-Pan, 2001). This study used three types of EMIX models to predict porewater conductivity. (1) Seven two-layer models used EM-31 data over single port wells. The upper model layer, designed to represent the unsaturated zone, was fixed to the shallow conductivity derived from a nearby resistivity line (200mS/m). This upper layer was set to the thickness of the unsaturated zone based on water level in the well. Inversions were run with an initial lower model layer conductivity set to values based on nearby resistivity data (blue circles in Figure 21). (2) Two two-layer models for EM-31 data over water incorporated upper layers with direct measurements of surface water depth and conductivity. Initial model lower layer conductivity was set to values based on nearby resistivity data (red points in Figure 21). (3) Three models had three-layers that used EM-34 VMD and HMD data at three coil spacings with bottom of the model layers set to the mid-point of the screened intervals of the TC1 well and starting values based on resistivity data (green points in Figure 21). Water levels and associated unsaturated zone effects were not accounted for in these three models.

EM-31 data at seven locations over land (blue circles) plot closer to the one-to-one line (black line) than the three EM-34 data points (green points), which may be due to the following factors. (1) The unsaturated zone accounted for by resistivity and well data in the EM-31 models has a significant effect on terrain conductivity not accounted for with the EM-34 models. Expanding the models for the EM-34 soundings to include an unsaturated zone layer may improve their predictive capabilities. (2) The EM-31 samples a smaller and thus probably more conductively homogenous volume relative to the EM-34. In addition, the general case where porewater predictions that are too low
may be caused by poor estimates of the unsaturated zone (determined at a nearby resistivity survey), where-as predictions that are too high may be caused by an increase in clay content at the EM site. The misfit between observed and predicted porewater conductivities over larger depth ranges derived from three-layer models of EM-34 data suggests that this method at best distinguishes the general range of salinity trends (freshwater, brackish, saline, or hypersaline). EM-31 readings targeting shallow porewaters, however, may be useful at distinguishing salinity trends within smaller areas.

Very small error is expected in the measured porewater conductivity measured by a calibrated YSI-30 probe relative to the predicted porewater conductivity (instrument specifications in Appendix 1). Predicted porewater conductivity error bars were not feasible in this study because they comprise an unknown combination of formation factor error caused by variations in clay content, saturation, EMIX, RES2DNV and 1IXD model error as well as other errors associated with EM and DC data acquisition, such as coil misalignment and instrument calibration.

While the uncertainties in estimating porewater conductivity from calibrated EM data may be considerable, this method appears adequate to establish trends of porewater conductivity on the TCSA. The following five factors probably influenced the relative success of using surface geophysical methods to sense porewaters at depth on the TCSA: 1) Relatively flat and consistent lithology in the upper 30 m of exploration depth, 2) predominately saturated formations overlain by a thin unsaturated zone, 3) large conductivity contrasts between targets (freshwater, saline and hypersaline water saturated formations), 4) predominately high salinity porewaters dominated the apparent
conductivity signal and limited the effects introduced by conductive clays, and 5) a lack of power transmission lines and conductive anthropogenic materials that interfere with EM soundings.
Figure 20 - Porewater and geophysical data used to calculate formation factors. Multiple depths were available from the TC1 well with Wenner array RES2DNV inversion profile, EM-34 and EM-31 data. Single depth porewater data was available from the Schlumberger array location (Figure 16). The remaining stations have single depth wells and EM-31 data.
Figure 21 - Predicted vs. measured porewater conductivity based on EM models and local resistivity derived formation factors. EM-31 over land (blue circles), EM-34 (green points) and EM-31 over water (red points). A one-to-one correlation would fall on the black line. Error and formation factors are discussed in this chapter.

Table 2 - Predicted and measured porewater conductivity based on EM models, direct samples and local resistivity formation factors from Table 1.

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<th>Coil height (m)</th>
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<td>EM-31</td>
<td>3.67</td>
<td>0.9</td>
<td>VMD</td>
<td>0.0</td>
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</tr>
<tr>
<td>2.42</td>
<td>0.0</td>
<td>3209</td>
<td>2647</td>
<td>2.56</td>
<td>EM-31</td>
<td>3.67</td>
<td>0.9</td>
<td>VMD</td>
<td>3.0E-03</td>
<td></td>
</tr>
<tr>
<td>2.53</td>
<td>0.0</td>
<td>189</td>
<td>197</td>
<td>2.56</td>
<td>EM-31</td>
<td>3.67</td>
<td>0.9</td>
<td>VMD</td>
<td>0.2</td>
<td></td>
</tr>
</tbody>
</table>
Imaging Submarine Groundwater Discharge

Discerning freshwater, seawater, and hypersaline porewaters was successful using the method discussed above. Locating zones of submarine groundwater discharge (SGD), however, typically requires the identification of more subtle conductivity anomalies that occur when an upward flux of fresher groundwater mixes with more saline surface water at a few meters below the sediment seawater interface. The magnitude of EM $\sigma_{\text{raw}}$ anomalies expected in association with SGD are examined next.

While the water table data on the TCSA suggests SGD may occur (Figures 9 and 10), as of the date of this publication, it has only been predicted in groundwater models and has not been directly measured here or elsewhere in Tampa Bay using seepage meters, piezometers or geochemical tracers (Swarzenski pers. comm.). Thus no sites were available within the TCSA for directly examining potential conductivity effects of SGD. Further investigations beyond the scope of this study are needed to determine if the TCSA or other sites within Tampa Bay have significant SGD.

For the purposes of estimating EM instrument response to SGD conductivity anomalies in a setting such as Tampa Bay, we can use the results from a low-conductivity anomaly recently identified from a Tampa Bay marine resistivity survey located 29 km north of the TCSA, in 3.7m of water and 1.1 km from shore. Porewaters squeezed from a vibracore at the site of the resistivity anomaly revealed salinity that was 6.1 ppt fresher than the surface water at 5.0 m below the sediment seawater interface (unpublished data).
This fresher porewater may indicate that an upward flux of fresher groundwater has mixed with saline surface water at 5m below the sediment seawater interface. Using the pressure and temperature from the vibracore site (see method in Appendix 1) and a formation factor of 3.7 that was measured within the TCSA and similarly within Tampa Bay (this study and Manheim pers. comm.), a porewater salinity low anomaly of 6.1 ppt would theoretically lower the bulk seabed conductivity by 270 mS/m.

Floating EM methods used in this study are not suited to the water depth of the sample described above. For testing purposes, the existence of a similar anomaly in seafloor sediments beneath shallower water depths is assumed in a suite of PCLOOP 2-layer forward models with a water column conductivity common on in the TCSA of 4600 mS/m and surface water depths of 0.3, 0.7 and 1m. Lower model layers were set at 1200 mS/m (based on a resistivity measurement at the TCSA) and then lowered to 930 mS/m to simulate the hypothetical “SGD” anomaly. The EM-31 models predict an apparent conductivity change of between 7.0 and 22 mS/m in VMD mode with HMD response falling outside the instruments range (Figure 22). A change of 7 mS/m in EM-31 readings is detectable when the instrument is held stationary, but would be within noise levels if the instrument were towed rapidly or run in any but calm conditions.

Identical EM-34 models were run in VMD and HMD mode at 10, 20 and 40 meter coil spacings (Figures 23 and 24) with an apparent conductivity change of between -34 and -79 mS/m, which is detectable using the float system in this study. The EM-34 HMD apparent conductivity response to this anomaly was on the order of 1-20 mS/m (Figure 24), which is most likely within noise levels and not detectable.
At 0.3 to 1.0 m water depth, the SGD anomaly discussed above thus appears detectable with the EM-34 in VMD mode and at or below the detection limit of the EM-31 and EM-34 HMD. Thus although EM methods offer access to terrain inaccessible to marine resistivity methods, they lack the resolution needed for identifying zones of diffuse SGD. Clearly these techniques will have greater success in identifying SGD anomalies that have a higher porewater conductivity contrast.

The shallow exploration depths and resolution of the EM methods used in this study preclude estimations of the 3-D volume of SGD anomalies, although estimates of aerial extent and concentration are feasible. Calculating SGD flux would be feasible with dual density numerical groundwater flow models based on hydraulic head distribution and information on the aerial extent and concentration of SGD zones (Voss, 1984; SUTRA, Souza, 1987; SEAWAT, Guo and Langevin, 2003).

Other processes in tropical and sub-tropical climates, such as Tampa Bay, can further complicate locating SGD anomalies, regardless of the geophysical or direct sampling techniques used. For example, the subtle anomaly discussed above was observed in open water 1.1 km from shore, but if it were closer to shore, it may have been reduced or completely masked due to the mangrove soil salinization process discussed next.
Figure 22 - PCLOOP 2-layer forward models of floating EM-31 VMD response at three saline water depths over a SGD anomaly. A background seabed conductivity (blue circles) and a lower SGD influenced seabed conductivity (black squares) are shown.
Figure 23 - PCLOOP 2-layer forward models of floating EM-34 VMD response at three saline water depths over a SGD anomaly. A background seabed conductivity (blue circles) and a lower SGD influenced seabed conductivity (black squares) are shown.
Figure 24 - PCLOOP 2-layer forward models of floating EM-34 HMD response at three saline water depths over a SGD anomaly. A background seabed conductivity (blue circles) and a lower SGD influenced seabed conductivity (black squares) are shown.
Effect of Mangroves on EM Measurements

At the TCSA, the very highest terrain conductivities are found at shallow depths near mangroves (Figure 25, 26D, Appendix 7). Previous studies indicate that mangrove roots can uptake saline water and exclude 90-99% of all salt; therefore, leaving behind a concentrated solution in the soil (Scholander, 1968; Passioura et al., 1992; Tomlinson, 1994). This process raises soil porewater salinities until a quasi-steady state is reached, in which the flow of salt in the soil by convection in the seawater traveling to the roots is equaled by diffusion of the concentrated solution of salt water back to the soil surface (Passioura et al., 1992). A model of this process took 20 days to double the salinity of the porewaters within the upper 40 cm of intertidal mud (Passioura et al., 1992), which would correspond to a porewater conductivity of 8000-12,000 mS/m in the TCSA (Figure 27).

As porewater salinity concentration by mangroves may dominate nearshore salinity patterns, and have not been widely described in hydrogeologic contexts, we sought to investigate their extent and associated electromagnetic anomalies. Mangrove root depth and extent are controlled by mangrove species, transpiration levels, porewater salinity, soil flushing and bioturbation (Passioura et al., 1992). Shallow hand auger samples in the TCSA show that the *Rizophora mangle* (red mangrove species) trees that typically line the shores of Tampa Bay and the saline and brackish water ponds in the TCSA extend a dense shallow network of feeding and drinking roots up to approximately
5 m from shore and to a depth of 10-15 cm below the sediment seawater interface. This root distribution is typical of this species of mangrove (Passioura et al., 1992; Tomlinson, 1994).

On a transect running seaward from a red mangrove forest into Tampa Bay, extremely high shallow porewater conductivities of two to three times surface water levels were found near (~2m) and at the same depth (0.3-0.7m) as this network of roots and then fell off to background surface water conductivities within 5-10 m (Figure 5 Area 2, Figure 27).

Figure 25 - Porewater conductivity transect at 0.3 - 0.7 m sediment depth leading away from a red mangrove forest going towards Tampa Bay. Note a high contrast with the surface water conductivity ~ 4220 mS/m near mangroves. Distances are in meters from the nearest red mangrove tree trunk (location in Figure 5, Area 2).

The dramatic mangrove conductivity effects are easily detectable with EM-31 surveys. Anomalous EM readings are expected to extend a few meters beyond the zone of elevated porewater conductivities, as mangrove root soil salinization may have a similar EM lateral detection limit as shallow buried metallic targets such as buried steel drums, unexploded ordnance and steel pipe. Studies of 0.5-1.0m depth metallic targets
reported the first detection of anomalies at a distance of 1-5 m using the EM-31 in VMD mode at 0.9m height (McNeill, 1980a; Westphalen and Rice, 1992; Vogelsand, 1995; Bailey and Sauck, 2000; Barrow et al., 2000). Two EM transects perpendicular to mangrove zones and a more detailed 3-D grid are discussed below.

The first EM mangrove transect was a combined marine and land profile shown in Figure 26 that travels from a grass covered upland area through a 10m wide line of red mangroves (*Rhizophora mangle*) and then out across approximately 1 m deep saline water. Both VMD and HMD readings were taken at a constant instrument height of 0.1m. The EM-31 was oriented parallel to the shoreline in order to maximize resolution of potential anomalies smaller than the coil spacing of the instrument (Geonics, 1995). A suite of two-layer EMIX models that incorporated water column measurements in the upper layer and starting values based on resistivity at Bishop Harbor (Figure 16) was used for locations over water. For locations over land, a suite of one-layer EMIX models was run with starting values based on a resistivity measurement at a similar upland site (Figure 19). The lower model layer terrain conductivity anomaly associated with the mangroves (Figure 26) extends approximately 1 m to either side of the expected mangrove root zone (see shallow auger samples discussed above), which agrees with the porewater profile extending into Tampa Bay shown in Figure 25 and with the lower end (~1m) of lateral detection limits for case studies of shallow high conductivity targets.
Figure 26 - Lower model layer conductivity beneath open marine water, mangrove trees (centered at 0m) and upland vegetation. Lower model layer terrain conductivity was calculated from EMIX two-layer models at a constant instrument height over water and one-layer EMIX models over land (location in Figure 5, Area 3). Porewater sample of 4270 mS/m from 1m depth is located at +8m. Surface water = 4140 mS/m.

One porewater sample was available near this profile at Bishop Harbor (Figure 26) from a sediment depth of 1m and located at +8m from the nearest mangrove trunk. Porewater conductivity from this sample (4270 mS/m) is very close to that of the surface water conductivity (4140 mS/m), suggesting that porewater flushing is sufficient at 8m from the nearest mangrove trunk to dilute to a background level close to that of surface water. This porewater sample is in accordance with the EM profile (Figure 26), which shows terrain conductivities close to open marine water values 8m from mangrove trunks. Thus the Bishop Harbor land-marine profile (Figure 26) and the porewater conductivity transect extending into Tampa Bay (Figure 25) show a similar scale (~5-10 m) for the extent of the zone of hypersaline porewaters surrounding mangroves.

A second EM transect perpendicular to mangrove zones indicates considerably broader zones of hypersaline waters. This second EM transect consists of floating EM-31 VMD data across Moses Hole pond (Figure 27). Raw apparent conductivities were
interpreted with two-layer EMIX models with water column data comprising the upper layer and the lower layer initially set based on resistivity data. The best fitting lower model layer (seabed) conductivity is plotted as a function of distance (easting) across Moses Hole pond in Figure 27. These data show decreasing seabed conductivities with distance from the mangrove shorelines of over 50 m from the western shore and over 125 meters from the eastern shore. The extended distances of the anomalously high conductivities present in Figure 27 (50-125m) are an order of magnitude larger than those seen on the transects extending into Tampa Bay and Bishop Harbor (Figures 25 and 26).

One possible explanation for the apparently different scales of mangrove effects is that flushing of Moses Hole Pond is much more restricted than the movement of Tampa Bay surface waters (Smith and Swarzenski pers. comm.). The Moses Hole Pond is only indirectly connected to Tampa Bay by mosquito ditches and a tidal creek (Smith and Swarzenski pers. comm.). More rapid flushing in Tampa Bay and Bishop Harbor may reduce the extent of the zone of mangrove salinization and associated EM anomalies relative to that preserved in Moses Hole.
Figure 27 - Profile of floating EM-31 VMD lower model layers across Moses Hole pond (Figure 5, Area 2). Data processed using two-layer EMIX models. Aerial view of raw data visible as East-West running transect in Appendix 7. Surface water ~ 4500 mS/m.

An EM-31 grid within a mangrove forest on the TCSA further indicates that terrain conductivities within and around mangrove vegetation zones are not uniform. Figure 28 shows a grid of EM-31 VMD and HMD soundings over very shallow water within the banks of a mosquito control ditch leading into Moses Hole pond (Figure 5, Area 2). At this site, direct sampling by a drive point piezometer in the ditch produced hyper-saline 6125 and 7155 mS/m porewater conductivities at 31 and 61 cm respectively (circle with cross on Figure 28D). Surface water conductivities were 4500 mS/m ± 80 and water depths ranged between 0.1 and 0.23 m. The elevation of the ditch banks was consistent at ~1m above the water level and straight, which allowed for the rectangular sampling grid shown in Figure 28. All EM-31 readings over the ditch were taken at a constant height above the water of (0.8m) by carrying the instrument with the antennae.
oriented parallel to the ditch in order to minimize the effect of conductive anomalies smaller than the coil spacing (McNeill, 1980a). The HMD mode raw apparent conductivities and water depth clearly correlate (Figure 28, A and C) which is expected from this modes sensitivity to near surface materials (McNeill, 1980a). To interpret the raw apparent conductivities (Figure 28 A and B), at each point a two-layer model was created in EMIX using surface water information in a fixed upper layer and a nearby floating Schlumberger resistivity model as a starting point for the unknown lower layer. These models converged with a mean RMS error of <1%. The lower model layer conductivity (Figure 28D) not only shows differences laterally across the ditch, but also significant variability along the length of the ditch. In particular, extremely high terrain conductivities are derived for a portion of the eastern shore just south of the porewater sampling site. This high conductivity anomaly in Figure 28D is associated with mangroves that appear sickly and smaller than surrounding trees, which may be due to stress from hypersaline porewaters (Smith pers. comm).

Using the resistivity derived formation factor of 3.65 from a nearby site on the TCSA, the predicted porewater value shown with the circle with cross symbol on Figure 28D is remarkably consistent with the measured value. The predicted porewater value is of 6620 mS/m, which differs from the measured value of 7155 mS/m by only ~8% (Table 2).

Observations at the TCSA suggest that the EM-31 is a useful tool for measuring variability in porewater salinity within as well as adjacent to mangroves. Porewater salinity extremes may be associated with poor mangrove health; however the causes and consequences of this relationship are beyond the scope of this paper.
Figure 28 - EM-31 survey in mosquito control ditch lined with red mangrove trees. Raw data (A,B), water depth (C) and lower EMIX model layer and porewater sample (D).
CONCLUSIONS

At the TCSA study site, discerning between freshwater, seawater, and hypersaline saturated formations by EM observations with resistivity-derived formation factors was reasonably successful. Porewater conductivities estimated from 12 unique EM models from 8 sites were compared against directly measured porewater samples. A reasonable degree of correlation exists between the measured and predicted porewater conductivity for these 12 samples.

Forward models predict that the floating EM-34 VMD and HMD modes have potential for measuring useful information on seabed conductivity. Field experiments, however, were unsuccessful at reproducing these theoretical results. Given the space needed for EM-34 measurements, instrument development efforts targeting these settings may be better focused on short marine resistivity streamers.

The small size of the floating EM-31, relative to resistivity streamers or the space needed between EM-34 coils, proved useful for profiling in otherwise inaccessible terrain. Results from floating EM-31 VMD experiments suggest that conductivity readings interpreted with two-layer models that incorporate calibration information from pore and surface water measurements and DC soundings can be used in areas of extremely high conductivity porewaters near mangroves to predict porewater conductivity, which may be useful for near shore SGD studies and multi-disciplinary studies in wetlands. It is important to note that in such very high conductivity terrains as the TCSA, without resistivity surveys for calibration, inversions of EM data alone were
inherently ambiguous. There is still considerably utility in the EM methods, however, as they are faster than the DC methods and can be used in shallower water and less accessible terrain, such a mangrove shorelines.

No sites were available at the TCSA for directly examining potential conductivity effects of SGD. A prospective diffuse SGD anomaly located by resistivity methods in deeper water in Tampa Bay was used to assess the capabilities of the EM methods used in this study. EM response models predict the floating EM-31 lacks the necessary resolution to identify diffuse SGD. However, the overall success of predicting porewater salinity distribution within the TCSA suggests that the floating EM-31 method can delineate more concentrated zones of SGD with higher porewater conductivity contrast.

The process of mangrove soil salinization was found to significantly effect apparent conductivity readings within 5m of the mangrove trunk and falling sharply off within 10m at the edge of Tampa Bay. Restrictions in surface water flow and associated slower porewater flushing in some ponds and ditches were associated with higher conductivities in general and an extension of this effect to 50-125m from the nearest mangrove trunk.
References Cited


APPENDICES

The following appendices provide background on the methods used and the locations of measurements taken for this thesis. Appendices 11-13 show examples of data provided to the USGS Tampa Bay Integrated Science Project.
Appendix 1 - Method for Measuring Water Conductivity and Depth

**Instruments:** Yellow Springs Instruments YSI-30 conductivity and temperature meter (accuracy ± 21mS/m). Van Essen model DI-219 conductivity, temperature and depth data logger (accuracy ± 50 mS/m and ±3 cm depth). The less accurate DI-219 was periodically corrected with the YSI-30 during time series logging.

**Calibration:** Instrument calibrated to using KCl solution; 12.85 mS/cm ± 0.35% at 25°C. This standard is traceable to Standard Reference Material 3193 produced by the National Institute of Standards and Technology in Gaithersburg, MD USA.

**Depth Measurements:** Water depth was measured with a barometrically compensated Van Essen DE-219 for time series data or with a weighted measuring tape for individual data points.

**Porewater Sampling:** Porewater samples were collected in the field using a peristaltic pump and stored in Nalgene HDPE bottles. Samples were collected from the USGS TC1 multi-port well (Figure 2, Appendix 3), a percussion hammer drive point piezometer, 0.5 cm hand pushed stainless steel piezometer, single port wells and directly from core samples using a Manheim porewater hydraulic press.

**Filtering:** Particulate clay in porewater samples on the TCSA, especially when total dissolved solids are low, may introduce a conductivity error. The high cation exchange capacity of the clays found on the TCSA may increase porewater conductivity measured by the YSI-30 probe and similar devices (Hyde and Huckle, 1983; Caldwell et al., 1986; McNeill, 1990). Marine salinity (4000-5000 mS/m) water on the TCSA probably had a negligible clay conductivity effect because the charge of clay particles is reduced in high TDS waters (McNeill, 1990). The clay content of porewater samples varied based on the sampling method used and was independent of TDS measured after filtration; therefore, all porewater samples were centrifuged and decanted or passed through a 4 µm filter in order to remove any variances introduced by clay particulates.

**Salinity Calculation and Units:** Conversion between salinity and conductivity was computed using the International Equation of State (IES 80) method (Lewis and Perkin, 1978; Lewis, 1980; Fofonoff, 1985). The apparent conductivity that EM and DC methods measure is in large part a function of the porewater and surface water conductivity, which is a function of temperature; therefore, calculations involving pore and surface waters and EM and DC readings did not use specific conductance ($C_s$), which is referenced to a common temperature, but instead used absolute conductivity ($C$) which is a function of the temperature at the time of sampling (McNeill, 1990).
PCLOOP forward model of EM-34 response over a homogenous half-space with infinite depth (Geonics, 1994). VMD and HMD response slope is negative beyond a terrain conductivity of 600 mS/m and 2000 mS/m respectively. VMD and HMD 10, 20 and 40 m coil spacing response is identical and the difference between HMD coil spacing data is too small to plot on this graph. Note the limits for readings with the analog EM-34.
Appendix 3: EM-34 HMD 10 Meter Coil Spacing Raw Data
Appendix 4: EM-34 HMD 20 Meter Coil Spacing Raw Data
Appendix 5: EM-34 HMD 40 Meter Coil Spacing Raw Data
Appendix 6: EM-34 VMD Sounding Locations
Appendix 7: Raw EM-31 Soundings on Land and Over Water
Appendix 8: Two-Layer Resistivity Modeling Programs and RMS Error.

Apparent Conductivity Calculation for Schlumberger Array:
\[
\sigma_r = \frac{1}{\rho_r} \quad \rho_r = \frac{\pi l^2}{I} \frac{\Delta V}{2l} \quad \text{with } x = 0 \text{ and } L > 10l
\]

Apparent Conductivity Calculation for Wenner Array:
\[
\sigma_r = \frac{1}{\rho_r} \quad \rho_r = 2\pi l (\Delta V / I)
\]

Apparent resistivity \((\rho_r)\), apparent conductivity \((\sigma_r)\), voltage \((V)\), ampere \((I)\), potential electrode spacing \((l)\), current electrode spacing \((L)\), offset distance between the current electrode spread \((x)\), array spacing \((a)\) (Koefoed, 1979; Sharma, 1997).

Modeling Programs: The IX1D inverse and forward modeling program by Interpex, Ltd. was used to create the two-layer floating Schlumberger array models and fit the measured apparent conductivity to within 8.5% of the same models run in VES and DCEL as an error check (Cooper, 2000; Interpex, 2002; Weller, 2003). Land based Wenner traverse resistivity surveys were inverted for apparent conductivity using the two dimensional RES2DINV resistivity inversion program (Loke, 2002). The inversion process used in these programs assigns each sub-surface grid node an initial terrain conductivity and then calculates the apparent conductivity that would result and iteratively adjusts the model layers until the RMS error is minimized.

Definition of RMS (Root Mean Square) Model Error: RMS error is used as an indication of the fit between the theoretical data generated from the model and the measured data. RMS error is calculated by summing the squares of the difference in the log of the data values (apparent conductivity) and then dividing by the number of data points and taking the square root of the result. The antilog of this result minus one multiplied by 100 gives the percent RMS error. This method of calculating model error ensures that high data values do not dominate the calculated error and leave large errors in the low data values (Interpex, 2002).

References:


Appendix 9: Location of Wenner Array Lines and USGS TC1 Multi-Port Well

Note: Location of this area is also visible in Figure 5, Area 4.
Appendix 10: RES2DNV 2-D Wenner Array Resistivity Inversions

TC LINE 1

TC LINE 2

TC LINE 3

Unit electrode spacing 1.00 m.

Unit electrode spacing 3.00 m.

Unit electrode spacing 0.250 m.
Pond salinity, road bed location and distance with EM-34 HMD raw apparent conductivity data at different coil spacings.
Appendix 12: Local Influence of Mosquito Control Ditches.
Appendix 13: Local Influence of Mosquito Control Ditches with Elevation.