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Hydrogeophysical Characterization of Anisotropy in the Biscayne Aquifer Using Geophysical Methods

Albert Yeboah-Forson

Florida International University, ayebo001@fiu.edu

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FLORIDA INTERNATIONAL UNIVERSITY
Miami, Florida

HYDROGEOPHYSICAL CHARACTERIZATION OF ANISOTROPY IN THE
BISCAYNE AQUIFER USING GEOPHYSICAL METHODS

A dissertation submitted in partial fulfillment of the
requirements for the degree of
DOCTOR OF PHILOSOPHY
in
GEOSCIENCES
by
Albert Yeboah-Forson

2013

To: Dean Kenneth Furton
College of Arts and Sciences

This dissertation, written by Albert Yeboah-Forson, and entitled Hydrogeophysical Characterization of Anisotropy in the Biscayne Aquifer using Geophysical Methods, having been approved in respect to style and intellectual content, is referred to you for judgment.

We have read this dissertation and recommend that it be approved.

René M. Price

Oren V. Maxwell

Xavier Comas

Fernando Miralles-Wilhelm

Dean Whitman, Major Professor

Date of Defense: June 13, 2013

The dissertation of Albert Yeboah-Forson is approved.

Dean Kenneth Furton
College of Arts and Sciences

Dean Lakshmi N. Reddi
University Graduate School

Florida International University, 2013

DEDICATION

I would like to dedicate this work to my wife Ruby for all the sacrifice. Thank You.

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ABSTRACT OF THE DISSERTATION

HYDROGEOPHYSICAL CHARACTERIZATION OF ANISOTROPY IN THE

BISCAYNE AQUIFER USING GEOPHYSICAL METHODS

by

Albert Yeboah-Forson

Florida International University, 2013

Miami, Florida

Professor Dean Whitman, Major Professor

The anisotropy of the Biscayne Aquifer which serves as the source of potable water for Miami-Dade County was investigated by applying geophysical methods. Electrical resistivity imaging, self potential and ground penetration radar techniques were employed in both regional and site specific studies. In the regional study, electrical anisotropy and resistivity variation with depth were investigated with azimuthal square array measurements at 13 sites. The observed coefficient of electrical anisotropy ranged from 1.01 to 1.36. The general direction of measured anisotropy is uniform for most sites and trends W-E or SE-NW irrespective of depth. Measured electrical properties were used to estimate anisotropic component of the secondary porosity and hydraulic anisotropy which ranged from 1 to 11% and 1.18 to 2.83 respectively. 1-D sounding analysis was used to model the variation of formation resistivity with depth. Resistivities decreased from NW (close to the margins of the everglades) to SE on the shores of Biscayne Bay. Porosity calculated from Archie's law, ranged from 18 to 61% with higher values found along the ridge. Higher anisotropy, porosities and hydraulic conductivities were on the Atlantic Coastal Ridge and lower values at low lying areas west of the ridge.

The cause of higher anisotropy and porosity is attributed to higher dissolution rates of the oolitic facies of the Miami Formation composing the ridge. The direction of minimum resistivity from this study is similar to the predevelopment groundwater flow direction indicated in published modeling studies. Detailed investigations were carried out to evaluate higher anisotropy at West Perrine Park located on the ridge and Snapper Creek Municipal well field where the anisotropy trend changes with depth. The higher anisotropy is attributed to the presence of solution cavities oriented in the E-SE direction on the ridge. Similarly, the change in hydraulic anisotropy at the well field might be related to solution cavities, the surface canal and groundwater extraction wells.

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LIST OF SYMBOLS

ρ_0	Pore fluid resistivity
α	toutorsity factor
γ	specific weight of fluid
ϵ_r	dielectric permittivity
θ	direction of minimum resistivity
θ	minimum resistivity
λ	coefficient of anisotropy
μ	viscosity of the fluid
$\mu\text{S/cm}$	specific conductance
ρ_l	longitudinal resistivity
ρ_m	mean resistivity
ρ_t	transverse resistivity
Φ	anisotropic porosity
Φ	secondary porosity
$\Omega\text{-m}$	ohm-meter
ft	feet
ft^2/d	feet^2/day
ψ_h	Hydraulic anisotropy
mV	millivolt
mi	mile
Hz	hertz
MHz	Megahertz
GHz	Gigahertz
K	hydraulic conductivity
km	kilometer
m	cementation factor
m	meter
ρ_R	Rock resistivity
Z_e	The equivalent effective depth

LIST OF ABBREVIATIONS AND ACRONYMS

1-D	One Dimension
2-D	Two Dimension
3-D	Three Dimension
A	Square array lengths
ACR	Atlantic Coastal Ridge
ARS	Azimuthal Resistivity Survey
ASPG	Azimuthal Self Potential Gradient
BL	Bird Lake Park
CMP	common midpoint
CO	Camp Owaissa Bauer
Co	common offset
CP	Continental Park
DC	direct current
DF	Dante Fascell Park
EC	FIU Engineering Center
EM	electromagnetic
ERI	Electrical resistivity imaging
ERT	Electrical resistivity tomography
ESI	Imaginary extent of saltwater interface
GPR	Ground Penetration Radar
GPS	global positioning system
HEM	helicopter electromagnetic
MDWS	Miami Dade Water and Sewerage
ML	Miller Pond Park
MP	Modello Park
NAD	North American datum
NAVD	North American vertical datum
NT	North Trail Park
PL	Palmland Park
PP	Palmer Park
RMS	Root Mean Square
RMSE	Root Mean Square Error

SC	Snapper Creek Well Field
SFWMD	South Florida Water Management District
SL	Sun lake Park
SP	Self Potential
TEM	Time-domain electromagnetic
USGS	United States Geological Survey
VES	vertical electrical soundings
WL	Wild Lime Center Park
WP	West Perrine Park

1 INTRODUCTION

1.1 IMPORTANCE OF THE STUDY

Groundwater contributes a sizeable percentage of available potable water to the citizens of the United States. In the State of Florida, over 90% of water resource come from groundwater (Solley et al., 1988) and for Miami-Dade County the number is 100%. The need to protect this water resource from depletion and contamination is the forefront of every water organization objective. The management of groundwater requires a detailed knowledge of fundamental flow properties and process behavior in the subsurface. Unfortunately, the physics of flow in geological media like an aquifer is a dynamic science and deep insight into the behavior at the microscopic level is still evolving (Nutzman et al., 2005). Hence, progressive understanding of aquifer systems requires not just the traditional hydrogeology but integrated scientific approaches previously never considered for groundwater characterization and investigation.

Physical hydrological properties like anisotropy, hydraulic conductivity, porosity and pore space orientation are key to the characterization of groundwater. In an ideal scenario where unlimited resources exist, information about these parameters can be obtained from hydrogeological field surveys, but in the real world, decision makers and managers of groundwater resources have limited resources, finite time and cannot run unlimited hydrogeological tests. Hence, groundwater model tools are often used in the quest to predict, plan and protect this vital resource which makes up 28% of all available freshwater resources (Jackson et al., 2001).

The development of accurate groundwater models relies heavily on available field data to constrain the models. However, the lack of adequate field data often leads to inaccurate representation of the sub-surface in geological modeling as a consequence of oversimplification of parameters. In carbonate aquifer systems like the Biscayne aquifer of South Florida which exhibit karst features (Cunningham et al., 2006), these problems are magnified because of the extremely high transmissivity ($1,000,000 \text{ ft}^2/\text{d}$ or greater) in the aquifer (Fish and Stewart, 1991) which limits drawdown from pump tests (Merritt, 1997) and makes field estimation of anisotropy difficult. In addition, dissolution features inherent in the rocks of south Florida have caused the porosities in the aquifer to change rapidly from point to point and even changes with depth at the same location. As a result extrapolating measurement of porosity from lab to the aquifer or from well survey to the aquifer might fail to capture the true extent of the aquifer. Another challenge facing researchers in South Florida is that the groundwater flow behavior in the aquifer is not natural and is highly influenced by artificial canal systems and extraction wells which control the local changes in an aquifer. Finally, the Biscayne aquifer like any coastal aquifer interacts with sea water which often leads to salt water intrusion. Hence the use of alternative pre-investigation tools like geophysical survey is highly employed to circumvent some of these critical factors.

Application of geophysical methods to hydrogeological problems has gained more ground recently in the scientific community leading to a research sub-field of hydrogeophysics (Yoram and Hubbard, 2005). Hydrogeophysics is the concept of using geophysical methods to investigate hydrogeological processes or parameters by providing

quantitative information about the subsurface. The attractiveness of hydrogeophysical methods like geoelectrical techniques for characterizing aquifer properties results from the noninvasive nature of the techniques, cost effectiveness, fast data acquisition and ability to map both geological layers and groundwater table when dealing with large scale field surveys (Fitterman et al., 2012; Hinnell et al., 2010; Huisman et al., 2010).

Although Archie's law which relates the electrical measurements of saturated rocks to porosity, water saturation and pore fluid conductivity in-situ have been in existence for more than half a century (Archie, 1942); the relationship between electrical properties and hydrological parameters is still an evolving topic. Many recent studies including Sandberg (2002); Slater (2007); Looms et al. (2007) and Margiotta et al. (2008) have demonstrated that an integrated approach of geophysical measurement for petrophysical properties is the way forward. Such integrated geophysical approaches have shown within reasonable accuracy that electrical properties like azimuthal resistivity can be used to predict the pore characteristic of carbonate systems (Hart and Rudman, 1997) as well as using electrical signal to predict transmissivity during pumping (Ritzi et al., 1992). The aforementioned success of these studies indicates that geoelectrical methods could be successfully employed in the Biscayne Aquifer to predict hydrological parameters.

Geophysical techniques such as electrical resistivity tomography (ERT), ground penetration radar (GPR), Airborne and Marine electromagnetic (EM) are used to routinely characterize the hydrological properties of surficial aquifer systems. In the Biscayne Aquifer this approach has been used as means to estimate the rate of salt-water

intrusion and validate either geochemical or hydrological studies (Fitterman and Stewart (1986); Fitterman and Discs-Pan, 1998; Swarzenski, 2004 and Renken et al., 2005). However, only a few studies have ever used geophysical methods to quantitatively assess the physical properties of the aquifer such as porosity (Grasmueck and Weger 2002; Cunningham, 2004; Neal et al. 2008).

1.2 REVIEW OF THE BISCAYNE AQUIFER

The Biscayne Aquifer is a Pleistocene unconfined carbonate aquifer located in southeast Florida and is the principal source of water for all of Miami-Dade, Broward and Monroe Counties (Miller, 1990). A detailed discussion of the geology and the hydrogeology of the Biscayne Aquifer can be found in studies like Parker et al. (1955), Fish and Stewart (1991), Cunningham et al. (2004, 2006, 2009), Renken et al. (2005) and Cunningham and Florea (2009). Geologically, the Biscayne aquifer is made up of several Pleistocene age surficial geologic units (Miami Limestone, Key Largo Limestone, Fort Thompson formation) but Pliocene and late Miocene rocks are also contained in the aquifer. The aquifer is underlain by the Tamiami formation except in the north where it consists of a transition zone that varies from Anastasia to Tamiami formation. The water table is at or near the land surface in many places, except at the Atlantic Coastal Ridge along the east coast which has a approximately 6m of unsaturated zone.

The rocks are very porous with estimated porosity ranging from 5.5 to 79% (Fish and Stewart 1991; Cunningham et al, 2004; Manda and Gross, 2006; Cunningham et al. 2006). The porosity for the Biscayne Aquifer has been shown to vary depending on the

depth (Cunningham, 2004) and location (Renken et al., 2005). Cunningham et al. (2006), classified porosity into three groups: (1) touching-vug porosity (high permeability - conduit flow); (2) interparticle matrix and separate-vug porosity (moderate permeability-diffuse-carbonate flow); and (3) conduit porosity (low permeability-fracture flow). The water table of the Biscayne Aquifer fluctuates rapidly in response to variations in recharge (precipitation), natural discharge (seepage into streams, canals, or the ocean; evaporation; and transpiration by plants), and pumping from wells. Water levels are higher on the west of the aquifer toward the Everglades and lowest along the Atlantic Coastal Ridge.

1.3 REVIEW OF GEOPHYSICAL METHODS USED IN THIS STUDY

There are two main classes of geophysical methods: active and passive methods. Active methods (e.g., electrical resistivity and GPR) send signals to the subsurface and measure the response while passive methods (e.g., self potential) measure the natural response to the subsurface. In this study both approaches were employed. The variation of induced or natural electrical current within surrounding geological unit provides information about the nature of the geological structure below the subsurface and is typified in the geo-electric section. A short overview of the geophysical methods used in this thesis is presented here.

1.3.1 Electrical Resistivity

Electrical resistivity (ρ) an inverse of electrical conductivity, is an inherent property of all earth materials and is defined as a measurement of material resistance to the flow of electrical current (Fretwell and Stuart, 1981). Electrical current may be propagated into the subsurface through conductive, electrolytic or dielectric conduction. Most rocks are poor conductors and generally conduct electricity due to the electrolytic conduction of pore fluid interconnection (Keller and Frscknecht, 1982). Resistivity of porous rocks such as limestone varies with porosity and pore volume when the rock is saturated (Archie, 1942). Resistivity surveys are usually carried out using the direct current (DC) method by inducing current into the ground via two current electrodes and measuring the potential difference across the subsurface through two additional potential electrodes. The apparent resistivity of the subsurface is proportional to the ratio of voltage to current. A detailed review of all the various resistivity imaging and sounding techniques to provide one dimensional, two dimensional or three dimensional images of subsurface electrical structure is available in Loke et al. (2013). Application of resistivity in aquifer studies includes determination of preferential flow directions, cavities, saltwater intrusion and depth analysis of surficial aquifers (Ritzi et al., 1992; Hart and Rudman, 1997; Slater, 2007 Loke et al. 2013).

1.3.2 Ground penetrating radar

Ground penetrating radar (GPR) is an electromagnetic geophysical technique for subsurface exploration that uses a transmitter to generate electromagnetic (EM) waves

(typically in a range between 10 MHz - 2 GHz) that travel the subsurface and return to a receiver as a sequence of reflections. Reflections result from contrast in dielectric permittivity (ϵ_r), a physical property highly dependent on water content. The contrast between limestone matrix and dissolution features (whether filled with water or air) represents a good target for GPR detection. The use of GPR has proven to be well-suited for detailed description of the epikarst (<20m for saturated zone and <30m for unsaturated zone of karst system Vadillo et al., 2012), the infiltration zone of the karst aquifer where limestone is prevalent (Al-fares et al., 2002) and cavity location in karst aquifers (Annan, 2005). For example, Cunningham (2004) used GPR to characterize hydrogeologic properties to understand the paleokarst in the Biscayne aquifer. Likewise, Grasmueck and Weger (2002) and Neal et al. (2008) employed 2-D and 3-D GPR method in many locations around Miami, Florida, to understand the radar stratigraphy of Pleistocene Miami Limestone.

1.3.3 Self Potential (SP)

Natural potential differences at the surface arise from various sources including fluid streaming bioelectric activity in vegetation and electrochemical concentration in ground water causing current flow in the subsurface. The SP method passively measures natural potential difference (no active injection of current) which generally exists between two points on the ground using two non-polarizing electrodes (Telford et al., 1990). Mechanisms such as electro-filtration, diffusion and membrane potential, mineral potentials, adsorption, and bioelectric potentials producing SP have not been

satisfactorily answered (Bolève et al., 2007). Self Potential is the only geophysical method with the capability to determine flux independent of any knowledge of available fracture orientation (Revil, 2002). The electrical field associated with the flow of the ground water is called the streaming potential and it indicates the direction of flow (Revil, 2002; Jardani et al., 2006; Wishart et al., 2006). Wishart et al. (2008) showed that it is possible to determine preferential directions of groundwater flow associated with induced anisotropy from SP measurement. Application of self potential methods over the last decades has lead to strong understanding of the behavior of groundwater streaming potentials in the subsurface (Revil, 2002; Rizzo et al., 2004; Jardani et al., 2006).

1.4 OBJECTIVE OF THE STUDY

The goal of the dissertation is to understand the petrophysical relationship between hydrogeological parameters and the geo-electric properties in the Biscayne Aquifer. The objective of this study will be accomplished by the use of geo-electrical methods and hydrogeological data to quantitatively determine hydrogeological parameters such as hydraulic anisotropy, porosity and pore orientation. In addition, the study also compares geo-electrically derived hydrogeological parameters with published studies of the Biscayne Aquifer.

These goals were addressed through the following specific objectives:

1. To determine the anisotropy of the Biscayne Aquifer in Eastern Miami-Dade County and establish if the orientation of the anisotropy of the aquifer is unidirectional.
2. To image the depth, thickness, and extent of layered conductivity and porosity in the saturated zones of the Biscayne Aquifer.
3. To characterize dissolution features in the Miami Limestone and understand their anisotropic behavior at a site on the Atlantic Coastal Ridge.
4. To investigate the cause(s) of the variation in flow direction of electrical anisotropy and to examine the role of groundwater extraction wells or canals, if any, on the observed trends in anisotropy direction with depth at Snapper Creek Municipal Well.

1.5 THE STRUCTURE OF THIS DISSERTATION

The dissertation is organized into six chapters. Motivation, literature review, study region and key questions to be addressed are contained in Chapter one as introduction while the conclusions and contributions of the study are presented in Chapter six. Chapters two, three, four and five, present independent studies that connect the ability of geophysical methods to provide information about aquifer properties. Each chapter addresses one of the four objectives highlighted. These chapters were written as one standalone papers and can be read individually.

Chapter two describes a regional study using a 28-electrode resistivity imaging system to investigate electrical anisotropy at 13 sites in the Biscayne Aquifer of SE Florida. The electrical parameters measured from azimuthal square array methods were used to estimate secondary porosity and the principal components and direction of the hydraulic conductivity tensor. The geographic patterns of the electrical anisotropy are characterized in both magnitude and direction. The geoelectrical estimation of these hydraulic properties can provide an alternative preliminary assessment tool for aquifer characterization. The results contained in this chapter have been submitted to the journal, *Groundwater* and are currently under review.

Chapter three addresses the specific objectives of using electrical resistivity sounding and associated 1-D resistivity model tools to determine the depth, thickness, and extent of porosity layered in the study area. Using a combined resistivity and hydrogeological methods, 1-D depth stratification of eastern Miami-Dade is generated for the shallow surficial aquifer unit. Iso-resistivity and porosity maps are created for different depths. The key result is that porosities are higher along the Atlantic Coastal Ridge compared to the areas west of the ridge and variations in resistivity with depth are mostly influenced by dissolution features and saltwater intrusion. The results in chapter three are intended for submission to the journal, *Environmental Earth Sciences*.

Chapter four integrates the use of electrical resistivity and ground penetration radar methods to image, characterize and delineate dissolution features at the West Perrine Park located on the Atlantic Coastal Ridge and to better understand the effect of these features on anisotropy. Quantitative analysis of the geophysical data and high-resolution borehole images from wells close to the site indicate potential solution cavities

in the study site. The presence of cavities, especially if along the preferential direction might cause an increase in the hydraulic anisotropy which could alter the magnitude of ground-water flow. A revised version of this chapter will be submitted to the Journal of Hydrogeology.

Chapter five is a case study that attempts to understand the factors that cause anisotropy to change with depth in a well field. Multiple resistivity data, azimuthal self potential gradient and well data are used to assess anisotropic change at the Snapper Creek Municipal Well Field. The SP data was collected during pumping and non-pumping periods to assess the directionality of flow. The study speculates that the change in electrical anisotropy direction with depth is the result of the complex relationship between the permeable vuggy and moldic rocks, surface water flow in the canal and groundwater behavior from the extraction wells at the site. An extract of this chapter will be submitted to Journal of Hydrology.

1.6 ORIGINAL CONTRIBUTION OF THE STUDY

The major contributions of the dissertation can be found in Chapters two, three, four and five, which discuss the characterization of anisotropy, porosity and hydraulic conductivity in the Biscayne Aquifer using geophysical approaches. The use of azimuthal resistivity technique to estimate hydraulic anisotropy and secondary porosities in the surficial carbonate aquifer system is an original concept. Secondly, although electromagnetic methods have been used for depth analysis of the Biscayne Aquifer, those studies focused on seawater intrusion; the azimuthal resistivity soundings used in

this study resolve depth stratification for hydrogeologic parameters analysis. In addition to imaging dissolution features in the aquifer, the study aims to understand the impact of cavities on anisotropy. Finally, the study shows the likely potential impact canals and extractions wells might have on the anisotropy in a groundwater-surface water interaction zone in urbanized areas.

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2 ELECTRICAL RESISTIVITY CHARACTERIZATION OF ANISOTROPY IN THE BISCAYNE AQUIFER

ABSTRACT

Electrical anisotropy can play an important role in estimating the orientation of hydraulic conductivity in aquifers. Knowledge of hydraulic anisotropy can aid modeling of groundwater flow as anisotropy in rocks of aquifer may indicate a preferred hydraulic conductivity direction, which has been formed through fracturing, sedimentary fabric or dissolution. In the present study, a 28-electrode resistivity imaging system was used to investigate electrical anisotropy at 13 sites in the Biscayne Aquifer of SE Florida using the rotated square array method. The measured coefficient of electrical anisotropy generally ranged from 1.01 to 1.12 with values as high as 1.36 found at one site. Higher values were generally located on the Atlantic Coastal Ridge while the lowest values were in low elevation areas on the margin of the Everglades to the west. The observed electrical anisotropy was used to estimate an anisotropic component of the secondary porosity and hydraulic anisotropy (ratio of maximum to minimum hydraulic conductivity) which ranged from 1 to 11% and 1.18 to 2.83 respectively. The predominate trend of minimum resistivity and maximum hydraulic conductivity was E-W/SE-NW beneath the ridge and E-W/SE-NW farther west. The high anisotropic values found on the ridge may be a result of increased dissolution rates of the oolitic facies of the Miami Formation limestone compared with the bryozoan facies to the west. The anisotropy directions are similar to the predevelopment of groundwater flow direction as

indicated in studies published by others. The finding suggests that, the paleo-groundwater flow in the Biscayne Aquifer might have resulted in the observed anisotropy.

2.1 INTRODUCTION

The Biscayne Aquifer of SE Florida is a carbonate aquifer which exhibits karst features and secondary porosity because of the presence of touching vugs, conduits and solution holes (Cunningham et al., 2006). Dissolution can alter the direction and velocity of ground-water flow resulting in unpredictable flow characteristics (Knochenmus and Robinson, 1996). A major challenge facing groundwater modeling in carbonate aquifer systems is that aquifers are often assumed to be isotropic though field measurements suggest that these aquifers are anisotropic (Neuman, 1975). Aquifer anisotropy occurs when hydraulic conductivity varies with direction. On a borehole scale, hydraulic anisotropy is usually determined by using direct methods such as core analysis or pump recovery tests which determines the principal components and orientation of the transmissivity tensor from drawdown data (Neuman, 1975; Ritzi and Andolsek, 1992). On a regional scale, these techniques are limited by factors such as high cost, sparseness of hydrologic data, and field scale heterogeneity (Rubin and Hubbard, 2005). In the Biscayne Aquifer of SE Florida, there is a general lack of published studies on anisotropy. The extremely high transmissivity ($1,000,000 \text{ ft}^2/\text{d}$ or greater) in the aquifer (Fish and Stewart, 1991) limits the drawdown from pump tests (Merritt, 1997) which makes field estimates of anisotropy difficult. Even in the less transmissive Floridan Aquifer in central Florida (transmissivity of $30,000\text{-}57,000 \text{ ft}^2/\text{d}$), attempts at measuring

hydraulic anisotropy from drawdown tests have given inconsistent results (Knochenmus and Robinson, 1996). These factors make direct measurements of anisotropy in the Biscayne aquifer difficult to achieve.

Indirect geophysical methods such as direct current (D.C) resistivity provide an inexpensive and noninvasive alternative for estimating anisotropy and other hydrologic parameters (Slater, 2007; Loke et al., 2013). Groundwater and electrical current flow are similar as both are transported through the interconnected pore volumes (Revil and Cathles, 1999). Electrical anisotropy occurs in the subsurface when electric current flows differently in one horizontal direction relative to the other and can occur in the presence of steeply dipping beds or fractures (Taylor and Fleming, 1988; Lane et al., 1995; Boadu et al., 2005; and Ramanujam et al., 2006). Therefore, electrical anisotropy measurements can provide a useful analog for estimating a preferred hydraulic conductivity direction in the aquifer.

In this study, a 28-electrode resistivity imaging system was used to investigate electrical anisotropy at several sites in the Biscayne Aquifer of SE Florida using the square array method. The geographic patterns of the electrical anisotropy are characterized in both magnitude and direction. These parameters are used to estimate hydraulic characteristics such as secondary porosity and the principal components of the hydraulic conductivity tensor. Successful geoelectrical estimation of these hydraulic properties can provide an alternative preliminary assessment tool for aquifer characterization.

2.2 GEOLOGY AND HYDROGEOLOGY OF THE BISCAYNE AQUIFER

The Biscayne Aquifer is a Pleistocene unconfined carbonate aquifer located in southeast Florida. It underlies an area of approximately 10,000 km² and is the principal source of water for all of Miami-Dade, Broward and Monroe Counties (Miller, 1990). The surface elevations increase from the low lying areas of the Everglades in the west to the Atlantic Coastal Ridge along the east coast with elevations reaching 6 m above sea level (Figure 2.1). A detailed discussion of the geology and the hydrogeology of the Biscayne Aquifer can be found in studies of Parker et al. (1955), Fish and Stewart (1991), Cunningham et al. (2006, 2009), Renken et al. (2008) and Cunningham and Florea (2009).

The rocks of the aquifer are composed of the Miami Limestone, the Key Largo Limestone, and the Fort Thompson Formation (Fish and Stewart 1991). The Miami Limestone, formed during a sea level high stand associated with the Sangamon Interglacial, is the predominant unit found at the surface and is subdivided into oolitic and the bryozoan facies (Hoffmeister et al., 1967). The oolitic facies is present beneath the Atlantic Coastal Ridge and can be grouped into cross-bedded and bioturbated facies. The bryozoan facies is found underneath the Everglades and consists of sandy fossiliferous rocks which were formed in lagoonal environments west of the Atlantic Coastal Ridge. The Fort Thompson Formation is made up of intercalated fresh and marine limestone and underlies the Miami Formation in Miami Dade County (Fish and Stewart, 1991).

Porosities in the aquifer range from 5.5 to 79% with most sites showing a maximum porosity around 40% (Fish and Stewart 1991; Cunningham, 2004; Manda and

Gross, 2006; Cunningham et al., 2006). The porosity for the Biscayne Aquifer has been shown to vary depending on the depth (Cunningham, 2004) and the location of observation (Renken et al., 2005) as a consequence of the complex nature of the carbonate rocks. Hence the total or effective porosity for the overall Biscayne Aquifer is rarely estimated since assigning a single porosity value would never be a true representative of the actual porosity. Cunningham et al. (2006) classified porosity into three groups: (1) touching-vug porosity (high permeability - conduit flow) (2) interparticle matrix and separate-vug porosity (moderate permeability- diffuse-carbonate flow); and (3) conduit porosity (low permeability- fracture flow). Both horizontal and vertical permeability increase with increases in porosity. Higher hydraulic conductivities in excess of 3.5 cm/s (10000 ft/day) are common for this aquifer owing to the well-developed secondary porosity. In Miami Dade County, the secondary porosity is greatest in the Fort Thompson Formation and as a result, most of the production wells in the Biscayne aquifer are screened in the Fort Thompson (Fish and Stewart, 1991).

2.3 ELECTRICAL ANISOTROPY

Electrical anisotropy occurs in the subsurface as a result of current flowing differently in one direction relative to another. The phenomenon, originally described by Maillet (1947), is usually the result of planar beds or fractures within the rocks providing lower electrical resistance than the surrounding media. Consider a geological model with electrical current flowing through a preexisting vertical rock fabric oriented at a strike of θ (Figure 2.2). The resistivity across the bedding plane (transverse resistivity, ρ_t) is

normally greater than the resistivity parallel to bedding (longitudinal resistivity, ρ_l). The anisotropic media is characterized by the coefficient of anisotropy, λ , and the mean resistivity, ρ_m ,

$$\lambda = \sqrt{\frac{\rho_t}{\rho_l}} \quad (2.1)$$

$$\rho_m = \sqrt{\rho_l \rho_t} \quad (2.2)$$

The coefficient of anisotropy, λ , for a homogeneous anisotropic geological unit is always greater than 1 since resistivity is always greatest in the transverse direction.

Electrical anisotropy is often measured by deploying a linear DC resistivity sounding array along a range of directions and plotting the measured apparent resistivity as a function of azimuth to define the anisotropy ellipse. An alternative to the linear array is the square array where the current and potential electrodes are deployed on opposite sides of a square (Figure 2.3). The square array has several advantages over the linear array, including faster set up time, smaller area requirement and greater sensitivity to anisotropy (Habberjam and Watkins, 1967; Habberjam, 1972; Habberjam, 1975). As the array is rotated through a series of angles, θ_0 , the apparent resistivity, ρ_a , varies in an elliptical-like pattern given by Habberjam (1972) as,

$$\rho_a = \frac{1}{2 - \sqrt{2}} \rho_m \left\{ \frac{\frac{2}{[1 + (\lambda^2 - 1)\cos^2(\theta_0 - \theta)]^{1/2}} - \frac{1}{[2 + (\lambda^2 - 1)(1 + \sin^2(\theta_0 - \theta))]^{1/2}}}{\frac{1}{[2 + (\lambda^2 - 1)(1 - \sin^2(\theta_0 - \theta))]^{1/2}}} \right\} \quad (2.3)$$

where θ is the minimum (longitudinal) resistivity direction.

Numerous studies have used measurements of electrical anisotropy to characterize steeply dipping fractures and beds in the sub-surface (Taylor and Fleming, 1988; Lane et al., 1995; Boadu et al., 2005; Ramanujam et al., 2006). In general, rocks in the Biscayne Aquifer are neither fractured nor well bedded. Instead, the rocks are composed of limestone in which dissolution processes have resulted in a well developed secondary porosity. Therefore electrical anisotropy observations may indicate a directional component of the secondary porosity corresponding to a dominant horizontal hydraulic conductivity direction.

2.4 DATA AND METHOD OF ANALYSIS

Data for this study were collected over a three year period from December 2008 to January 2012 at 15 sites in eastern Miami Dade County (Figure 2.1). Site locations ranged from low elevation regions near the Everglades in the west to the higher elevations of the Atlantic Coastal Ridge in the east. As a results of space limitations, the sites were restricted to public parks and other areas where open space is accessible. Site locations required natural undisturbed areas, open space and absence of artificially buried materials and utilities. Survey locations included sites (1) near the shoreline of Biscayne Bay where seawater has intruded into the aquifer; (2) on higher elevation regions of the Atlantic Coastal Ridge where the unsaturated zone is relatively thick; and (3) in lower elevation regions to the west, close to the Everglades where the unsaturated zone is relatively thin. Two of the sites were near the artificially dredged canals in the county.

These canals are situated in natural topographic lows known as transverse glades and were constructed in the early 20th century for flood control purposes.

The field measurements were made with an Advance Geoscience Incorporated (AGI) Super-Sting R1/IP 28-electrode resistivity imaging system. The system utilizes a multi-core cable controlled by a programmable switch box which controls the channels of the current and potential electrodes for each measurement. The 28 electrodes were placed at equal angles on a circle forming 7 separate square array configurations rotated at 12.86° intervals (Figure 2.4). The array was oriented to magnetic north with a Brunton compass and results were corrected to true north using the local declination of 6° . A custom command file was created which recorded 28 separate measurements of the square array rotated through 360° .

The 28 electrodes were initially deployed at a radius of 2.83 m and after each set of azimuthal measurements, the radius of the array was expanded in increments of $\sqrt{2}$. The array resulted in sets of azimuthal measurements for square array lengths, A , of 4.0, 5.7, 8, 11.3, 16, 22.6, 32 and 45.3 m. The procedure produced a set of anisotropic soundings which provided a measure of variation of anisotropy with depth. The equivalent effective depth, Z_e , for each azimuthal measurement is approximately one half the square size, A (Habberjam and Watkins, 1967; Edwards, 1977). While the exact relationship between the measurements and depth require numerical modeling, the effective depth provides a useful approximation for estimating the depth of the anisotropy.

The field data for each array radius were inverted for the three parameters: mean resistivity, ρ_m , coefficient of anisotropy, λ , and direction of minimum resistivity, θ , in

Equation 2.3 using an iterative non-linear least square approach .The iterative procedure started with an initial guess for the three parameters which formed the data vector and the kernel matrix through the partial derivatives of the parameters. The matrix was then inverted for the perturbation vectors and new parameters were estimated. This procedure was repeated until the parameters converged to better than 0.1%.

Like all non-linear inverse models, the method is dependent on accurate starting values to preclude convergence to a local minimum and to provide quick convergence of the parameters. Initial guesses for ρ_m , λ and θ were determined from the average resistivity measurement, the ratio of maximum to minimum resistivity, and the minimum resistivity square angle respectively. In some cases where this did not provide convergence, a parameter search algorithm was used to generate starting values. In addition to the inverted parameters, the procedure estimated statistical uncertainties associated with each parameter from the diagonal elements of the covariance matrix scaled by the root mean square error between observed data and the calculated resistivities. The off diagonal elements of the covariance matrix were small indicating little or no correlation between the parameters. The statistical significances of the inverted parameters were tested by applying the parametric T-test at a 95% confidence interval.

2.5 RESULTS

In total, 100 sets of azimuthal measurements were made at the 13 study sites. The data were modeled and plotted using the data analysis approach described in the previous

section. Statistically insignificant model fits resulted in the elimination of seven additional azimuthal measurement sets. The eliminated data set failed the T-test at a 95% confidence interval and have R^2 of less than 0.1.

Examples of typical survey results are shown in Figure 2.5. Figure 2.5a is representative of measurements with a low coefficient of anisotropy where the azimuthal pattern is nearly circular. As the anisotropy increases, the azimuthal pattern becomes more elliptical with a well-resolved orientation (Figure 2.5b). This pattern is representative of most of the measurements in this study. At sites with very high coefficients of anisotropy, the pattern defined by Equation 2.3 deviates from a pure ellipse (Figure 2.5c). This type of pattern was observed at only one study site, the West Perrine Park. Figure 5d is an example of a measurement having two minimum resistivity directions. Taylor and Fleming (1988) described such a scenario as 2 distinct directions of fracture orientation and in the absence of a defined fracture the true minimum resistivity direction obtained from Equation 2.3, is the average of the two unique directions. Approximately 5% of the measurements exhibit these characteristics.

The coefficient of anisotropy generally ranged from 1.01 to 1.11 with a mean of 1.06 and a mode of 1.03 (Figure 2.6). At one site, West Perrine Park (WP, Figure 2.1) anisotropy was as high as 1.36. Most of the low magnitudes of anisotropy were associated with smaller square sizes and were mostly found at low elevation sites. However, the depth at which they occurred varied from site to site. The coefficients of anisotropy higher than the mean were generally found at sites along the Atlantic Coastal Ridge where the unsaturated zone is thick and usually at square sizes of 11.3-23 m (effective depth, Z_e , of 5-11 m).

The direction of minimum resistivity, θ , shows that anisotropy on the regional scale, exhibits a range of different orientations (Figure 2.7). The predominant direction of the minimum resistivity generally trends in the E-W direction. Smaller populations trending in the NE-SW and NW-SE directions were also observed. The rose diagram also indicates a minor peak of minimum resistivity in the SSE direction.

To illustrate how anisotropy varies with depth, maps of the coefficient of anisotropy and the minimum resistivity direction for all the sites were created for the different square sizes used in the study (Figure 2.8). In general, the coefficient of anisotropy was marginally higher for sites on the Atlantic Coastal Ridge compared to those close to the Everglades. For example, at a square size of 11.3m (Z_e , 5.6 m), the magnitude of anisotropy for sites on the ridge ranges from 1.07- 1.08 while those behind the ridge were 1.03. This is also the case for other square sizes in this study. The maps, in general with few exceptions, also showed that the minimum resistivity on the Atlantic Coastal Ridge trends mostly in the E or SE direction, regardless of the square size. Directions for sites close to the Everglades are more variable for different square sizes. For instance at square size of 11.3m (Z_e , 5.6m) some sites trend NE while others trend SE or SSE.

Three of the sites (BL, PL, SC, Figure 2.8) exhibit directions of electrical anisotropy that vary with depth. For example at the Snapper Creek Well Field site (SC, Figure 8), for square sizes of 4-8 m (Z_e , 2-4 m), the minimum resistivity trends in the SE direction whereas for square sizes greater than 11m ($Z_e \geq 5$ m), it trends in the NE direction. In contrast to the abrupt change in the minimum resistivity direction with depth at SC, the change with depth at BL and PL was gradual ranging from SE to NE and NE to

SE respectively. Further analysis of these variations of anisotropy with depth will require detailed modeling but this analysis is beyond the scope of this paper.

2.5.1 Analysis of Secondary Porosity

In karst aquifer systems, secondary porosity dominates the overall porosity as a consequence of the presence of fractures, solution conduits and caves (Milanovic, 1981). In general, the rocks of the Biscayne Aquifer are not fractured and analysis of induction logs and core logs have shown that any secondary porosity is generally because dissolution of the rocks (Renken et al., 2008). This dissolution process has resulted in well developed secondary pores that enhance groundwater storage and transmission through the pore network in the form of isolated and interconnected vugs (Cunningham et al., 2006). The coefficient of electrical anisotropy has been shown to have the same functional form as anisotropic permeability to the first order (Bespalov et al., 2002) and higher coefficient of anisotropy implies higher anisotropic permeability (Boadu et al., 2005). Hence, an increase in the secondary porosity would mean more connectivity in the isotropic pores of the rock.

Taylor and Flemming (1988) developed a method for determining the additional porosity of fluid filled vertical joints using electrical anisotropy measurements by extending the basic analog of resistors connected in parallel and in series to a jointed rock unit. The effective resistivity in the transverse and longitudinal directions is dependent on the material resistivities (rock, ρ_R , and pore fluid, ρ_0) and anisotropic porosity, Φ ,

because of the fractures. The current flowing perpendicular to the fractures is analogous to resistors in series with the transverse resistivity ρ_t given by

$$\rho_t = \rho_0 \Phi + (1 - \Phi) \rho_R \quad (2.4)$$

while the current flowing parallel to the fractures is analogous to the resistors in parallel with the effective longitudinal resistivity,

$$\frac{1}{\rho_l} = \frac{\Phi}{\rho_0} + \frac{(1 - \Phi)}{\rho_R}. \quad (2.5)$$

In the present study, we extend this concept to an idealized homogeneous anisotropic rock characterized by linear conduits. The linear conduits act as resistors in series for current flowing perpendicular to the conduits (equation 2.4), and as resistors in parallel for current flowing parallel to the conduits (equation 2.5). The effective transverse resistivity is the sum of the resistivity perpendicular to the interconnected conduits and the surrounding saturated rock matrix of resistivity, ρ_R , weighted by their volume percentages while the effective longitudinal resistivity is in the direction parallel to the conduit. These equations may be combined with equations 2.1 and 2.2 to solve for the secondary porosity, Φ , as a function of mean resistivity, ρ_m , the coefficient of anisotropy λ and a known fluid resistivity, ρ_0 :

$$\Phi = \frac{\rho_0 \rho_m (\lambda^2 - 1)}{\lambda \rho_0^2 + \lambda \rho_m^2 - 2 \rho_m \rho_0} \quad (2.6)$$

Taylor and Flemming (1988) applied their method to fractured terrains where the rocks are largely nonconductive and the secondary porosity is solely because of the fracture sets. In carbonate systems like the Biscayne Aquifer, dissolution processes can create secondary porosity which is largely isotropic in nature. Hence, the secondary

porosity derived in this study from the electrical measurements is an anisotropic component of the secondary porosity which has a preferred direction and which should be parallel to and contribute to the direction of maximum hydraulic conductivity.

The anisotropic secondary porosity, Φ , is a factor of the coefficient of anisotropy, λ , and the mean resistivity, ρ_m , inverted from Equation 2.3 as well as the groundwater resistivity, ρ_0 . A groundwater resistivity of 16.7 Ω -m (specific conductance of 600 μ S/cm) (Fish and Stewart, 1991) was assumed for all sites in the study. While three of the sites (MC, DF and PL, Figure 2.1) were near coastline where the groundwater is affected by saltwater intrusion, the measured mean resistivity, ρ_m , was in all cases greater than 100 Ω -m which suggests that these measurements only sampled the freshwater portions of the aquifer. Anisotropic secondary porosities were calculated only for the saturated zone for each site by excluding the square sizes with measurements of mean resistivity, ρ_m , higher than 150 Ω -m

The distribution of anisotropic secondary porosity estimated from Equation 6 ranged from 0 to 11% (Figure 2.9). The calculated porosities are highly skewed towards low values with most values below 3%. The low values, mostly, correspond to the coastal and western sites. Higher values of porosity ($\geq 5\%$) are confined to the center of the ridge. These estimates are accurate to better than 0.2% based on the error estimates of the inverted parameters in (Equation 2.6). Average anisotropic secondary porosity was determined at each field site from the arithmetic mean of the porosities for each square size (Table 2.1; Figure 2.10a). The estimated mean anisotropic secondary porosity for each site ranged from 0.8% to 5.5%. The highest observed anisotropic porosities are

found in the center of the ridge (3-5.5%), with smaller values found near the coastline and towards the west.

2.5.2 Hydraulic anisotropy

Porosity unlike hydraulic conductivity is a scalar property. Hence the electrical anisotropy results are best described by directional variations in hydraulic conductivity (K) tensor. Although there are no direct relationships between electrical measurements and hydraulic conductivity properties, Slater (2007) listed a number of studies where petrophysical relationships have been shown to exist between electrical resistivity and hydraulic conductivity on a local scale. The reason is groundwater flow and electric current have obvious analogies (Ohms law to Darcy law) but are governed by different physics of flow principles (Ahmed et al., 1988). Hence, it is evident that K could be related to electrical properties where information about the local rock properties is obtained from laboratory measurement or aquifer studies. This approach was employed by Niwas et al. (2011) and Soupious et al. (2007) to accurately estimate aquifer hydraulic conductivity and porosity from surface geoelectrical measurements.

Koudadaki et al. (2007) combined the empirical relationships between permeability, porosity and electrical formation factor of Archie (1942) to obtain a relationship for estimating hydraulic conductivity, K , from electrical resistivity measurements,

$$K = a_2 * \frac{\gamma}{\mu} \left(\alpha \frac{\rho_0}{\rho} \right)^{(b_2/m)} \quad (2.7)$$

where m , α , a_2 and b_2 are empirically derived constants which depend on rock type, ρ_R , and ρ_0 are the formation and the fluid resistivities, and γ and μ are the specific weight and viscosity of the fluid. The two principal components of the hydraulic conductivity tensor, K_{ht} and K_{hl} are then calculated by substituting the transverse and longitudinal resistivities, ρ_t and ρ_l into (Equation 2.7). These values are then divided to obtain the hydraulic anisotropy, ψ_h ,

$$\Psi_h = \frac{K_{hl}}{K_{ht}} = (\lambda^2)^{(b_2/m)} \quad (2.8)$$

where λ is coefficient of electrical anisotropy defined in (Equation 2.1).

The dimensionless parameters, a_2 and b_2 are determined from regression analysis of log porosity and permeability using the empirical expression given by Archie (1942),

$$k = a_2 \Phi^{b_2} \quad (2.9)$$

In the present study a_2 and b_2 were estimated from 250 porosity and permeability measurements obtained from 25 fully penetrating wells across the Biscayne Aquifer (Cunningham et al., 2006). The regression plot for the logarithm parameters is shown in Figure 2.11. Analysis of Figure 2.11 derived estimates for a_2 and b_2 of 4.22×10^{-06} and 5.69, respectively. A cementation factor, m , of 1.7 and a tortuosity factor, α , of 1 were assumed (Kwader, 1985). These parameters and the average electrical anisotropy were used to calculate K_{hl} , K_{ht} and ψ_h at each site.

The calculated values for K_{hl} , K_{ht} and ψ_h for each site are shown in Table 1 and Figure 10b. A mean value of 8 cm/s and 5 cm/s were obtained for K_{hl} and K_{ht} , respectively and compares favorably with the hydraulic conductivity of 3.5 cm/s

estimated by Fish and Stewart (1991). Hydraulic anisotropy ranged from 1.18 to 2.83 with a mean of 1.51. Similar to the anisotropic porosity, values on the ridge are higher with lower values west of the ridge. In general, most sites show a good correlation between anisotropic porosity and hydraulic anisotropy with the exception of MC (Figure 2.1), where the secondary porosity is low while the hydraulic anisotropy is high.

2.6 DISCUSSION

The coefficient of electrical anisotropy obtained from the study generally ranged from 1.01 - 1.12, though most measurements were 1.04 or less. These results are well within the expected range of electrical anisotropy for limestone rock which varies from 1 to 1.2 (Telford et al., 1990). In general, the anisotropy was lower than that reported in other azimuthal resistivity studies which showed values ranging between 1.1 and 1.5 (Taylor and Fleming, 1988; Lane et al., 1995; Boadu et al., 2005). However, these studies were conducted in fractured non-carbonate rock terrains. At one site (WP, Figures 2.1 and 2.4) the anisotropy was as high as 1.36. The high anisotropy may be the result of large solution cavities and is the subject of further investigation.

Previous studies have found total porosities in the Biscayne Aquifer of 40% or higher which is largely a results of secondary dissolution (Fish and Steward, 1991; Cunningham et al., 2006). These values are similar to estimates of the total porosity calculated from Archie's law (Table 2.1). The electrical measurements in this study were used to estimate a component of this porosity which contributes hydraulic anisotropy in the aquifer. At most sites, this anisotropic porosity is 2% or less and therefore contributes

less than 5% to the total porosity. At some sites on the Atlantic Coastal Ridge, this anisotropic porosity may contribute up to 10% of the overall porosity. Although a 2% anisotropic porosity might not seem significant in the numerical sense, a recent study by Worthington et al. (2012) in a carbonate aquifer with hydraulic conductivities far less than that of the Biscayne Aquifer showed that a net gain of 0.1% in effective porosity can double the normal rate of groundwater velocity.

Anisotropic porosity and hydraulic anisotropy are highest on the Atlantic Coastal Ridge and generally decrease to the west (Figure 2.10). The higher values suggests that the geology and topography is an important factor in the development of the anisotropy in the Miami Limestone of the Biscayne Aquifer. Evans and Ginsburg (1987) noted that dissolution is higher on the Atlantic Coastal Ridge than the low elevation areas west of the ridge. The high dissolution areas coincide with the oolitic facies of the Miami Formation which contains more soluble aragonite (Hoffmeister et al., 1967). The rocks to the west of the ridge grade into the calcite rich bryozoan facies which dissolves more slowly. Hence, the difference in the rate of dissolution could generate the higher hydraulic anisotropy observed on the ridge.

The direction of minimum resistivity may also be related to the preexisting sedimentary fabric. While outcrops are rare in the study area, one site, MC (Figure 2.1), is located near a 3m high bluff which contains cross beds dipping 20-30°. The strike of these beds, $70 \pm 2^\circ$ is similar to the direction of minimum resistivity, $60 \pm 5^\circ$, at this location. This suggests that in some cases, the minimum resistivity may be associated with cross bedding in the Miami Formation.

The observed anisotropy may have developed in response to dissolution in the direction of the paleo groundwater flow. The current groundwater flow direction in the study area generally trends E/SE and is influenced by man-made features such as canals, levees and wells (Fish and Stewart, 1991). This differs from predevelopment conditions where a water table high beneath the Atlantic Coastal Ridge acted as a barrier to groundwater flow. Groundwater flowed away from both E/SE towards Biscayne Bay and W/SW towards the Everglades (Parker et al., 1955). When drainage canals were constructed across the ridge, the hydrostatic head decreased allowing regional groundwater flow towards the E/SE. Fennema et al. (1994) modeled the groundwater flow under predevelopment conditions (Figure 2.12). Beneath the Atlantic Coastal Ridge, the flow was highest and was oriented towards the SE consistent with the observed direction of the minimum resistivity and the higher magnitude of anisotropy. Farther west, on the margin of the Everglades, the flow was oriented towards the W and SW and was smaller in magnitude. This is consistent with most of the observed anisotropy to the west of the ridge (Figure 2.10). This suggests a causal mechanism linking the direction of the minimum resistivity and maximum hydraulic conductivity to dissolution caused by the paleo-groundwater flow.

The findings of the study demonstrate the capability of azimuthal resistivity square array method to characterize the anisotropy of the surficial aquifer. The study shows that in the absences of transmissivity drawdown data and observe interconnected vugs in the karsts environments, indirect geoelectrical measurement be useful in obtaining valuable information about a subsurface. Such information is important in terms of geology, groundwater flow and transport properties of non-fractures karts

terrain. Ultimately, incorporating anisotropy information into groundwater models like MODFLOW would provide a clearer picture about the subsurface instead of assuming isotropic conditions.

2.7 SUMMARY

Over 100 azimuthal resistivity measurements were collected in eastern Miami-Dade County of Florida through the use of the square array to investigate electrical and hydraulic anisotropy in the Biscayne aquifer. These measurements were inverted for the mean resistivity, coefficient of electrical anisotropy and minimum resistivity direction for each site. The electrical parameters were then used to estimate the anisotropic component of the secondary porosity and the principal components and direction of the hydraulic conductivity tensor.

The coefficient of electrical anisotropy ranged from 1.01 - 1.36 with a modal value of 1.03. The estimated anisotropic component of the secondary porosity ranged from 1 to 11% with a modal value of 2%. The hydraulic anisotropy (the ratio of the minimum to maximum hydraulic conductivity) ranged from 1.18 to 2.83. The anisotropy is greatest in the central portions of the Atlantic Coastal Ridge, with the maximum hydraulic conductivity at most sites trending E-W/SE-NW. In regions to the west of ridge the hydraulic anisotropy is lower and trends both towards the SE-NW and E-W. In both cases, the magnitude and direction of the anisotropy is consistent with modeled predevelopment groundwater flow. This suggests that the observed anisotropy may be

attributed to the dissolution of limestone which enhanced the hydraulic conductivity in the predominant direction of the paleo-groundwater flow.

The study shows that azimuthal resistivity surveys can provide reasonable estimate of primary hydrological parameters such as anisotropy, secondary porosity and conductivity tensors in carbonate aquifer systems. The multi-electrode square array technique employed in this study enhances the speed of measurements and is able to effectively measure even small anisotropy values with very high accuracy. This paper further advances the hydrogeophysical approach which calls for the use of geophysical methods for quantitative measurement in regional hydrological studies.

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Table 2.1: Summary hydraulic parameters at each site calculated from electrical resistivity. Geographic locations of sites are shown in Figure 5.1. K_{ht} and K_{hl} are the transverse and longitudinal horizontal hydraulic conductivity respectively.

Site	Secondary porosity (Φ)	Total porosity	Hydraulic Conductivity K_{ht} (min) m/s	Hydraulic Conductivity K_{hl} (max) m/s	Hydraulic Anisotropy Ψ_h
BL	0.021	0.32	0.046	0.069	1.51
CO	0.033	0.33	0.057	0.097	1.71
DF	0.011	0.34	0.079	0.094	1.19
EC	0.008	0.29	0.029	0.035	1.19
MC	0.011	0.29	0.028	0.039	1.78
ML	0.016	0.34	0.069	0.092	1.33
NT	0.011	0.31	0.043	0.050	1.18
PP	0.041	0.42	0.229	0.333	1.45
PL	0.010	0.22	0.032	0.044	1.39
SC	0.020	0.33	0.057	0.087	1.52
SL	0.010	0.30	0.038	0.048	1.26
WP	0.055	0.34	0.048	0.136	2.83
WL	0.020	0.38	0.133	0.167	1.26

Figure 2.1: Map of azimuthal resistivity survey sites and topography. Topographic source USGS 30m DEMs. BL-Bird Lake Park; CO-Camp Owaissa Bauer; DF-Dante Fascell Park; EC-FIU Engineering Center; ML-Miller Pond Park; NT-North Trail Park; PP-Palmer Park; PL-Palmlan; PL-Palmland Park; SC-Snapper Creek Well Field; SL-Sun lake Park; WP-West Perrine Park; WL-Wild Lime Center Park; MP-Modello Park; CP-Continental Park.

Figure 2.2: Conceptual geophysical model of a homogeneous anisotropic Earth caused by vertically dipping beds or fractures. ρ_t : transverse resistivity perpendicular to the bedding planes or fractures; ρ_l : longitudinal resistivity parallel to the bedding planes or fractures; ρ_l : longitudinal resistivity parallel to Figure the bedding planes or fractures; θ : strike direction.

Figure 2.3: Schematic diagram of the square array. $\theta_0-\theta$ is the angle between the array and the direction of longitudinal (minimum) resistivity as defined in Habberjam (1972) and Equation 3. A and B correspond to the current electrodes and M and N correspond to the potential electrodes in the alpha configuration.

Figure 2.4: Schematic illustration of square array deployment using the 28 electrode system. 1 and 22 are current electrodes and 8 and 15 are potential electrodes in the first of the rotated. The array was increased from a radius of 2.83 m to 32 m in increments of $\sqrt{2}$.

Figure 2.5: Examples of polar plots of square array apparent resistivity plotted against azimuth (deg). The thick solid line is the best fitting apparent resistivity ellipse obtained from Equation 3. Locations of sites are shown on Figure 5.1. See Appendix 1 for all the polar plots.

Figure 2.6: Distribution of the coefficient of anisotropy (λ) measurements for the all the sites and square sizes in the study.

Figure 2.7: Directional distribution of minimum resistivity, θ , for the all the sites and square sizes in the study.

Figure 2.8: Map showing the direction of minimum resistivity in study area for the square array sizes with effective depth in parenthesis. The size of the arrows is proportional to the magnitude of the coefficient of anisotropy.

Figure 2.9: Distribution of the anisotropic secondary porosity measurements for sites grouped by geographic region. Coastal sites: PL, MC, DF; Ridge sites: CO, PP, WP; West sites: SC, WL, SL, BL, ML, EC, NT. Location of the sites is shown in Figure 5.1.

Figure 2.10: Map showing the average direction of minimum resistivity of all measurements at each site (Davis, 1986) and a) calculated anisotropic secondary porosity (%) and b) horizontal anisotropy for each site in the study area. The size of the arrow is proportional to the magnitude of anisotropic porosity and hydraulic anisotropy.

Figure 2.11: Field derived logarithmic relationship between porosity and permeability from 25 fully penetration wells into the Biscayne Aquifer. Data source (Cunningham et al., 2006).

Figure 2.12: Pre-development groundwater flow direction modified from Fennema et al. (1994). The size of the arrows is proportional to the groundwater flow rate.

3 SQUARE ARRAY RESISTIVITY SOUNDING OF THE BISCAYNE AQUIFER IN MIAMI-DADE COUNTY, FLORIDA

ABSTRACT

Azimuthal geoelectrical sounding was carried out across 13 sites in the Biscayne Aquifer of SE, FL in order to determine the resistivity variation with depth and the bulk conductivity extent of the coastal aquifer. Observed resistivity and well data close to the sites were used to generate a series of resistivity-depth models for 5, 10 and 16 layers. Depth mapping of resistivity and porosity was estimated from resistivity data using Archie's Law to understand the extent of saltwater intrusion. Resistivity of 30 Ω -m or higher was generally observed in the saturated zone in most of the study sites which is interpreted as freshwater. However, along the coastline, resistivities as low as 1 Ω -m was present, an indication of saltwater intrusion. In general, resistivity decreases from the NW (close to the margins of the Everglades) to the SE of the study area along the Biscayne Bay. The estimated porosity ranged from 18 to 61 % with a mean of 30 % at 10 m depth. The variation might be attributed to the changes in porosity because of cavities in the fresh water zones. At the regional scale, the calculated porosities are higher along the Atlantic Coastal Ridge compared to the areas west of the ridge. Results of the azimuthal soundings compare favorably with similar published sounding studies using time domain and helicopter-based electromagnetic methods. The study shows that azimuthal sounding resistivity surveys can be used to provide fast resistivity and porosity data for surficial coastal aquifers or as a primary interpretation tool to serve as background for a more detail hydrogeophysical studies.

3.1 INTRODUCTION

The Biscayne Aquifer is a Pleistocene unconfined carbonate aquifer located in southeast Florida and serves as the principal source of water for all of Miami-Dade, Broward, and Monroe Counties (Miller, 1990). Surficial coastal aquifers systems like the Biscayne Aquifer are often prone to contaminant pollution from saltwater intrusion (Fitterman et al., 1999). Contaminated land and groundwater constitute a complex system with numerous interacting processes; hence identifying possible area of pollution is important step towards the management and protection of coastal aquifer. In the Biscayne aquifer, a network of hydrogeological monitoring wells and induction logs along the imaginary extent of saltwater interface (ESI) are implemented to monitor periodic levels of chloride in the groundwater (Fitterman et al., 2012). Due to the scale of the Biscayne Aquifer, it is not economically feasible to cover the full extent of the aquifer with wells. In addition wells usually capture the immediate surroundings and cannot reliably provide accurate information about area not in its cone of influence. Furthermore, the complexity of the Biscayne Aquifer as a result of its inherent karst features makes extending point measurement to the whole aquifer unreliable.

Surface geophysical methods (e.g., electrical resistivity, time domain and helicopter electromagnetic soundings) are widely used as preliminary assessment tools to routinely provide electrical resistivity (inverse of electrical conductivity) information relating depth of conductive fluids and map the ESI in the Biscayne Aquifer (Fitterman et al. 1999; Fitterman and Prinos 2011; Fitterman et al. 2012; Loke et al.2013). The electrical resistivity of a medium depends mainly on groundwater salinity, saturation,

aquifer lithology, and porosity (Shaaban, 2002). The attractiveness of these methods for characterizing aquifer properties is because of advantages such as the noninvasive nature of the techniques, cost effectiveness, fast data acquisition, ability to generate large data set for mapping the subsurface and reduction in the number of installed monitoring wells (Fitterman and Prinos 2011; Hinnell et al., 2010; Huisman et al., 2010). Although resistivity can be inferred from electromagnetic (EM) soundings, since it is the inverse of conductivity; studies have shown that EM surveys are more sensitive to conductive bodies than to resistivity bodies (Burger et al., 2006). Hence for geological material like limestone which has low conductivity, the use of direct current (DC) resistivity methods to measure apparent resistivity is highly advantageous. Besides, DC resistivity unlike EM methods has physical contact with the subsurface during the entire duration of the survey. This reduces the effects of uncontrollable factors such as overhead power lines that might affect helicopter EM survey in urban centers. Also resistivity survey is relatively cheaper and less resource intensive compared to sounding method like the helicopter EM. In a highly populated urban setting like the Miami-Dade County, it is evident that the use of DC method to obtain resistivity cannot be underestimated.

The focus of the current paper is to apply square array resistivity sounding techniques to understand the variation of resistivity with depth in the Biscayne Aquifer. The study aims to get a holistic picture of spatial variation for both the geological layers and the groundwater table on a large scale (20 km inland from the coastline). The specific objectives of the study are to (a) determine the spatial variation of resistivity at different depth within the aquifer through resistivity modeling constrain with existing well data,

(b) generate a resistivity depth map to delineate the extent of seawater intrusion and (c) estimate and map the porosity of the study area through Archie's law. Knowledge of landwards extent of saltwater and porosity in the saturated zone will play a critical role in effective water protection and management of the Biscayne Aquifer.

3.2 STUDY SETTING

The rocks of the aquifer in the study area composed of the Miami Limestone on the surface underlain by the Fort Thompson Formation. A detailed discussion of the hydrogeological setting of the Biscayne Aquifer can be found in studies like Parker et al. (1955), Fish and Stewart (1991), Cunningham et al. (2004, 2006, 2009), Renken et al. (2008) and Cunningham and Florea (2009).

In the study area, the surface elevation increases from the low lying areas on the margins of the Everglades in the west to the Atlantic Coastal Ridge along the east coast with elevations reaching 6 m above sea level (Figure 3.1). The thickness of the aquifer in the study ranges from 17 m to 32 m (Fish and Stewart, 1991; Fitterman et al., 2012). The study sites were restricted to natural undisturbed areas, open space with absence of artificially buried materials and utilities. Survey locations included sites (1) near the shoreline of Biscayne Bay where seawater has intruded into the aquifer; (2) on higher elevation regions of the Atlantic Coastal Ridge where the unsaturated zone is relatively thick; and (3) in lower elevation regions to the west, close to the Everglades where the unsaturated zone is relatively thin. Two of the sites were near the artificially dredged canals in the county. These canals are situated in natural topographic lows known as

transverse glades and were constructed in the early 20th century for flood control purposes. However the canals on numerous occasions lead to saltwater contamination of freshwater inland from the Biscayne Bay either through high tide, drought or groundwater pumping (Parker et al. 1955; Fish and Stewart 1991).

Porosities in the aquifer range from 5.5% to 79% with most sites showing a maximum porosity around 40% (Fish and Stewart 1991; Cunningham, 2004; Manda and Gross, 2006; Cunningham et al., 2006). The porosity for the Biscayne Aquifer has been shown to vary depending on the depth (Cunningham, 2004) and the location of observation (Renken et al., 2005) that results from complex dissolution features of the carbonate rocks. Induction logs collected in Miami-Dade County indicate that the bulk resistivity freshwater in the saturated zone ranges from 35 to 220 Ω -m (Fitterman et al., 2012). Fitterman and Prinos (2011) concluded that, the changes in bulk conductivity with depth observed in induction logs are most likely the result of abrupt changes in porosity similar to those illustrated in Figure 3.2.

3.3 METHODOLOGY

3.3.1 1-D RESISTIVITY MODELING

One dimensional (1-D) vertical electrical soundings (VES) are routinely used to locate conductive targets such as water table depth, aquifer thickness and in coastal areas, salt water intrusion as it provides good vertical resolution with depth. One dimensional resistivity sounding works best in complex geological situation where there is rapid

vertical change with depth. The analysis involves modeling the variation of resistivity with depth. A key assumption used in 1-D resistivity modeling is that the subsurface consists of uniform horizontal layers and resistivity changes only with depth. A detailed discussion of the various resistivity techniques including 1-D sounding and its limitations can be found in a recent review paper by Loke et al. (2013).

In this study, resistivity soundings were conducted at 13 sites in the study area (Figure 3.1) using the square array resistivity sounding technique (Habberjam, 1972). The square array is an alternative electrode configuration to linear measurement of resistivity where current and potential electrodes are deployed on opposite sides of a square and the electrode spacing is varied for each measurement over a fixed center (Habberjam and Watson, 1967). The use of square array sounding have been shown by Merlanti and Pavan, (1996) to provide greater significant information compared to classical linear vertical linear electric soundings (Wenner and Schlumberger) as demonstrated in Figure 3.3. The square array is exhaustive and provides higher resolution at greater spacing when VES is performed than Wenner and Schlumberger (Merlanti and Pavan, 1996). The improvement is to the result of the ability of the square array to be less susceptible to obscured heterogeneities, bedrock relief, electrode placement errors and other sources of noise (Lane et al., 1995).

Field measurements of 1-D sounding were made with an Advance Geoscience Incorporated (AGI) Super-Sting R1/IP 28-electrode resistivity imaging system. The system utilizes a multi-core cable controlled by a programmable switch box which controls the channels of the current and potential electrodes for each measurement. The

28 electrodes were placed at equal angles on a circle forming seven separate square array configurations rotated at 12.86° intervals (Figure 3.4). The array was oriented to magnetic north with a Brunton compass and results were corrected to true north using the local declination of six degrees. A custom command file was created which recorded 28 separate measurements of the square array alpha configuration rotated through 360° . The 28 electrodes were initially deployed at a radius of 2.83m and after each set of azimuthal measurement; the radius of the array was expanded in increments of $\sqrt{2}$ to 32 m. This resulted in sets of azimuthal measurements for square array lengths, A , of 4.0, 5.7, 8, 11.3, 16, 22.6, 32 and 45.3 m. The procedure produced a set of soundings which was used to investigate the resistivity variation with depth.

In order to model the apparent resistivity in the EarthImager 1-D software, the azimuthal sounding 'A' spacing were converted to the equivalent Wenner 'a' spacing. Most 1-D modeling software does not use the square array hence the need to convert the square array to Wenner. The equivalent spacing relation between square array and Wenner array is $a = 0.75A$. At each A-spacing the square array sounding provide 28 resistivity data points. These points are averaged for each A-spacing resistivity. The Wenner a spacing and measured average apparent resistivity for each sounding is used as the input file for the model.

The data were modeled using the EarthImager 1-D modeling program (AGI, 2012). Occam's inversion algorithm is used in EarthImager 1-D to resolve layered resistivity on the basis of a set number of layers as inversion thickness remains constant during the inversion process (Constable et al., 1987). There are limitations to 1-D

sounding model, as resistivity-depth models determined by inversion are usually suffer from non-uniqueness, equivalence and often difficult to resolve a thin layer. To overcome some of these challenges due to equivalency and non-uniqueness; three (3) different layered models are employed in this study as the final outcome of the sounding is dependent the number of layers., Three (3) different layered models were generated in the Earth Imager 1-D using 5, 10 and 16 layers together with the default inversion parameters (depth factor of 1.1, thresholds (%) of 10, damping factor of 10, etc.). For example, the 16 layer model was parameterized with layer thickness ranging from 0.5 m at the surface to 3.9 m at 22m depth. A smooth inversion technique was chosen to eliminate the need to guess starting values close to the actual formation resistivity (Turesson, 2006). The observed apparent resistivities (measured field data) were compared with the model (calculated) resistivity of 16 layer model for sensitivity analysis.

3.3.2 DATA ANALYSIS

In order to constrain the measured geoelectrical data set, water table data were obtained from a network of groundwater wells close to the sites (Figure 3.5) monitored by United State Geological Survey (USGS) , South Florida Water Management District (SFWMD) and Miami Dade Water and Sewerage (MDWS). Most of these data are in the DBHYDRO database (<http://www.sfwmd.gov/dbhydroplsql/>). Where two or more monitoring wells are present at the site (e.g., SC) the average of them is used to estimate the water table for the site. In the event where the closest well to the site is only monitored periodically, the historical mean is chosen from available data. Using the

North American Vertical Datum of 1988 (NAVD 88) as the reference surface elevation below the sea level, surface topographic data and groundwater level in NGDV 29 were converted using the VERTCON software (http://www.ngs.noaa.gov/cgi-bin/VERTCON/vert_con.prl).

The study employs ordinary kriging geostatistics analyst tool in the ArcGIS software to interpolate data from 13 sites across the study area to create spatial distribution maps for further hydrological analysis. Spatial interpolation is the process of predicting the values of attributes at unsampled sites (Xie et al., 2011). Ordinary kriging is one of the spatial interpolation methods used to characterize patterns in geological properties over various spatial scales (Childs, 2004; Anselin and Getis, 2010) that quantify the spatial autocorrelation among sample points based on the entire dataset within the study area. It assumes a constant but unknown mean, $\mu(x) = \mu$, over the study region; and the estimated locations of the data points increase based on the globally calculated semi-variogram (Meng et al., 2010; Anselin and Getis, 2010). The limitation of spatial interpolation method is that different interpolation methods will almost always produce different results (Childs, 2004).

3.3.3 POROSITY ESTIMATION

The porosity of the study area was estimated using Archie's law (Archie, 1942). Archie's Law (Equation 1) is the basic equation used by petrophysicists to determine whether a formation has water or hydrocarbons in the pore space (Ellis and Singer, 2007). In general Archie's law relates electrical properties to rocks and is commonly stated as:

$$\rho_R = \rho_w \Phi^{-m} S_w^{-n} \quad (1)$$

where ρ_R is the bulk resistivity of the rock, ρ_w is the resistivity of water contained in pore structure, Φ is the porosity, S_w refers to the saturation level, m and n are constants of rock cementation and saturation exponent. In the saturated zone where rock is filled with water ($S_w = 1$) the Equation 1 reduces to

$$\Phi = (\rho_w / \rho_R)^{-1/m} \quad (2)$$

Where ρ_R is the bulk resistivity of the aquifer estimated from the 1-D models, ρ_w is pore water resistivity estimated from groundwater well data (Figure 3.5), and m is an empirically derived constant which depend on rock type. In this study a cementation factor m , of 1.7 was assumed (Kwader, 1985). Using the Equation 2, an estimate of the porosity at 5 and 10 m below sea level was obtained based on well data (Figure 3.5). Typical groundwater resistivity for sites on the center (CO, PP, WP; Figure 3.1) and west of the ridge (SC, WL, SL, BL, ML, EC, NT, Figure 3.1) ranges from 16-23 Ω -m (specific conductance of 400-600 μ S/cm). For sites behind or near the ESI (MC, DF and PL, Figure 3.1), pore fluid resistivity from nearby groundwater wells (3.0 Ω -m, 8.6 Ω -m and 10.0 Ω -m respectively; Figure 3.5) was used.

3.4 RESULTS

The calculated apparent resistivity from the model was nearly identical to the measured observed resistivity for all sites except MC (Figure 3.6) where less than 8 sounding spacing's were collected. At MC the model provided a poor fit as evidence by

the high RMS associated with it. The model apparent resistivity lies outside the observed resistivity measured at some of the depths at these sites. Overall, the 1-D model did a good job predicting the interface between the saturated and unsaturated zone. The geoelectrical water table which is defined as resistivity below 220 Ω -m in this study agrees with the mapped water table for all of the sites except MC (Figure 3.6e).

Model resistivity in the saturated zone with few exceptions ranges from 30 Ω -m – 200 Ω -m. The apparent resistivity-depth sections were nearly identical for each layered (5, 10, 16) model. At most sites, the resistivity depth profile was very uniform in the saturated zone with minor changes in resistivity (< 20 Ω -m). The resistivity depth profiles for some of the sites (BL, DF, ML, PP, SC, WL, WP) show a sharp decrease below the water table followed by an increase in resistivity. For example at BL (Figure 3.6a) the resistivity increases from ~ 70 Ω -m just below the water table to ~ 130 Ω -m at -5m elevation. Similarly, at WL (Figure 3.6l), the resistivity increases from ~ 30 Ω -m below the water table to almost 150 Ω -m at -5 m elevation before decreasing to 40 Ω -m after 15 m. The observed resistivity decreasing below the water table and increasing afterward is not unique to this study. In similar resistivity-depth profiles using both time domain and helicopter electromagnetic soundings and induction logs; south of our study area (Figure 3.1) Fitterman and Prinos (2011) observed similar resistivity depth characteristics. They attributed this behavior is to existing solution cavities in the saturated zone which cause bulk resistivity to decrease around the solution hole and increase afterwards (Figure 3.2). Hence the sharp change in resistivity in this study could be attributed to the presence of similar solution cavities.

In contrast the resistivity for sites like CO, EC, NT, PL, SL either increases slightly in the saturated zone or remains the same with depth. For example at NT (Figure 3.6g), the resistivity does not decrease below the water table but rather increases from ~50 ohm-m just below the water table to ~230 Ω -m at -5m elevation. At CO (Figure 3.6b) the resistivity for all the 3 different layers remains the same throughout the saturated zone. There is an exception to these trends particularly for MC (Figure 3.6e) where the resistivity decreases from a high of 800 Ω -m in the unsaturated zone to 1 Ω -m at 20 m below sea level. The uniqueness of the results in MC is not surprising as it is located on Atlantic Coastal Ridge which has higher elevation (by south Florida standard) and also very close to the Biscayne Bay. The Biscayne Bay is a ~2m deep and 20 km wide coastal lagoon system located east of the study area (Figure 3.1) where inland fresh water from artificial canal system interact with the water from the Atlantic Ocean (Fish and Stewart,1991 and Stalker et al. 2009). Previous study at the site by Stalker (2008) shows the presence of both blackish and saline water.

3.4.1 Depth - Resistivity Map

The 10 layer 1-D resistivity models were used to generate iso - resistivity maps at depths of 5, 10, and 15 m using ordinary kriging interpolation in ArcGIS (Figure 3.7). Results indicate spatial variation between the 3 maps but in general, the resistivity at each of the depths decreases from NW and to SE. At 5 m depth (Figure 3.7), high resistivity was dominant in the study area, even including coastal sites like MC (Figure 3.1). At MC there is a thin fresh water lens on the surface of the saturated zone, which is most likely

responsible for the high resistivity observed at the 5 m depth. The only exceptions are low resistivity zones observed at WP and PP (Figure 3.7) both located on the ACR. Typically in an aquifer, low resistivity is due to saltwater intrusion, contaminant or presence of solution cavities. In this instance, it is likely due to solution cavities as the first two does not apply because the sites are inland of the saltwater water intrusion line (Fitterman and Prinos, 2011) and pore fluid data from wells close to them (Figure 3.5) shows fresh water without any contamination. For example at the WP site where detail hydrogeophysical depth analysis was carried out (Chapter 4); a sharp decrease in resistivity at depth of 4-11 m was most likely attributed to solution cavities in the bedrock.

The resistivity depth map at 10 m and 15 m below the NAVD88 show higher resistivity in the NW and W part of the study area and decreases towards Biscayne Bay especially in the NE. The saltwater zone was depicted well in the NE section of the study area on the 10-15m map, but not in the SE section of the study area due to a lack of data collected in that region. This is a limitation of the spatial interpolation method employed in the study where in the absence of local neighborhood data sets, the interpolation gravitates towards the regional mean.

The resistivity in the study area with few exceptions decreases with depth as evident by lower resistivity at 15 m compared to 5 m and 10 m. Most of the study area behind the imaginary salt water intrusion line still contains fresh water in the saturated zone and is not affected by saltwater intrusion (Figure 3.7). In contrast there is significant change in resistivity with depth at MC which is located adjacent to the Biscayne Bay.

Sampled groundwater resistivity from an uncased groundwater well 10.9 m deep at the site showed 3 Ω -m pore fluid resistivity at the base of the well which is consistent with saltwater (Stalker, 2008). Similarly, an induction log done by Fitterman and Prinos (2011) shows saltwater intrudes at the sites between the extent of salt water intrusion line and the Biscayne Bay.

3.4.2 Depth Porosity Map

Kriging interpolation in ArcGIS was used to create a porosity map of the study area (Figure 3.8). The spatial pattern of porosity from interpolation showed a clear SE-NW/W gradient. This clearly is a reflection of estimated porosity values from Archie's Law which ranged from 18 to 61% with a mean of 30% (Table 1) with the distribution of peak intensity corresponding to the sites with higher porosity. The porosity determined from this study compares favorably with other using direct methods (Fish and Stewart 1991; Cunningham, 2004; Manda and Gross, 2006; Manda and Culpepper, 2013) as well as data from core and in-situ logging from 250 data points from 25 fully penetrating wells across the Biscayne Aquifer Figure 3.9 (Cunningham et al., 2006). Porosity data presented in Figure 3.9 are from N and NW outside of the study area and range between 5.5 and 50.2 % with a mean of 26 %.

The nature of the porosity observed at both 5 m and 10 m below sea level indicate higher porosity values in the SE of the study area along the ACR and lower values in the W and NW away from the ACR. Almost all the sites west of the ACR had porosity values less than 30%. The lowest porosity of 18% at 5m depth in Figure 6 is observed at

NT while the highest porosity of 61% was observed in the middle of the ACR at PP -10 m elevation. The estimated higher porosities were generally found in the SE and E regions of the study areas and transition into the lower porosities areas in the NW. In general, higher porosities (30-60 %) were observed following areas of higher elevation (Figure 3.1), while lower porosities (18-33 %) were observed in low lying areas (Figure 3.8).

3.5 DISCUSSION

The results of the resistivity depth model demonstrate the ability of using 1-D resistivity sounding methods to provide depth information about the Biscayne Aquifer as well as to capture the extent and distribution of resistivity values of the Miami limestone and Fort Thompson formations in the saturated zone. The resistivity depth profile from this study compares favorably with recent TEM and HEM sounding data interpreted for resistivity (Fitterman and Prinos 2011; Fitterman et al. 2012). For example these studies determine the resistivity in the saturated zone for fresh water to range from 220 to 30 Ω -m which corresponds to the range of values (230 – 30 Ω -m) obtained in this study. At MC the observed low resistivity value less than 10 Ω -m can be interpreted as saltwater corresponding with Fitterman et al., (2012) conclusions that in the Biscayne Aquifer resistivity values of less than 10 Ω -m are an indication of saltwater.

Overall, for most of study area, the resistivity depth profile is relatively uniform in the saturated zone with only minor changes in resistivity observed. The resistivity for some of the sites show either a sharp decrease below the water table followed by an

increase in resistivity or an increase below the water table followed by a decrease in resistivity. The karst features of the aquifer including solution cavities might be responsible for these minor changes in resistivity with depth. Fitterman and Prinos, (2011) attributed a similar pattern in changing resistivity below the water table to the presence of cavities in the Biscayne Aquifer like those shown Figure 3.2. However, the major change in resistivity with depth observed in this study, such as at site MC (Figure 3.6e) is driven by saltwater intrusion in the aquifer.

The resistivity - depth map of the study area enables the interpretation of the approximate extent of the saltwater intrusion and the water quality at various depths in the aquifer. For most of the sites in the study area, the resistivities were above 30 Ω -m independent of depth. Fitterman and Prinos (2011) concluded from their hydrogeophysical study that in the Biscayne Aquifer, resistivity above 30 Ω -m in the saturated zone is indicative of fresh water while those values less than 10 Ω -m are attributed to saltwater. Evidence from this study shows current levels of saltwater intrusion are moderate to small and does not pose present danger to water pumping areas like the Snapper Creek Municipal Well Field where resistivity of ~ 50 Ω -m is apparent even at -20 m elevation which is the base of the aquifer at the site. This might change in the future, but studies conducted over the past two decades (Fitterman et al., 1999; Fitterman and Prinos 2011; Fitterman et al., 2012) show only minor changes of saltwater intrusion in the Biscayne Aquifer beyond the ESI in the study area.

In general, the study indicates that estimated porosity is dependent on the elevation of 1-D soundings in the study area. This is not surprising as the porosity of the

Biscayne Aquifer has been shown to vary with depth (Cunningham, 2004) and the location of observation (Renken et al., 2005). Even at the same spot, data from core and in-situ logging sometimes have different porosities at different depths (Cunningham et al., 2006). On the regional scale, the higher porosity areas are generally observed along the Atlantic Coastal Ridge while lower porosities are observed in the sites to the west of the ridge. Geological studies have shown the rate of limestone dissolution is higher on the ridge where more soluble aragonite is present compared to the west of the ridge where the calcite rich bryozoan facies which dissolves more slowly (Hoffmeister et al., 1967). The higher dissolution rates along the ridge are believed to have contributed to the increased cavities inherent in the ACR of Miami-Dade County (Cressler, 1993). In a limestone carbonate aquifer with karst features, porosity can be governed by many factors but in the Biscayne Aquifer higher dissolution in cavities (e.g. solution holes) play a prominent role.

3.6 CONCLUSIONS

In this study, 1-D soundings were made across 13 sites in the Biscayne Aquifer to investigate how resistivity varies regionally and with depth. These soundings were modeled in EarthImager1-D to provide different depth - resistivity models. The bulk resistivity of the saturated zone from the model resistivity and porosity estimated using Archie's law were used to generated depth map of eastern Miami – Dade County.

The model resistivity depth profile was relatively uniform in the saturated zone. For most of the study sites, resistivity in the saturated zone was above 30 Ω -m indicative

of freshwater in the bulk formation regardless of depth in the Biscayne Aquifer. The minor variation in resistivity ($< 30\text{-}50\ \Omega\text{-m}$) with depth is attributed to changes in porosity due to solution cavities in the fresh water zones. However, along the SE coastline, resistivity decreases from a high of $220\ \Omega\text{-m}$ to $1\ \Omega\text{-m}$. This major major changes in resistivity in the saturated zone ($> 50\ \Omega\text{-m}$) is due to saltwater intrusion along the coastal areas. These findings are consistent with similar published sounding studies which showed that the freshwater resistivity of the aquifer ranges from $35\text{--}220\ \Omega\text{-m}$, with saltwater exhibiting resistivity less than $10\ \Omega\text{-m}$. The estimated porosity in the saturated zone ranges from 18 to 61 % with a mean of 30%. This compares favorably with other porosity data from hydrogeological studies on the aquifer. On the regional scale, the Atlantic Coastal Ridge tended to have higher porosity values as compared to areas west of the ridge.

The study suggests that the 1-D resistivity model in the Biscayne Aquifer can provide useful information about the subsurface bulk formation features and water quality parameters if the model is well constrained with hydrogeological data. Such information may be coupled with groundwater models for accurate representation of surficial karst aquifer systems in the absences of hydrogeologic data. This approach could improve primary environmental assessment prior to detail investigation of transport and contaminant investigations.

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Table 3.1: Summary model electrical resistivity at each site calculated from and porosity. Geographic locations of sites are shown in Figure 5.1. All depth measurements are referenced to NAVD 88.

Site	Surface Elev (m)	Water Table- m	Resistivity at 5m	Resistivity at 10m	Resistivity at 15m	porosity 5m	porosity 10m
BL	1.66	0.91	183	183	183	24%	24%
CO	3.81	0.38	116	116	103	32%	32%
DF	1.82	0.35	186	186	96	24%	24%
EC	1.44	0.92	119	117	101	32%	32%
MC	5.18	-0.87	149	6	2	28%	38%
ML	1.78	0.82	186	186	124	24%	24%
NT	1.48	0.79	252	247	114	18%	20%
PP	1.92	0.45	45	39	38	56%	61%
PL	3.62	-1.22	162	138	109	26%	29%
SC	1.42	-0.38	110	302	153	33%	18%
SL	1.83	1.01	186	148	99	24%	28%
WP	2.03	0.75	65	87	109	45%	38%
WL	3.01	0.45	190	154	77	24%	27%

Figure 3.1: Map of azimuthal resistivity survey sites and topography. Topographic source USGS 30m DEMs. BL-Bird Lake Park; CO-Camp Owaissa Bauer; DF-Dante Fascell Park; EC-FIU Engineering Center; ML-Miller Pond Park; NT-North Trail Park; PP-Palmer Park; PL-Palmland Park; SC-Snapper Creek Well Field; SL-Sun lake Park; WP-West Perrine Park; WL-Wild Lime Center Park; MP-Modello Park; CP-Continental Park.

Figure 3.2: Digital borehole image of monitoring well G-3885 with induction log showing the effect of cavities on conductivity in the Biscayne Aquifer (Fitterman and Prinos 2011).

Figure 3.3: Vertical electrical sounding comparison performed with square (blue), Wenner (red) and Schlumberger (green) from a given bedrock. (Adapted from Merlanti and Pavan, 1996).

Figure 3.4: Schematic illustration of square array deployment using the 28 electrode system. 1 and 22 are current electrodes and 8 and 15 are potential electrodes in the first of the rotated squares.

Figure 3.5: Location of azimuthal resistivity soundings and groundwater wells close to the sites where water table data was obtained.

Figure 3.6(a-m): 1-D electrical model obtained from the resistivity sounding modeling and analysis (Figure 5.1). A) Observed and calculated resistivity from the 16 layer model. Error bars represent $\pm$ one standard deviation of the measured resistivity at each a spacing. B) The equivalent resistivity vs. depth section for the 5 (red), 10 (green) and 16 (blue) layer models with water table determine from wells close to the site. Model resistivities are nearly identical for each model. a) BL; (b) CO; (c) DF; (d) EC; (e) MC; (f) ML; (g) NT; (h) PL; (i) PP; (j) SC; (k) SL; (l) WL; (m) WP.

Figure 3.7: Modeled resistivity versus depth map at 5, 10 and 15 meters depth relative to the NAVD 88.

Figure 3.8: Porosity map at 5 and 10 meters depth relative to the NAVD 88.

Figure 3.9: Porosity distribution across 25 fully penetration wells into the Biscayne Aquifer (data source Cunningham et al., 2006)

4 DETERMINATION OF ANISOTROPIC KARST FEATURES IN THE BISCAYNE AQUIFER USING ELECTRICAL RESISTIVITY IMAGING (ERI) AND GROUND PENETRATING RADAR (GPR)

ABSTRACT

The Biscayne Aquifer is a highly permeable surficial limestone aquifer that covers a total of about 10000 km² in South Florida. The Biscayne Aquifer is characterized by cavities and dissolution features below the ground surfaces that are difficult to detect and quantify accurately because of their heterogeneous spatial distribution. Such heterogeneities exert a strong influence in the direction of groundwater flow as proposed by others. In this study we use an integrated array of hydrogeophysical methods to test their ability for detecting the lateral extent and distribution of dissolution features and to investigate their patterns of anisotropy in the Biscayne Aquifer. Hydrogeophysical methods including electrical resistivity imaging (ERI) and ground penetration radar (GPR) were constrained with direct borehole information collected in previous studies. Our geophysical results suggest the presence of a high conductivity zone (from ERI) and low EM wave velocity (from GPR) below the water table at depths of 4-9 m that correspond to the depth of solution conduits derived from digital borehole images. Electrical anisotropy derived from rotated square array measurements reported coefficients of anisotropy as high as 1.36. The cause of this higher anisotropy here is attributed to the presence of a solution cavity oriented in the E-SE direction. Evidence from this study indicates that groundwater flow in highly anisotropic and heterogeneous karst systems like the Biscayne Aquifer may be directly influenced by the nature and

orientation of dissolution features, and shows the potential of hydrogeophysical methods for better understanding such complexities.

4.1 INTRODUCTION

The presence of cavities and other dissolution features play an important role in the anisotropy of karst systems and often complicate studies related to groundwater flow modeling. A key factor that affects flow through rocks in aquifer systems is the dissolution of the pore network leading to interconnected conduit systems. The Biscayne Aquifer of South Florida is a surficial Pleistocene karst system characterized by dissolution features such as cavities, touching-vugs, conduits and solution holes that contribute directly to groundwater flow dynamics (Fish and Stewart, 1991; Cressler, 1993; Cunningham et al., 2009). The heterogeneous nature in physical properties of the rocks in the Biscayne Aquifer is mainly due to the variability in matrix porosity and the presence of isolated touching vug conduits (Cunningham et al., 2006). Well connected vugs can form major conduits resulting in preferential flow of groundwater, especially in carbonate rocks (Cunningham, 2004) of young eogenetic karst aquifers such as the Biscayne Aquifer) (Cunningham, 2004; Ginés and Ginés, 2007; Renken et al 2008). However, the presence of caves and dissolution features across the Biscayne Aquifer in Miami-Dade County is not fully understood particularly due to the limited number of exploration studies (Cressler, 1993). Flow in the Miami Limestone and the underlying Fort Thompson Formation in the Biscayne Aquifer is not uniform but localized through

secondary porosity and permeability caused by dissolution features that are highly anisotropic not only in terms of size but also in their spatial distribution (Fish 1988; Fish and Stewart 1991). Moreover, the likely connectivity of solution features in the Biscayne Aquifer would further indicate a more evolving role for localized anisotropy in karst features. Geyer et al. (2010) showed that drainage in karst aquifers at the regional scale may be controlled by the anisotropic nature of the conduit system. For all these reasons, investigation of dissolution features both in terms of spatial distribution and anisotropic properties may provide information about the lateral and vertical variability in groundwater movement and thus may help current modeling efforts to predict groundwater flow dynamics and contaminant transport in the Biscayne Aquifer.

Surface caves, sinkhole collapse and karst springs are commonly used to define the subsurface nature of dissolution features (White, 2002). However, such surface features are almost nonexistent or very isolated in the Biscayne Aquifer. Quantification of dissolution features at the aquifer scale often relies on single-point measurements such as borehole and core logging to investigate heterogeneity or tracer tests to determine connectivity (Ptak et al. 2004; Le Borgne et al. 2007). These methods tend to be invasive and suffer from limitations such as (1) the inability of single point measurement to determine interconnected conduit systems at larger scales and (2) the problematic nature of performing tracer tests in highly populated areas like Miami, Florida. Alternatively, indirect method such as near surface geophysics can provide dense spatial coverage and a means for locating karst features non-invasively (Bowling et al. 2005; Gibson et al. 2004). Specifically, ground penetrating radar (GPR) has been successfully used to

delineate cavities ranging from small vugs to caves and to qualitatively infer changes in porosity associated with dissolution features in the limestone of SE Florida (Grasmueck and Weger, 2002; Cunningham, 2004; Neal et al. 2008). This approach is not unique to the Biscayne Aquifer as surface geophysical methods like electrical resistivity and ground penetrating radar have routinely been applied to karst environments for cavity detection since the pioneering experiment for water-filled cavity detection by Holub and Dumitresku (1994). Chalikakis et al. (2011) provides a detailed review of contributions of geophysical methods to the exploration of dissolution features and parameters in surficial karst systems. Geophysical methods for imaging and detection of cavities in karst systems tend to be minimally invasive, provide good lateral resolution (typically in the order of cm), and typically result in faster and cheaper data acquisition when compared to traditional hydrological methods such as wells and boreholes. For all these reasons and despite the fact that they provide an indirect measure of the physical properties of an aquifer, hydrogeophysical methods are powerful tools for investigating flow dynamics at the regional scale particularly when constrained with some direct point measurements.

In addition, hydrogeophysical methods can also be used to investigate the anisotropic nature of rocks in the Biscayne aquifer. Electrical anisotropy occurs in the subsurface when current flows differently along different azimuthal directions. In a previous study by Yeboah-Forson and Whitman, (submitted), hydraulic anisotropy measured in the Biscayne Aquifer using geoelectrical methods determined coefficients of anisotropy less than 1.12 in most of the study area (Figure 4.1). However at one study site (West Perrine Park, Figure 4.1), the coefficient of anisotropy was found to be as high as 1.36 which is outside the typical range of coefficient of electrical anisotropy (1.0-1.2)

expected for limestone rocks (Telford, 1990). This paper intends to explore the local anisotropic nature (considered twice the regional average) of a particular site on the Atlantic Coastal Ridge (Figure 4.1) by expanding the electrical resistivity measurements conducted in Yeboah-Forson and Whitman (submitted), and including ground penetrating radar (GPR) surveys to constrain electrical resistivity results and characterize dissolution features. Indirect geophysical measurements are further constrained with existing high-resolution borehole images collected from nearby wells. The study has implications for groundwater flow and contaminant transport modeling in karst systems, particularly when considering how structural anisotropy may determine groundwater flow pattern in highly heterogeneous karst systems like the Biscayne Aquifer.

4.2 STUDY SETTING

The study site is located in West Perrine Park, and is bounded by latitude 25.609° N to 25.611° N and Longitude 80.359° W to 80.361° W in South East Miami, FL (Figure 4.1). The topographic elevation at the site is 3.01 m (NAVD 88). Existing digital borehole images from USGS wells G-3836 (located 2.00 km NE) and G-3889 (located 2.81 km SW) of the site (Figure 4.2) shows solution cavities starting at depth 4m below the surface. In general, the limestone rocks in the Biscayne Aquifer are very porous and have very high hydraulic conductivity due to the presence of similar moldic and vuggy porosity (Simon et al., 2007; Cunningham, 2004; Manda and Gross, 2006; Renken et al., 2008). Cunningham et al., (2006) classified the porosity into three groups: (1) touching-vug porosity (high permeability - conduit flow) (2) interparticle matrix and separate-vug

porosity (moderate permeability- diffuse-carbonate flow); and (3) conduit porosity (low permeability- fracture flow). These dissolution features in the saturated zones results in the high productivity of the extraction wells in the aquifer which supply water needs for the inhabitants of Miami-Dade, Broward and Monroe Counties in South Florida (Fish and Stuart, 1991).

The study site is situated on the Atlantic Coastal Ridge. The Atlantic Coastal Ridge is an 8 km wide low ridge of sand atop limestone formed along the eastern coast of Florida. It ranges in altitude from about 3 to 6 m above sea level in Miami Dade County but can be as high as 16m in northern Florida (McPherson and Halley, 1996). In the study area the ridge is made up of Pleistocene Limestone rocks which are highly permeable and lies at shallow depths with well developed secondary porosity due to dissolution of limestone (Evans and Ginsburg, 1987). The Miami Limestone, formed during a sea level high stand associated with the Sangamon Interglacial, is the predominant unit found at the surface and is made up of the cross-bedded and bioturbated oolitic facies beneath the ridge and bryozoan facies in the lagoonal environments west of the Atlantic Costal Ridge (Hoffmeister et al., 1967). The Fort Thompson Formation is made up of intercalated fresh and marine limestone and underlines the Miami limestone in Miami Dade County (Fish and Stewart, 1991).

Cave systems in present day Miami-Dade County includes caves at the Montgomery Botanical Center and Camp Owassar Baur and are all located on the ridge. Cressler (1993) showed that there are 22 large caves in the Miami metropolitan area as a result of the geomorphological features of the Atlantic Costal Ridge (ACR) and the

Everglades which leads to the formation of the cave systems. Most of the caves are associated with the old transverse glades (currently canal terrain) due to the slightly acidic nature of the dissolution and have vertical resolution up to 10ft in the unsaturated zones and deeper when filled with water (Cressler, 1993). The concentration of the cave systems along in the ridge shows that dissolution features are more likely to be found in the ACR.

4.3 METHODS

Geophysical field measurements were conducted from December 2011 to August 2012 using electrical resistivity and ground penetration radar (GPR). Resistivity techniques included 1-D azimuthal square array soundings, azimuthal 2-D profiles, and 3-D tomography. GPR techniques included common offset (CO) and common midpoints (CMP). Geophysical data was compared to existing hydrological data from a groundwater monitoring well (S-182) 1 km from the site and digital borehole images from two wells close to the study site (Figure 4.2).

4.3.1 Electrical Resistivity

The electrical resistivity method is based on inputting current into the ground via two current electrodes and measuring the potential difference across the subsurface through two additional potential electrodes. Electrical resistivity methods have been employed in numerous karst studies for cavity detection (Sumanovac and Weisser, 2001;

Roth et al., 2002; Seaton and Burbey, 2002; Gibson et al 2004; Vadillo et al., 2012). A detailed discussion of the various resistivity techniques and applications can be found in Loke et al., (2013). Briefly, in this study a total of three resistivity surveying techniques were used including: a) a 1-D square array sounding: performed to assess the variation of resistivity with depth, 2) an azimuthal Wenner 2-D survey: used to generate a 2-D image of the subsurface along different compass directions and investigate lateral variability of electrical properties of the subsurface, and 3) a 3-D resistivity survey: conducted to further investigate lateral variability to improve the interpolation of 2-D resistivity profiles in order to generate depth slices. All field resistivity surveys were conducted with an Advance Geoscience Incorporated (AGI) Super-Sting R1/IP 28-electrode resistivity imaging system. This single channel resistivity instrument system utilizes a multi-core cable, controlled by a programmable switch box which automatically determine the current and potential electrodes for each measurement.

4.3.2 One Dimension Resistivity Sounding

The azimuthal square array is an alternative electrode configuration for the measurement of electrical anisotropy where current and potential electrodes are deployed on opposite sides of a square (Habberjam and Watson, 1967). As the array is rotated through a series of angles (θ), the apparent resistivities vary in an elliptical pattern. Azimuthal resistivity arrays have been used to image features in karst regions (Hart and Rudman, 1997; Busby, 2000). The 28 electrodes were initially at equal angles on a circle forming 7 separate square array configurations rotated in increments of 12.87° at a radius

of 2.83m. The square size, A , was expanded in increments of $\sqrt{2}$ which resulted in sets of azimuthal measurements for square array sizes, A , of 4.0, 5.7, 8, 11.3, 16, 22.6, 32 and 45.3m to investigate the variation of anisotropy with depth. The equivalent effective depth, Z_e , of each azimuthal measurement is approximately one half the square size, A (Habberjam and Watkins, 1967; Edwards, 1977). The square array field data was inverted using the non-linear least square approach to determine the coefficient of anisotropy, λ , the mean resistivity, ρ_m , minimum resistivity direction, θ and associated parameter errors for each square array size, A (Yeboah-Forson and Whitman, submitted).

One dimensional square sounding was performed to assess the variation of resistivity with depth. This approach is routinely used to locate conductive targets such as the water table and steeply bedded fractures in karst systems as it provides good vertical depth resolution. In order to model with the EarthImage 1-D software, the azimuthal soundings, 'A' spacing's, were converted to their equivalent Wenner 'a' spacing's using the equations derived by Lane et al., (1995). Three (3) different layered models were generated in the Earth Imager 1-D using 5, 10 and 16 layers together with the default inversion parameters (depth factor of 1.1, thresholds (%) of 10, damping factor of 10, etc.). For example, the 16 layer model was parameterized with layer thickness ranging from 0.5 m at the surface to 3.9 m at 22 m depth. A smooth inversion technique was chosen to eliminate the need to guess starting values close to the actual formation resistivity (Turesson, 2006). The observed apparent resistivities (measured field data) were compared with the model (calculated) resistivity of 16 layer model for sensitivity analysis.

4.3.3 Two-Dimension Wenner Profiles

The 2-D Wenner profile was used to address the limitations of the 1-D models and to give a more accurate representation of resistivity changes in both vertical and horizontal direction which is a limitation in 1-D sounding (Loke and Barker, 1996). Five separate 2-D Wenner profiles were deployed at azimuths of 30, 65, 103, 129, and 167 degrees (Figure 4.3a). The center of these five profiles coincide with the center of the 1-D sounding method (Figure 4.3a). Electrodes were placed at 3m intervals for a total profile length of 81 m. Each profile consisted of a total of 118 measurements. All data was inverted using R2 resistivity modeling inversion software (Binley, 2012). R2 is a forward/inverse solution for 3-D or 2-D current flow in a quadrilateral or triangular mesh. The inverse solution is based on a regularized objective function combined with weighted least-square (Binley and Kemna, 2005). The Occam's algorithm which optimizes the smoothness in the model and eliminates the artifacts due to inherent non-uniqueness (Constable et al., 1987) is used in R2 inversion software. Generally in a 2-D inversion model the program works by generating an inverse model section, which is a tomogram representing the modeled depth and formation resistivities (Loke and Baker, 1996).

The model divides the subsurface into rectangular blocks and chooses optimum inversion parameters for the data which includes the damping factor, vertical to horizontal flatness filter ratio, convergence limit, and number of iterations discretize into a number of finite elements and resistivity blocks. The boundary conditions were set at infinity by adding finite elements and resistivity blocks outside the region of interest (Binley, 2012). In this study, a quadrilateral mesh was created with 4 nodes between

electrodes, leading to 0.5 m horizontal spacing. The vertical grids nodes were generated starting at 0.1 m thick at the surface increasing in thickness by a factor of 1.1 to a maximum depth of 13 m depth. The mesh for the boundary conditions was set to increase exponentially before 0 and after 81 m along the profile. Likewise the mesh extends past the maximum depth of 14 m.

4.3.4 Three-Dimension Electrical Resistivity Survey

This survey approach was employed to overcome the lack of detailed directional information from 2-D profiles so as to investigate the full extent of the anisotropic nature of dissolution features in the study area. All geological structures are 3-D in nature and a full 3-D resistivity survey and 3-D inversion model for interpretation should in theory give the most accurate results (Loke et al., 2013) as more data points are collected per measurement leading to better resolution of the subsurface (Bentley and Gharibi, 2004). Also a 3-D resistivity survey can improve upon artifacts and the resolution at depth which can provide better assessment of structures on a finer scale (Slater and Binley, 2003).

Data was collected using a mixed dipole gradient array. This array combines the radial dipole-dipole with the radial gradient array. As a result, the current and potential arrays are almost always on a straight line in the different grid orientation (Nyquist and Roth, 2005; AGI, 2012). The cable was laid out with X in the W-E and Y in the N-S directions respectively (Figure 4.3a). Electrodes were spaced at 6 m intervals, and a roll-along technique was employed to cover an area of 36 m x 42 m. The center of the 3-D survey was designed to match the azimuthal resistivity survey. A total of 708 data points

were modeled in EarthImager 3-D. EarthImager 3-D is a three-dimensional resistivity inversion program that interprets a three dimensional volume and two dimension sections of inverted resistivity data (AGI, 2012). The model was parameterized with a horizontal cell size of 1 m and layer thickness ranging from 1 m at the surface to 2 m at 8.75 m depth. Dahlin et al. (2002) showed that the use of a cell size smaller than 1 m does not significantly improve the accuracy for 3-D resistivity model. The inverse process converges at the 5th iteration with root mean square error of the correlation between the measured and modeled apparent resistivity of 4.8 % and a L2-Norm of 0.9.

4.3.5 Ground Penetrating Radar (GPR)

Ground penetrating radar (GPR) is an electromagnetic geophysical technique for subsurface exploration that uses a transmitter to generate electromagnetic (EM) waves (typically in a range between 10 MHz-2 GHz) that travel the subsurface and return to a receiver as a sequence of reflections. Reflections result from contrasts in dielectric permittivity (ϵ_r), a physical property that is highly dependent on water content. For that reason, the contrast between limestone matrix and dissolution features (whether full of water or air) represents a good target for GPR detection. The use of GPR has proved useful for describing the depth of penetration in the epikarst (< 20 m for saturated zone and < 30 m for unsaturated zone of karst system) and the infiltration zone of the karst aquifer where limestone is prevalent (Al-fares et al, 2002). Other studies have shown the utility of GPR for investigating shallow faulting, and cavity location in karst aquifers (Huisman et al., 2003; Annan, 2005). Application of GPR in the Biscayne Aquifer has

also proved effective for better understanding certain karst features. For example, Cunningham (2004) used GPR to characterize hydrogeologic properties such as the distribution of high porosity zones and to explore formation connectivity in the paleokarst of the Biscayne aquifer. Likewise, Grasmueck and Weger (2002) and Neal et al. (2008) employed 2-D and 3-D GPR method in many locations around Miami, Florida to better understand the complex oolitic sedimentary architecture and to reveal the complex internal structure of Pleistocene Miami Limestone.

GPR data in this study was collected with a Mala-RAMAC GPR with 100 and 200 MHz unshielded antennas. Two types of surveys were collected: a) common offsets (COs) with antenna separation of 1 m for 100 MHz and 0.6 m for 200 MHz; and b) common midpoints (CMPs) with 0.1 m spacing between traces in both cases. Two GPR CO profiles perpendicular to each other were collected following resistivity profiles A(129°) and D (65°) with five GPR CMPs distributed across the COs (Figure 4.3b). Stacking for both COs and CMPs was 32, with a time window of 500 ns. An average velocity of 0.11 m ns^{-1} was used to convert time to depth in COs as calculated from CMP surveys (as shown below). Furthermore one dimensional (1D) models of EM wave velocity were calculated from CMP surveys based on the application of Dix equation (Neal, 2004) which allows for estimating velocity variation with depth based on velocity intervals. The data processing routine for all GPR data consisted of: a) a “dewow” filter over a 10 ns or 5 ns time-window (for 100 and 200 MHz antennas respectively), b) application of a time-varying gain, c) a band-pass filter, and d) a static correction. All data processing was performed using ReflexW by Sandmeier Scientific Software.

4.4 RESULTS

4.4.1 Electrical Resistivity

Electrical anisotropy results are best presented in the form of polar plots. The plot of azimuthal resistivity survey is circular when there is little variation between the vertical axis (ρ_t) and the horizontal axis (ρ_l) exists. When anisotropy is present (i.e. $\rho_t > \rho_l$), an ellipse is formed with differences between the axes increasing as anisotropy increases (Habberjam, 1972). Polar plots of apparent resistivity from square arrays are shown in Figure 4.4 displaying different trends in anisotropy. The anisotropy for square size (A) of 4 m presents a very low coefficient of anisotropy ($\lambda = 1.06$) with the minimum resistivity direction of 119° (ESE). However, a higher magnitude of anisotropy was observed for A equal to 5.6, 8, 11.6 and 16 m with λ ranging from 1.18 to 1.36 with directions between 97 - 103° (E-ESE) (Figure 4.4b-e). An average anisotropy of 1.10 is found for directions between $(84$ - $93^\circ)$ E-W (Figure 4.4 f-g). In general, the coefficient of anisotropy ranged from 1.05 to 1.36 with a mean of 1.18. The highest anisotropy was found at square sizes corresponding to 8-16 m. The direction of minimum resistivity ranged from 84 - 119° with a mean of 96° trending E-W. The observed mean resistivity, ρ_m , is determined from the various square sizes and ranged from a high value of 742 Ω -m on the surface to 84 Ω -m at a square size of 45 m.

The results of the 1-D sounding model are presented in Figure 4.5 and show that the calculated apparent resistivity from the model is nearly identical to the measured observed resistivity. The variations in resistivity-depth prediction by the different layer models (5, 10, and 16) in Figure 4.5 all indicate a similar trend in resistivity changes with

depth. The modeled geoelectrical water table agrees with the water table determined from the monitoring wells at the time of the survey. The results from the model show a steep decrease in resistivity between 3-5 m below the surface, followed by a low resistivity layer ($55 \text{ } \Omega\text{-m}$) between 5-12 m and a higher resistivity at deeper depths. The depth model indicate higher resistivity ($>300 \text{ } \Omega\text{-m}$) on the surface to a depth of 3 m after which resistivity is less than $150 \text{ } \Omega\text{-m}$. The depth of this low resistivity zone ($55 \text{ } \Omega\text{-m}$) from the 1-D sounding corresponds well with the higher anisotropy observed in the anisotropy survey.

2-D Wenner results collected for each azimuth are shown in Figure 4.6. Profiles mostly show a high resistivity zone ($\rho > 400 \text{ } \Omega\text{-m}$) between the ground surface and about 3 m depth. This resistivity value is consistent with general observed resistivity values for limestone rocks in unsaturated zone. Below this zone resistivity generally decreases (reaching values of less than $150 \text{ } \Omega\text{-m}$) in the saturated zone. Additionally, low resistivity (highly conductive) features ($\rho \leq 40 \text{ } \Omega\text{-m}$) are observed below 4 m depth in all profiles (blue shading in Figure 4.6). However low resistivity areas are more pronounced along the 103° and 129° azimuth (Figure 4.6c and 4.6d), where very low resistivity ($\rho \leq 15 \text{ } \Omega\text{-m}$) values are found centered at 10 m, 45 m, and 60-65 m along the profile. These results are indicative of a low resistivity zone oriented in the E-W direction.

A set of six horizontal depth slices from the 3-D resistivity model obtained from EarthImager 3-D is shown in Figure 4.7. The depth slices were located at 0, 3, 5, 6, 7, and 8 m depth below the surface. Similar to the 2D results, high resistivity $\rho > 400 \text{ } \Omega\text{-m}$ is evident at the surface (i.e. 0-3 m depth). Below this surface layer a resistivity layer ($\rho <$

400 $\Omega\cdot\text{m}$) is observed. Low resistivity zones ($\rho \sim 100 \Omega\cdot\text{m}$) are generally found below 3m corresponding to the saturated zone. However in this zone very low resistivity (14 - 43 $\Omega\cdot\text{m}$) features trending in E-W direction are found at depths below ≥ 6 m.

4.4.2 Ground Penetration Radar

Results for the GPR survey from the common offset along Line A (Figure 4.3b), and the common midpoint profiles and their associated 1-D velocity models for locations 10, 25 and 45m are shown in Figure 4.8a and 4.8b-d respectively. Results from the common offset (Figure 4.8a) show a sequence of reflectors between 0-100 ns (corresponding to approximately 5.5 m depth assuming an average velocity of 0.11 m ns^{-1}) followed by a marked attenuation of the signal for depths below 100 ns. Two laterally continuous reflectors are also shown extending: 1) throughout the entire profile between 0-85 m and at approximately 2.75 m depth; and 2) between 25-40 m and 60-75m along the profiles and depths of approximately 4 meters. Results from common midpoints at 10, 25 and 42 m along the profile and their associated 1D models of velocity show some trends consistent with common offset results characterized by: 1) an area of high velocity ranging between $0.11\text{-}0.10 \text{ m ns}^{-1}$ from 0 to 3-3.25 m depth (in all CMPs, Figure 4.7b-d); 2) an area of low velocity ranging between 0.065 to 0.07 m ns^{-1} below 3.25 m depth (in CMP 10 and 42, Figure 4.8b and 4.8d); and 3) an area of lower velocity of approximately 0.095 m ns^{-1} between 3-4 m depth followed by an increase in velocity of about 0.11 below 4 m depth (in CMP 25, Figure 4.8c).

The common offset (shown as a non-migrated profile) depicted in Figure 4.8a also shows a series of diffraction hyperbolae distributed throughout the profile. Figure 4.9a depicts these hyperbolae and their associated velocities inferred by fitting. A two-dimensional model of velocity distribution throughout the profile is depicted in Figure 4.9b as based on velocity estimates from diffractions. It is important to note that velocities inferred for each particular diffraction represent average bulk values from the surface to the apex of that particular hyperbola. In contrast, estimates from 1D models in Figures 4.8b-d are not average bulk values but represent interval velocities for particular layers after Dix equation. Furthermore, higher density of diffractions are particularly visible in Figure 4.9a between 5-25 m and 40-60 m along the profile at depths between 3-6 m. Velocity estimates from these diffractions results in an overall velocity distribution (as shown in Figure 4.9b) characterized by: 1) an area of high velocity ranging between 0.11 to 0.12 m ns⁻¹ at depths of 0- 4 m; and 2) an area of low velocity (0.07-0.08 m ns⁻¹) between 0- 20 m along the profile at 4 to 6 m depth and 35-60 m along the profile at 3 to 6 m depth.

4.5 INTERPRETATION AND DISCUSSION

The resistivity results from the azimuthal survey, 1-D sounding, 2-D and 3-D inversion are in general agreement with one another. Evidence from a daily monitored USGS well S-182 (<http://my.sfwmd.gov/dbhydro>) less than 1km south of the site indicated that the water table varies between 2.4 to 2.9 m from the surface depending on the season. The data from the well shows that the thickness of the unsaturated zone in the

study area is around 3m depth. In the unsaturated zone ($\leq 3\text{m}$) all the data show a high resistivity zone ($\rho > 400 \Omega\text{-m}$). Lower resistivity ($\leq 150 \Omega\text{-m}$) is generally found below the water table. Since increased water content will tend to decrease electrical resistivity, the results fit well with the fact that limestone rocks above the groundwater table are characterized by higher resistivities than those in the aquifer (Milanovic, 1981). The GPR results are also consistent with this model. They depict a high velocity zone associated with an unsaturated zone for approximately the first 3 m (Figure 4.8 and 4.9). Furthermore, common offsets also show laterally continuous reflectors at about 2.75 m depth, which is consistent with the depth of the water table.

The 1-D square sounding (Figure 4.5) show resistivity changes with depth in a low resistivity zone (5-12 m) which corresponds to the effective depth of the higher anisotropy measurements. While a general decrease in resistivity is expected below the water table, a higher than usual change might be associated with porosity change due to the presence of solution cavities (Cardimona 2002; El-Qady et al., 2005). Both the 2-D and 3-D inversion results showed a similar trend of low resistivity zones. This is evident at 40-60 m along two of the 2-D profiles (Figure 4.6c, d) and in the 3-D survey at depth greater than 5 m. The low resistivity zones are particularly visible in the azimuth profiles in the E to SE trend as observed at depths greater than 5m in Figure 4.6 (c, d) 5-10 m and 40-60 m along the 2-D profiles. This is not evident in the other 2-D profiles with other azimuth orientations (Figures 4.6 a, b and e). Further evidence of this is found in the 3-D image where an E -W trending low resistivity zone is present (Figure 4.7). Hence the low resistivity area seems to trend in the E-SE direction as indicated by the resistivity data

GPR results also correspond well with resistivity results as related to low resistivity zones. For example, the 2D model of velocity inferred from GPR common offset Line 1 (Figure 4.9b) corresponds with the 2D resistivity cross-section following that same direction (i.e. 129° azimuth or Figure 4.6d), and depict two low velocity areas between 0- 20 m along the profile at 4 to 6 m depth and 35-60 m along the profile at 3 to 6 m depth. These GPR low velocity areas correspond well with low electrical resistivity values depicted in the 2D resistivity profile (Figure 4.6d). One important consideration when interpreting GPR results as based on EM wave velocity, relates to the saturation conditions and whether dissolution features are filled with water or air. For instance, under saturated conditions areas with enhanced porosity, dissolution features or cavities will most likely be filled up with water and therefore may induce lower EM wave velocity and potentially some attenuation of EM waves due to increased electrical conductivity (Lane et al., 2000). Likewise water-filled solution cavities will most likely result in low resistivity in resistivity surveys. Hence, low velocities in GPR under saturated conditions (i.e. below the water table) will most likely correspond to low resistivity (high conductivity) in resistivity surveys as related to water-filled solution cavities (Grandjean and Leparoux, 2004). In the unsaturated zone cavities will be most likely full of air and therefore the resistivity above the water table will likely result in a high resistivity (low conductivity) in the resistivity profiles and areas with high EM wave velocity in the GPR profiles. The presence of high resistivity zones in the shallow zone mostly at 4 m of the resistivity profiles (Figure 4.6) and the presence of high EM wave velocity between 0-3 m depth in GPR profiles (both from common midpoints in Figure 4.8b-d, and 2D velocity model in Figure 4.9b) confirms the correspondence between

resistivity and GPR results along the unsaturated portion of the column. Results for GPR common offset Line 2 (not presented here for brevity) did not show any major differences as compared to GPR common offset Line 1. Despite the good correspondence between GPR and resistivity surveys, further GPR profiling along different azimuthal directions or 3D surveys (Grasmueck and Weger, 2002; Neal et al. 2008) would be needed in order to further investigate whether areas of low EM wave velocity below the water table show a preferential direction.

Integration of our geophysical methods suggests the presence of dissolution features below the water table correspond with areas of: 1) high electrical conductivity in 2D and 3D resistivity surveys; and 2) low electromagnetic wave velocity in GPR common offset and common midpoint surveys. Furthermore, the presence of dissolution features may be acting as point reflectors and therefore result in the presence of diffraction hyperbolae in GPR common offsets (Neal, 2004). High density of diffractions are most likely the cause of the reflector depicted in Figure 4.7a at about 4 meters as it loses its lateral continuity between 0-25 m and 40-60 m along the profile. Although it is not entirely clear what that reflectors may represent (and considering the water table is slightly higher as previously described), changes in internal properties of the limestone (such as increased porosity) may be the most likely cause of a strong contrast in reflective properties (i.e. reflection coefficient). These interpretation matches well with nearby borehole images collected in previous studies that show certain dissolution features below the water table and at about 4 m depth (Figure 4.2a). Borehole images in the Biscayne Aquifer described in previous studies (Cunningham, 2004; Manda and Gross,

2006; Renken et al., 2008) shows evidence of solution holes in the saturated zones similar to those observed in the borehole images (Figure 4.2). Although high electrical conductivity may be associated with high water conductivity, pore fluid data from the wells G-3836 and G-3889 (Figure 4.2) showed pore resistivity (ρ_o) of 17.3-22 Ω -m, therefore indicating no evidence of salt water intrusion or contaminant at the site

The presence of dissolution features and cavities observed in the borehole images NE and SW of the site show the possibility of interconnected solution holes in the study area. These interconnected conduits would explain the high anisotropy oriented in the E-W direction. The direction of the minimum resistivity from the azimuthal resistivity survey which is interpreted as the hydraulic conductivity direction trends in the E or SE (Yeboah-Forson and Whitman, submitted). The site as noted sits in an area of the Atlantic Coastal Ridge where evidence of cave systems exist (Cressler, 1993). Hence, flow through the solution holes in the E or SE direction might have increased the dissolution rate in this direction. This is because paleo groundwater flow inferred from Fish and Stewart (1991) ; Cressler, (1993) and Fennema et al., (1994) and current groundwater modeling by South Florida Water Management District (SFWMD) indicated the general direction of groundwater flow is E and SE on the Atlantic Coastal Ridge towards the Biscayne Bay. On the regional scale the horizontal hydraulic anisotropy in the Biscayne Aquifer has been attributed to either the paleo groundwater flow or the different dissolution features of the rocks (Yeboah-Forson and Whitman submitted). If solution holes and torching vug features are present at the site then anisotropy would be higher due to these localized cavities. Numerous studies (Cunningham, 2004; Fish and Stewart,

1991; Renken et al., 2008) conducted in the Biscayne aquifer have shown that porosity features of the aquifer change rapidly both laterally and vertically from one location to another depending on the dominant matrix and/or touching vug porosity in karst. The touching vug zone was observed in G-3836 (Figure 4.2a) and approximately 4.5 m and 4 m in G-3689 (Figure 4.2b) with low permeability in the diffuse zones.

The nature of the vug-to-vug and vug-to-matrix connectivity has the tendency to increase the impact of the overall anisotropic flow in areas with significant dissolution cavities. Although the permeability and porosity nature of the Biscayne Aquifer has been attributed to matrix heterogeneity (Cunningham et al., 2006), it is possible that, along Atlantic Coastal Ridge, the interconnected nature of solution cavities might be highly anisotropic. This is not unique to our study site as the Miami Limestone and the underlying Fort Thompson formation have been shown to have solution holes at other locations in the aquifer (Cunningham et al., 2006; Cunningham, 2004; Manda and Gross, 2006; Renken et al., 2008). Understanding of the nature of the touching vugs and conduit in the Biscayne aquifer is still a work in progress, however evidence of a highly interconnected conduit system (and thus greater anisotropy) would enhance groundwater model development, which is particularly important when considering aquifer contamination in areas with a highly dense population. Ultimately, the geophysical methods and its correlation with borehole images have the predictive power to make fast characterization of subsurface anisotropic features.

4.6 CONCLUSIONS

The study presented here combines an array of electrical resistivity surveys with ground penetrating radar with a set of nearby direct borehole images to investigate the anisotropic natures of dissolution features within the Miami limestone. The azimuthal resistivity anisotropic data indicated a magnitude of anisotropy from 1.05 - 1.36 trending in the E-W direction in the saturated zone. The highest anisotropy was found at square sizes corresponding to effective depths of 4-8 m. Integration of our geophysical methods suggests the presence of geophysical zone with high conductivity below the water table at depths ≥ 4 m below the surface. The high conductivity zone derived from the ERI/GPR study matches the depth feature of the solution conduit derived from nearby borehole images providing further evidence that high conductive zones are most likely related to solution cavities in the aquifer. The cause of the higher anisotropy is attributed to the presence of solution cavities oriented in the E-W direction and situated at a depth >5 m below the surface. We hypothesize that preferential flow and enlargement of solution holes may be caused by the paleo groundwater flow. Despite the local nature of our survey, the approach can be easily applied to other areas with geological materials similar to the highly porous Miami limestone that may contain heterogeneous distributions of dissolution features or areas of preferential flow. Furthermore, these results show the potential of hydrogeophysical methods for quickly delineating the subsurface stratigraphy of karst environments.

4.7 REFERENCES

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Figure 4.1: The topographic map of study area in SE Miami-Dade County, Florida. Biscayne Aquifer (insert).

Figure 4.2: Digital borehole optical logs from nearby wells (A) G-3836 (located 2 km NE) and (B) G-3889 located 2.8km SW from the site showing macroporosity and solution features throughout the vertical thickness of the Biscayne Aquifer.

Figure 4.3: Site map showing the locations of the geophysical surveys at West Perrine Park. (a) Resistivity measurements for 1-D square array (red), 2-D profiles (green), 3-D (brown). (b) GPR common offset and common midpoint profiles.

Figure 4.4: Polar plots of square array apparent resistivity plotted against azimuth (deg). The thick solid line is the best fitting apparent resistivity function.

Figure 4.5: 1-D azimuthal equivalent Wenner modeling and analysis. (A) Observed and calculated resistivity from the 16th layer model. Error bars represent $\pm$ one standard deviation of the measured resistivity at each a spacing. B) Modeled apparent resistivity with depth for the 5(red), 10(green) and 16 (blue) layer models with water table determine from S-182 well. Model resistivities are nearly identical for each model.

Figure 4.6(a-e): 2- D Inverted cross-section for (a) 30°, (b) 65°, (c) 103°, (d) 129° and (e) 165°, azimuths, WT is the water table determine from USGS well 182.

Figure 4.7: Horizontal depth slice of EarthImager 3-D inverse model.

Figure 4.8: a) GPR common offset in Line 1 using 200 MHz antennas. Arrows indicate location of common midpoint (CMP) surveys. Common midpoint profiles and inferred one-dimensional (1D) model of velocity (after Dix equation) for locations: b) 10 m; c) 25 m; and d) 42 m along common offset in Line 1 using 200 MHz antennas.

Figure 4.9: a) GPR common offset in Line 1 showing a representative sample of diffraction hyperbolas and associated velocities inferred. Arrows indicate location of common midpoint (CMP) surveys for reference; b) two-dimensional (2D) model of velocity inferred from interpolation of diffraction hyperbolas ($n=50$) in a) using a x -weight equal to 1. Note that velocities at each point represent average values from the surface to that point (i.e. no Dix calculated values).

5 GEOPHYSICAL FLOW ANALYSIS OF ANISOTROPY: A CASE STUDY OF SNAPPER CREEK MUNICIPAL WELL FIELD.

ABSTRACT

A case study of hydrogeophysical measurement of anisotropy and preferential permeability in an active hydrological site with continual pumping and canal network is presented. Electrical anisotropy determined at the site as part of a regional study indicates that anisotropy changes from NNW-SSE to NE-SW with depth. Subsequently one, two, three dimension resistivity and azimuthal self potential gradient (ASPG) techniques were employed together with well data to map the subsurface at Snapper Creek Municipal Well Field, in Miami, Florida. The resistivity sounding model showed a decrease in resistivity with depth between 2 and 4 m below the surface, followed by a higher resistive layer at depth. The two and three dimension imaging illustrates higher resistivity at the surface in the unsaturated zone and a decrease in the saturated zone with significant horizontal and vertical changes due to low resistivity zones. The variation is attributed to the solution cavities at the site similar to those evidenced by monitoring well lithology. The measured self potential during pumping and non-pumping of groundwater changes on the average by 3 mV. The estimated potential polarities indicate that flow direction trends SSE which is similar to the trend of azimuthal resistivity survey from the surface to 4 m and the direction of the surface water in the canal at the site. The results demonstrate that the change in hydraulic anisotropy might be related to solution cavities, the surface canal and the groundwater extraction wells. This study shows the potential for

hydrogeophysical measurement as a useful tool in providing information about the anisotropy in areas of complex surface and groundwater interaction.

5.1 INTRODUCTION

Management and protection of water resources in surficial aquifer systems require detailed knowledge of the interaction between groundwater and surface water. Groundwater and surface water are not isolated components of the hydrologic system, but instead interact in a variety of physiographic and climatic landscapes (Sophocleous, 2002). In South Florida, water management activities have increased the extent of interaction between surface water and groundwater over the past century through a network of canals (McCormick et al., 2011). These canals originally constructed to drain the wetland south of Lake Okeechobee, have altered the timing, connectivity, movement and direction of surface water and ground water.

Surface water and groundwater interaction in the Biscayne Aquifer especially in the Everglades and natural areas (Taylor Slough and Tree Islands), has been extensively studied (Harvey and McCormick, 2009; Larsen et al., 2011; Sullivan et al., 2012). However, the relationship between groundwater and surface water is less studied and understood in urbanized and highly populated areas in eastern Miami-Dade County, where numerous canals, water-control structures, and municipal well fields exist. The relationship between groundwater and surface water is often investigated by using techniques such as isotope analysis and numerical models which are constrained by hydraulic data required to simulate the entire flow and leakage between canals and

groundwater (Swain, 2012). However, the complexity of the shallow subsurface and lack of aquifer characterization models lead to failures of inaccurate representation of field conditions. Alternatively, hydrogeophysical investigations using geoelectrical methods can provide specific parameter information about the hydrogeological makeup of surface and groundwater systems. Despite the disadvantages associated with geophysical methods which include ambiguities and uncertainties in interpretation due to their indirect nature, there are many advantages including quick survey at relatively cheap cost, larger area coverage and non-invasiveness, which makes it attractive for subsurface investigation in sensitive hydrogeological environments (Reynolds, 2010).

In a recent regional study presented in chapter 2 of this thesis, electrical anisotropy was investigated at several sites in the Biscayne Aquifer. The direction of anisotropy was uniform with depth at most sites. However, at one of the sites located at the Snapper Creek Municipal Well Field (Figure 5.1), the direction of electrical anisotropy changes abruptly with depth after 4 m below the surface from NNW-SSE to NE-SW. This paper investigates the causes of this variation in the direction of electrical anisotropy and also examines the role of groundwater extraction wells or canals, if any, on the observed trends in anisotropy direction with depth. The study employs geoelectrical techniques (resistivity and self potential) with digital borehole images and well data to characterize, image, delineate subsurface features and measure hydraulic flux. The uniqueness of this case study is to examine the effectiveness of self potential method to predict preferential flow path during pumping and non pumping periods at the well field. The study has implications for groundwater and surface water interactions and development of accurate hydrogeologic models.

5.2 HYDROGEOLOGICAL SETTINGS

The Snapper Creek Municipal Well Field (SC) is bounded by latitude 25.700° N and Longitude 80.359° W in Miami-Dade County of South Florida (Figure 5.1). The surface elevation of SC is 1.45 m (NAVD 88) and it sits on the western margin of the Atlantic Coastal Ridge where elevations can reach 6 m above sea level. The site was chosen because it is a microcosm of the current man-made influence on the Biscayne Aquifer with a canal traversing the site and groundwater extraction wells on either side of the canal.

The site geology consists of the Pleistocene Miami Limestone on the surface underlain by the Fort Thompson Formation in the Biscayne Aquifer, with the base of the aquifer sitting on top of the Tamiami Formation (Wacker et al., 2013, in review). Lithological evidence obtained from four South Florida Water Management District (SFWMD) monitoring wells (C2GSW1_GW1, C2GSW1_GW2, C2GW1_GW1 and C2GW1_GW2; see Simon et al., 2007, Appendix B for further details) showed a mixture of some vugs with light brown quartz sand (fine to medium rounded grains, poorly sorted from the surface to 4 m depth) followed by light brown oolitic limestone (grainstone) with moldic and vuggy porosity at greater depth (Simon et al., 2007). In general, the limestone rocks in the Biscayne Aquifer are very porous and have very high hydraulic conductivity, resulting in the productivity of the extraction wells which supply water needs for the inhabitants of Miami-Dade, Broward and Monroe Counties in South Florida (Fish and Stuart 1991).

Snapper Creek Well Field (Figure 5.1) is part of Alexander Orr, Jr. Subarea Well fields with 4 pumping wells. These four active wells, SC 21-24 (Figure 5.1), were

constructed in 1976 to boost the water supply of Miami-Dade County. Since its construction, at least one of the wells has been in operation at any particular time except when a major maintenance is required or during few occasions of hurricane impact. The wells are 24 inches (0.61 m) in diameter, 108 feet (32.9 m) deep and have casing depths of 50 feet (15.24 m). The total well field capacity is 40.0 mgal/day (151ML/d) with 10 mgal/day for each well. The Snapper Creek Well Field is divided north and south by the C2 Canal (Figure 5.1). The C2 Canal and others like it in South Florida were constructed as drainage and flood control facilities.

Prior to the onset of effective man-made alteration to the hydrology in 1907 (Blake, 1980) and the completed construction of the canals in Miami in 1937 (McCally, 1999), groundwater flow direction was generally seaward, towards Florida Bay in the SW or Biscayne Bay in the E-SE. The post-development groundwater flow direction in Miami, and South Florida in general, is controlled by man-made features (canals, levees, pumps, wells, etc.) in eastern Miami-Dade County (Fish and Stewart, 1991). Hence, the local variations in the water table are due to other causes, such as local topographic highs (e.g. Atlantic Coastal Ridge), or large-scale withdrawals from major well fields and canal operations (Simon et al., 2007). Currently at SC Well Field, the local groundwater flow is in the NE direction toward the Alexander Orr Water Treatment Plant while the Surface water flow is in the SSE. The interaction of surface water and groundwater along these canals in the vicinity of public supply wells is the subject of many scientific and technical studies to determine whether the canal are losing water to the groundwater as a result of the pumping (Simon et al., 2007; Swain, 2012; Wacker et al., in review).

5.3 INSTRUMENTATION AND SURVEY DESIGN

Data collection was done using an Advanced Geoscience Incorporated (AGI, 2013) Super-Sting R1/IP 28-electrode resistivity imaging system. This single-channel resistivity instrument system utilizes a multi-core cable controlled by a programmable switch box, which automatically determines the current and potential electrodes for each measurement. Capabilities of this instrument include surface 2-D survey, 3-D survey, and time-lapse survey (AGI, 2012). The overall geophysical program consisted of the following consecutive tasks:

1. Azimuthal resistivity survey (ARS): to determine electrical anisotropy and resistivity variation with depth.
2. Detailed resistivity using 2-D Wenner survey: to generate 2-D images of the subsurface along multiple profiles to investigate lateral variability of electrical properties and delineate subsurface features
3. 3-D electrical imaging: to locate the zone of resistivity anomalies and further improve the interpretation of 2-D resistivity profiles
4. Azimuthal Self Potential Gradient (ASPG) survey: to detect the preferred direction of groundwater flux at the site during pumping and zero pumping from the extraction wells.

5.3.1 Azimuthal Resistivity Survey

The azimuthal square array is an alternative electrode configuration for the measurement of electrical anisotropy where current and potential electrodes are deployed

on opposite sides of a square (Habberjam and Watson, 1967). As the array is rotated through a series of angles (θ), the apparent resistivities vary in an elliptical pattern. Twenty-eight electrodes were initially deployed at equal angles on a circle forming seven separate square array configurations rotated in increments of 12.87° at a radius of 2.83 m from the center (Figure 5.2). The square size, A , was expanded in increments of $\sqrt{2}$ and resulted in sets of azimuthal measurements for square array sizes of 4.0, 5.7, 8, 11.3, 16, 22.6, 32 and 45.3 m to investigate the variation of anisotropy with depth. The equivalent effective depth, Z_e , of each azimuthal measurement is approximately one half the square size, A (Habberjam and Watkins, 1967; Edwards, 1977). The square array field data was inverted using a non-linear least square approach to determine the coefficient of anisotropy, λ ; the mean resistivity, ρ_m ; minimum resistivity direction, θ ; and associated parameter errors.

5.3.2 1-D Sounding

1-D square sounding was performed to assess the variation of resistivity with depth. This approach is routinely used to locate conductive targets such as the water table and steeply bedded fractures in karst systems as it provides good vertical depth resolution. In order to model with the EarthImager, the azimuthal soundings, ' A ' spacings, were converted to their equivalent Wenner ' a ' spacings using the equations derived by Lane et al. (1995). Three (3) different layered models were generated in the Earth Imager 1-D using 5, 10 and 16 layers together with the default inversion parameters (depth factor of 1.1, thresholds (%) of 10, damping factor of 10, etc.). For example, the 16-layer model

was parameterized with layer thickness ranging from 0.5 m at the surface to 3.9 m at 22 m depth. A smooth inversion technique was chosen to eliminate the need to guess at starting values close to the actual formation resistivity (Turesson, 2006). The observed apparent resistivities (measured field data) were compared with the model (calculated) resistivity of the 16-layer model for sensitivity analysis. The azimuthal measurements at each square size, A , were averaged and modeled to assess the variation of resistivity with depth.

5.3.3 Wenner Two Dimensional (2-D) Imaging

The Wenner Array was used to generate a 2-D image of the subsurface along different profiles to assess lateral variability of electrical properties of the subsurface. Originally, a 2-m spacing cross profile was collected diagonally in the layout shown in Figure 5.2. Subsequently, four parallel lines oriented N–S and 18 m apart and four parallel lines oriented W–E at the same spacing were surveyed together forming a square of side 54 m (Figure 5.2). The data collected from the layout (Figure 5.2) produced 8 different profiles, 4 in the N-S direction and 4 in the W-E direction. These were imaged at depths 0 to 10 m along a 54 m profile. The profiles were acquired during pumping of wells SC 23 and 22 with low water table conditions (-0.72 m, NAVD 88). All data was inverted using R2 resistivity modeling inversion software (Binley, 2012).

5.3.4 Three Dimensional (3-D) Imaging

Detailed subsurface analysis of heterogeneity at the site was performed using the 3-D electrical resistivity tomography. The goal of the 3-D data (collected a day after the

2-D) was to get a full tomography of the subsurface and image any complex 3-D features which were not captured by the 2-D profiles. This is because the 2D profiles generated from the models are limited and cannot provide detailed directional information. All geological structures are 3-D in nature and a full 3-D resistivity survey and 3-D inversion model for interpretation should, in theory give the most accurate representation of the the geologic structure (Loke et al., 2013) as more data points are collected per measurement leading to better resolution of the subsurface (Bentley and Gharibi, 2004).

A field scale 3-D resistivity survey was conducted using the mixed dipole gradient array. This is a mixed radial dipole-dipole (the current and potential electrodes A, B, M and N are either on the same line or across four lines straight line in a 3-D rectangular grid) and radial gradient array (the potential electrodes may be between or outside two current electrodes AB) (Nyquist and Roth, 2005; AGI, 2012). As a result, a large number of measurements are generated with stronger signal leading to higher model resolution (AGI, 2012). The survey covers an area of 54 m by 54 m. The 28 electrodes were placed at 9 m increments along four parallel lines spaced nine meters apart. After measurements were recorded, the array was shifted (rolled along) two lines to the right producing 50% overlap on the previous array. Finally, the array was rolled along a third time. In total, 1,038 measurements were collected. The data were modeled in EarthImager 3-D (AGI, 2013). The model was parameterized with a horizontal cell size of 1 m, and layer thicknesses increased from 1 m at the surface to 3 m at 13.75 m depth. The inverse process converged at the 8th iteration with root mean square error (RSME) of 14.4% and an L2-Norm of 14.28.

5.3.5 Azimuthal Self Potential Gradient (ASPG)

In order to visualize the relationship between the predicted direction of measured potential and groundwater characteristic, the azimuthal self-potential gradient (ASPG) technique proposed by Wishart (2008) was applied at the site. The ASPG approach is based on a collinear electrode array over azimuthal spread where two electrodes are kept at fixed distance from each other and rotated simultaneously at specific angles through 360°. The purpose for the self-potential survey was to determine the flow direction at the site when the wells were turned on and off. Self or streaming potentials can arise from different sources, including charge separations as a result of coupled electrokinetic fluid flow via applied pressure gradient (Rizzo et. al., 2004 and Wishart et al., 2006). During pumping, the flow of groundwater is responsible for a measurable potential at the ground surface owing to the electrokinetic coupling between the Darcy velocity and the electrical current density (Rizzo et. al., 2004). Revil et al. (2002) showed that the streaming potential is always positive along flow direction. In an azimuthal self-potential gradient survey, the direction of the groundwater is reflected in the positive potential polarities from the gradient measurement (Wishart et. al., 2006).

Since non-polarizing electrodes are required for SP studies, two AGI non-polarizing electrodes were used for the survey. To eliminate positional effects and recapture exact spots, the center spot of the survey was referenced to magnetic north and marked with flags for the duration of the survey. The ASPG was rotated at 25.7° and careful effort was made to reoccupy the same spot to record the azimuth. Three sets of measurement were collected for each survey. The baseline (first, NP) survey was conducted when all the four pumping stations were turned off (Figure 5.1). Subsequently

SP data set (2, P1) was collected when pumping wells SC 21 and SC 22 were in operation. Finally, SP data set (3, P2) was obtained when SC 21 and 23 were in operation.

5.4 RESULTS AND INTERPRETATION

5.4.1 Azimuthal Resistivity Survey

The results of apparent resistivity from the square array survey are shown in Figure 5.3. Electrical anisotropy results are best presented in the form of polar plots. The plot of azimuthal resistivity is circular when there is little variation between the transverse resistivity (ρ_t) and the longitudinal resistivity (ρ_l). When anisotropy is present (i.e. $\rho_t > \rho_l$), an ellipse is formed and the differences between the axes increases as anisotropy increases (Habberjam, 1972). Figure 5.3a shows measurements of very low coefficient of anisotropy (isotropic) where the azimuthal pattern is nearly circular. In contrast, observed polar plots in Figures 5.3b-h indicate a pattern with clear preferential direction associated with anisotropy. The coefficient of anisotropy generally ranged from 1.01 to 1.08, with a mean of 1.06. The lowest anisotropy is associated with the smallest A-spacing. The mean resistivity ranged from a high value of 338 ohm-m in the unsaturated zone to 104 ohm-m in the saturated zone. The direction of minimum resistivity, θ , exhibits different orientations at the site. For square sizes of 4-8 m (effective depth Z_e , 2-4 m), the minimum resistivity trends around 162° but trends 50-65° for square sizes greater than 11m ($Z_e \geq 5$ m). This trend indicates that electrical

anisotropy varies with depth at the well field. The change with depth is abrupt and occurs between 4 and 5 m of depth.

5.4.2 1-D Sounding

The results of the 1-D sounding model presented in Figure 5.4 show that the calculated apparent resistivity from the model is nearly identical to the measured observed resistivity. The variations in resistivity-depth prediction by the different layer models (5, 10, and 16) in Figure 5.4 indicate a similar trend in resistivity changes with depth. The modeled geoelectrical water table which is defined as resistivity less than 220 Ω -m is below the surveyed water table determined from the monitoring wells on the day of the geoelectrical measurement. This is not surprising as the water table in the well field changes rapidly in response to the groundwater extraction rate. The results from the model showed a steep decrease in resistivity (40 Ω -m) between 3 and 5 m below the surface followed by a higher resistivity layer ($\sim$ 100 Ω -m) at deeper depths. While a general decrease in resistivity is expected below the water table, a higher-than-usual change might be associated with porosity change due to the presence of solution cavities (Cardimona 2002; El-Qady et al., 2005). The depth of this low resistivity zone from the 1-D sounding corresponds well with the anisotropy change in direction from NNW-SSE to NE-SW below this zone.

5.4.3 Wenner Two Dimensional (2-D) Imaging

Figure 5.5a-c shows a diagonal 2 m spacing and 1 m spacing survey in the study area. The survey was conducted along two profiles trending NE-SW and NW-SE with depths 15 m and 6m and lengths 27m and 54 m respectively. In both Figure 5.5a and b, evidence of a low resistivity zone less than 30 Ω -m was found around 2 m depth followed by a higher resistivity (100 Ω -m) 4 m below the surface. The low resistivity zone below the water table is consistent with the low resistivity zone in the 1-D model.

Figure 5.6a shows the inverted resistivity sections along the N-S direction. The profiles generally have high resistivity of 200 Ω -m or greater close to the surface, which is the unsaturated zone. The thin high resistivity zone near the surface is underlain by resistivity less than 175 Ω -m in the saturated zone. Additionally, a few very low resistivity zones (<10 Ω -m) were observed on the profile, which might be interpreted as pockets of solution cavities. The W-E profiles show evidence of low resistivity (<5 Ω -m) at 15 m along Y0 and Y18 sections. However, this finding is not visible on the on Y36 and Y54 profiles, indicating the absence of a long-connected feature (Figure 5.6b). These observations might indicate solution cavities due to the known vuggy and moldic porosities at the site.

In general, there appears to be a low resistivity channel at ~ 3 m depth on X0, Y0, X18, Y18, and Y36 (Figure 5.6). The results indicate a shallow low resistivity zone below the water table that is consistent with the low resistivity zone in the 1-D model. However, assuming Y0 and Y18 are accurate descriptions of the geology, then, there are substantial differences between the N-S and W-E profiles in the saturated zone. This might explain the electrical anisotropy variations with depth observed in the data from

the ARS survey. In terms of depth, the resistivities decrease with depth below the water table.

5.4.4 Three Dimensional (3-D) Imaging

Depth slices of the inverted 3D resistivity tomogram are shown in Figure 5.7. The depth slices were located at 0, 2, 4, 6, 8, and 10 m depth below the surface. The results indicate higher resistivity ($>200 \Omega\text{-m}$) as evidence in depth slices ($Z=0$ and 2 m) in the unsaturated zone. On the other hand, below 4 m resistivity less than $175 \Omega\text{-m}$ was observed in the saturated zone (Figure 5.7). Evidence of low resistivity zone (blue $<30 \Omega\text{-m}$) is observed in the depth slices. This is consistent with the low resistivity zone observed in both 1-D and 2-D resistivity surveys below 6 m depth. The root mean square error and L2-Norm is higher than expected for the inversion as evidenced by the distribution of the observed and calculated resistivity shown in Figure 5.8. The lack of total agreement in Figure 5.8 is due to the inability of the EarthImager 3-D software to run at a finer grid and mesh sizes owing to memory limitations.

5.4.5 Azimuthal Self Potential Gradient (ASPG)

Figures 5.9 and 5.10 show the ASPG results from the gradient survey. The maximum SP anomaly measured was 2.2 mV and correlates with 135° (SE) during pumping scenario P1 while the minimum potential of -1.4 mV observed at 180° (S) was during for NP (baseline) (Figure 5.9). Also the magnitude of the measured potential for P1 was not significantly different from P2. However, there was change of 3 mV was observed between the baseline and pumping scenarios. The directions of self-potential

measured for the pumping scenarios are shown in polar plots (Figure 5.10). Polar plots of SP data indicate flow direction ranging from SE to S with a maximum flow direction of 155° (SSE) (Figure 5.10). This direction is similar to the azimuthal resistivity direction from the surface to 4 m below the surface but different from those greater than 4 m. Also, the trend of the SP generally corresponds to the orientation of the canal at the site (Figure 5.1), which trends in the SSE direction.

5.5 DISCUSSION

The geoelectrical measurements applied in this study were able to provide quantitative and qualitative information about the anisotropic behavior of the subsurface at the Snapper Creek Well Field. Overview of resistivity data indicates that at the well field, electrical anisotropy ranged from 1.01 to 1.08 (which is similar to those measured in the regional study) and the direction of minimum resistivity changes with depths between 4 and 5 m. The one-dimensional sounding investigation showed a steep decrease in resistivity between 2 and 4 m below the surface followed by an increase in resistivity at ~ 5 m, which corresponds to the change in direction. Similarly a thin low resistivity zone at ~ 3 m below the surface is observed along the 2-D profiles in the saturated zone. Furthermore, the 3-D depth slice analysis showed a major change in resistivity before and after 4 m below the surface. The change in resistivity zones was interpreted as porosity variation below the water table. Lithological evidence from the monitoring wells at the site by Simon et al., (2007) indicates major porosity changes due to cavities between the surface and 4 m below the surface. For example data from monitoring well

C2GSW1_GW2 indicate a tan to white limestone (wackestone) with both moldic and vuggy porosity between 10-15 ft (~3.5 to 5 m). Similarly, interpretation from monitoring well C2GW1_GW1 show the presence of very permeable olive brown to white limestone with vuggy and moldic porosity between 10 - 20 ft (~3.5 to 6.5 m) below the surface (Simon et al., 2007).. These water-bearing cavities with vuggy and moldic porosities at the site might be a key indicator of the directional change observed from the geoelectrical survey at the site.

Besides the likely scenario that the change in anisotropic direction is caused by vuggy and moldic porosities at the site, there is also a theory that the abrupt change in the direction between 4-5 m below the surface might be due to the influence of the canal at the site. The NNW-SSE anisotropy direction at the surface is probably highly influenced by the canal which is oriented in the same NNW-SSE direction and the prehistorical flow at SC is more in line the NE-SW direction observed from ARS. Evidence supporting this theory came from the fact that SC is west of the Atlantic Coastal Ridge and, prior to the man-made alteration to South Florida hydrology, most of the flow west of the ridge was toward the Florida Bay, SW of the study area (Cressler, 1993). However, Swain (2012) concluded that water level in the C2 canal is not a local phenomenon due to the hydraulic interconnectivity of the canals system around the study area.

Alternatively, assuming historical flow was SSE towards the Biscayne Bay, pumping for the past 4 decades may have altered the preferential hydraulic conductivity direction below 4 m depth where the full impact of extraction is felt. Swain (2012), using stochastic analysis to identify wellfield withdrawal effect on surface water and groundwater in the SE of the study area concluded that although the canal acts as a

barrier between the north and south pumping wells, it is ultimately affected by water withdrawal from the extraction wells. Hence, the change in direction of anisotropy might be due to the extraction wells which impact the anisotropic direction in the study area below 5 m depth.

The magnitude of the measured potential difference of 3mV from the ASPG techniques is lower compared with similar streaming potentials associated with pumping tests which measure a maximum of difference baseline and pumping to be ~5 mV (Rizzo et al., 2004). The low measured potential might be due to the effects of the canal surface water at the site. At Snapper Creek, the water level in the canal which is around 2 m below surface is always higher than the water level of the aquifer, as groundwater levels are affected by well field withdrawals while surface water levels are not (Simon et. al., 2007). Evidently even at peak pumping, the measured natural potential might be influenced more by the low surface water hydraulic gradient than the expected high gradient potential from groundwater wells screened at 15 m below the subsurface.

Nevertheless, the direction of flow does not depend on the magnitude of the measured potential but on the polarity. The results from the ASPG (which measure flux) directions matched that of the ARS (which measured the minimum resistivity direction in the geological unit) in terms of predicted direction from the surface to an effective depth of 4 m, which trends NNW-SSE. The geoelectrical predicted direction is consistent with the direction of the surface water in the C2-canal at the site. Hence, it is highly likely that the gradient directional signals picked up the SP method might be surface water and not groundwater gradient as intended in this study.

Geologically, the development of preferential pathways for hydraulic conductivity takes thousands of years. Even in the large solution interconnected limestone in the Biscayne Aquifer, the study questions if a change in preferential geological direction can occur in less than four decades of pumping and a century of active canals at the site. Based on the findings of this study, a clear distinction on the cause(s) of anisotropic change with depth from geoelectrical data alone is not adequate to resolve the ambiguity relating surface water and groundwater interactions in active hydrological zones where there is continual pumping and a canal network. The process that developed the electrical anisotropy features observed at Snapper Creek Well Field cannot be easily explained from this hydrogeophysical study without detailed geological evidence from digital borehole images and geological cross-section which is unavailable at the moment.

5.6 SUMMARY

The impact of water extraction on the interaction of surface water and groundwater along the canals in south Florida is an ongoing scientific research subject and this study uses hydrogeophysical concepts to contribute to knowledge of the subject area. The study presented here evaluates the characteristics of the subsurface for spatial variability in anisotropy with depth using geoelectrical measurements and well data in an active hydrological zone where there is surface water and groundwater interaction due to continual pumping and a canal network.

Results from azimuthal resistivity data indicates that at the well field, low electrical anisotropy range from 1.01 to 1.08 and the direction of minimum resistivity

changes with depth from NNW-SSE in the upper 4 m and NE-SW at depths above 5 m. 1-D sounding analysis using 5, 10 and 16 layer model showed a steep decrease in resistivity between 2- 4m below the surface followed by a higher resistive layer at depth. The 2-D Wenner profiles and 3-D imaging revealed a higher resistivity at the surface and decreased in the saturated zone with significant horizontal and vertical changes. Similarly, a thin low resistivity zone at ~ 3m below the surface is observed along the 2-D profiles in the saturated zone. Furthermore the 3-D depth slice analysis showed a major change in resistivity before and after 4 m below the surface. These changes are attributed to presences of solution cavities at the site; these presence are supported by lithological evidence from monitoring wells and unpublished cross-section from digital borehole images across the site. The ASPG data showed that the direction of flow was towards SSE which is similar to the direction of the surface water and measured anisotropy from the surface to depth of 4m.

The study hypothesizes that the change in electrical anisotropy direction with depth is due to (1) the complex relationship between the permeable vuggy and moldic rocks at the site, (2) the effect of the surface water flow in the canal constructed a century ago or (3) groundwater behavior due to the extraction wells over the past four decades. However, the exact impacts of these factors could not be fully resolved by hydrogeophysical data sets derived from this study. Thus, further work which compares this study with core analysis, induction log and isotope analysis at the site should determine the level of impact by the canals and the extraction wells (if any) on anisotropy at the Snapper Creek Municipal Well Field.

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Figure 5.1: Map of the study area with Insert map of Florida.

Figure 5.2: Layout of the schematic view of different (a) N-S Profile 2m Spacing resistivity.

Figure 5.3: Polar plots of square array apparent resistivity plotted against azimuth (deg). The thick solid line is the best fitting apparent resistivity ellipse.

Figure 5.4: Vertical electrical sounding model in the study area A) observed and calculated resistivity from the model. Error bars represent $\pm$ one standard deviation of the measured resistivity at each a spacing. B) Modeled apparent resistivity with depth for the 5(red), 10 (green) and 16 (blue) layer models with water table determine from Monitoring well. Model resistivities are nearly identical for each model.

Figure 5.5(a-c): 2 m spacing diagonal inversion (a, b) and 1-m spacing (c) Inverse model resistivity section.

Figure 5.6(a-b): 2- D Inverse model resistivity section along the a) North- South direction and b) East–West direction.

Figure 5.7: Distribution of Measured and Calculated Resistivity from EarthImager 3-D Model.

Iteration No. 8. RMS = 14.4%. L2 = 8.3

Figure 5.8: A plot of the correlation between measured and modeled apparent resistivity

Figure 5.9: Azimuthal Self Potential Gradient at Snapper Creek Well Field. NP-(no pumping at the site), P1 (SC 21 and SC 22 were in operation) and P2 (SC 21 and 23 were in operation).

Figure 5.10: Polar plot of azimuthal self-potential gradient for P1(blue) and P2 (red).The plot is for positive values of streaming potential recorded which interpretive at the flow direction during pumping.

6 SUMMARY, CONTRIBUTIONS AND FUTURE WORK

In urbanized or protected natural areas of south Florida the use of minimal invasive geophysical methods for estimating quantitative information about hydraulic properties of surficial aquifer systems cannot be overemphasized. Advantages of these methods over single-point measurements (e.g. hydraulic wells, direct core sampling) includes cost effectiveness, fast and continuous data collection and ability to capture larger survey area. The case studies in this dissertation involve the use of multiple geophysical surveys constrained by borehole images, well data, and geological data to provide accurate groundwater information in order to understand the surficial Biscayne Aquifer in SE Florida. A better understanding of hydrogeologic process and parameters of this complex aquifer system would help in the overall protection and management of this vital water resource.

This chapter reviews the findings highlighted in Chapters 2 to 5. The main findings and conclusions for each of the objectives mentioned in section 1.4 of the dissertation are individually summarized (section 6.1), and later integrated within a broader context (Section 6.2) of the overall study. Also, a brief discussion on the significance of the study to the Biscayne Aquifer (Section 6.3), contribution to the field of Hydrogeophysics (Section 6.4) and recommendations for future research (Section 6.5) are presented.

6.1 SUMMARY OF CONCLUSIONS

In Chapter 2, azimuthal resistivity measurements were inverted for the mean resistivity, coefficient of electrical anisotropy and minimum resistivity direction for sites in a regional study. Regression models were developed using measured electrical parameters and hydrogeological data to predict the behavior of the hydraulic properties such as secondary porosity, the principal components and direction of the hydraulic conductivity tensor. The anisotropy is greatest in the central portions of the Atlantic Coastal Ridge (which plays a critical role in the paleo groundwater flow) with the maximum hydraulic conductivity trending E-W/SE-NW at most sites. In regions west of the ridge, the hydraulic anisotropy is lower and trends SE-NW and E-W. This is consistent with predevelopment groundwater flow. The chapter provides a reasonable baseline estimate of hydrological parameters such as anisotropy, secondary porosity and conductivity tensors for the Biscayne Aquifer. The multi-electrode square array technique employed in this study is able to effectively measure even small anisotropy values with very high accuracy.

In Chapter 3, the depth, thickness and resistivity and extent of porosity were investigated through 1-D sounding constrained by well data in the Biscayne Aquifer of eastern Miami-Dade County. The formation resistivity was determined from three different layered 1-D models to generate a depth map of resistivity and porosity of the saturated zone. Porosity was estimated using Archie's law and pore fluid data from nearby wells at the study sites. The resistivity depth profile was uniform in the saturated zone for most of the study sites. Resistivity of the saturated zone decreased from a higher of 250 Ω -m in the NW to 1 Ω -m in the SE along the coastline in the saltwater zone. The

estimated porosity in the saturated zone ranged from 18 to 60% with a mean of 30%. This compare favorably with other porosity data from published studies on the aquifer. The variation in formation resistivity with depth is attributed to either change in porosity due to solution cavities in the fresh water zones or saltwater intrusion along the coastal areas surveyed.

Surface resistivity imaging and ground penetrating radar (GPR) techniques as well as digital borehole images were used to explore the high anisotropy observed along the Atlantic Coastal Ridge in Chapter 4. Integration of the geophysical methods and borehole images provides further evidence that low resistivity zones are most likely related to solution cavities in the aquifer. The study shows that the presence of interconnected dissolution features can greatly enhance measurable anisotropy and preferential flow direction in carbonate aquifer systems. The approach used in this study shows the applicability of geoelectrical methods to detect geological materials that may contain heterogeneous dissolution features associated with preferential flow.

In chapter 5, a combination of resistivity of multiple resistivity techniques and time-lapse self potential methods were used to investigate the spatial variability of anisotropy with depth and flow characteristics in a municipal well field where there is surface water and groundwater interaction due to continual pumping and a canal network. The direction of anisotropy at this site shows variation with depth. The results from lithological evidence from monitoring wells and unpublished cross-sections from digital borehole images across the site indicate the presence of discrete solution cavities at the site. Self potential data showed that the direction of flow is similar to canal direction and the measured direction of anisotropy from the surface to depth of 4m. The change in

electrical anisotropy direction with depth is probably due to either permeable vuggy and moldic rocks or surface water flow in the canal and groundwater behavior as a result of extraction wells at the site. However more extensive surveying is needed to provide additional details about the site.

6.2 GENERAL CONCLUSIONS

Multi-geophysical techniques and hydrogeological data were used in this research to understand the behavior of anisotropy and other hydrological properties of the Biscayne Aquifer in Miami-Dade County. Focus was placed on the ability of these geoelectrical methods to accurately determine these properties. The key finding of this study is that measurable anisotropy exists in the Biscayne Aquifer, and it is dependent on the location and elevation of the surveyed site. Overall, most of the measured coefficients of electrical anisotropy were generally less than 1.12, except at one site, which had a value as high as 1.36. Higher values generally were located on the Atlantic Coastal Ridge while the lowest values were on the margin to the west of the Everglades (low elevation areas). The study attributes higher values of anisotropy found on the ridge to increased dissolution rates of the oolitic facies of the Miami Formation limestone. This is evident in Chapter 4 where interconnected dissolution features exist. The cause of higher anisotropy is likely due to the presence of a solution cavity oriented in the E-SE direction.

Additionally, 1-D modeling which was used to investigate how resistivity varies regionally with depth, highlights the formation resistivity in the study area. In general, formation resistivity decreased uniformly with depth in the saturated zone, with major

variations attributed to the influence of solution holes or saltwater intrusion along the coast. The predominant trend of minimum resistivity which we interpreted as the maximum hydraulic conductivity is identified to trend generally in the E-W/SE-NW beneath the ridge, and E-W/SE-NW farther west. The study attributes changes in this established pattern to the complex relationship between the permeable vuggy and moldic rocks at the site, and the canal and groundwater extraction wells. Finally, total porosity, secondary porosity, and hydraulic conductivity tensors estimated from geoelectrical properties in this study were consistent with published hydrogeologic studies in the Biscayne Aquifer. The findings from this dissertation demonstrate the capability of geophysical method to characterize the anisotropy of the surficial aquifer and provide valuable information about karsts environments.

6.3 SIGNIFICANCE OF THIS RESEARCH TO THE BISCAYNE AQUIFER

This dissertation provides baseline hydraulic anisotropic information about the Biscayne Aquifer (Chapter 2), a parameter difficult to predict from hydraulic measurement in this aquifer due to its high transmissivity. The results of the study (Chapter 2-4) confirm the pivotal role of the Atlantic Coastal Ridge in groundwater behavior especially in the study area as a result of higher porosity and hydraulic conductivity along the ridge. Hence, regional groundwater models in Miami-Dade County must account for the role of the ridge in water dynamics of the aquifer. Chapter 3 which modeled the sounding data to determine the variation of resistivity with depth confirmed previous sounding studies that showed that saltwater intrusion extent has

remained relatively stable over the past two decades. This is useful information to water organizations in south Florida for effective management of the groundwater resource. Chapter 5 indicates that constant pumping at high rates in the well field coupled with the canal systems have the tendency to alter the preferential flow path (a process that takes thousands of year to develop). This justifies the focus of organizations like USGS, SFWMD and MDWS who are currently researching these issues to understand the effect of the extraction wells on the canal and vice versa. Ultimately, this study could potentially assist these agencies with baseline regional data on parameters like hydraulic anisotropy, porosity and conductivity that can easily be incorporated into their ever-demanding data models.

6.4 CONTRIBUTION TO THE FIELD OF HYDROGEOPHYSICS

The successful integration of geophysical and hydrological measurements for hydrogeological parameter characterization is a key objective in the field of hydrogeophysics. This dissertation makes significant contribution to the field including application of azimuthal resistivity survey to a non fractured-limestone environment for electrical anisotropy measurement. The study modifies the concept of fractured secondary porosity derived for fractured rocks to estimate secondary porosity for linear conduits in carbonate aquifers assuming they are interconnected (Chapter 2). This implies that the concept of fractured secondary porosity can be reasonably applied to non-fractured terrain surficial aquifer system. The dissertation shows that in the Biscayne Aquifer, geophysical properties are related to hydrological parameters and geo-electrical

methods can be used to quantitatively determine anisotropy, porosity and pore orientation (Chapter 2-5). In general, the undertaken research improves our knowledge and understanding of anisotropy in carbonate aquifer through geophysical methods by providing baseline information about this often difficult to determine parameter.

6.5 LIMITATION OF STUDY AND FUTURE RESEARCH

Geophysical methods like those used in this study are influence by artificially buried metallic features in the ground and hence the need for natural areas for the field survey. However in urbanized area like Miami, FL the lack of access to natural areas (no buried utilities) and open grassland spaces restricted the field survey. Hence additional integrated anisotropy studies need to be conducted outside the urban areas of Miami-Dade County (e.g., Everglades National Park) in order to be able to generate and capture the full extent of anisotropy in the Biscayne Aquifer. A key limitation of the study especially in chapter 2 is the absence of detailed geological or hydrological data on anisotropy in the Biscayne aquifer to compare with the measured electrical anisotropy in the study area. Although studies by others have shown the interconnected nature of dissolution features like touching vugs in the Biscayne Aquifer, their impact on anisotropy is still unknown.

Although porosity and hydraulic conductivity values estimated from this study are similar to those measures by others on the aquifer, the impact of scale on these parameters cannot be underestimated. For example in Chapter 3, the krigging interpolation approach used to generate regional map of porosity and resistivity might not

be representative of every point since technically more weights are given to neighboring observations. Hence, almost all hydrogeological studies warn against using regional studies to make site-specific judgment. Future hydrogeological research on carbonate aquifer systems should also include laboratory experiments on rock samples. This would also allow sensitivity analysis to be performed on the impact of scale on geo-electric techniques for measured hydraulic properties. Furthermore, the determined parameters in this study should be incorporated into groundwater modeling software to observe the impact of anisotropy data on regional groundwater models in karst terrains. This might help to understand the level of uncertainty associated with parameters like hydraulic conductivity and porosity used in current groundwater models.

Hydrogeophysical measurement in this study did have limited success in resolving the groundwater and surface water relationship with anisotropy at a hydrological active site (chapter 5). Thus further research which compares this study with core analysis, induction log and isotope analysis at the site should be conducted. This might reveal the impact of canals and extraction wells (if any) on anisotropy at the Snapper Creek Municipal Well Field. Also, because of the inherent complex and non-linear relationships that exist between the hydrogeological and electrical properties, it is likely that single channel instruments for resistivity measurement like those used in this study could miss the complex resistivity components which could easily be related to flow characteristics. Hence, multi-channel instruments might be needed for future research to capture the spectral induced polarization and complex resistivity.

APPENDICES A

Polar plot for azimuthal resistivity survey

BL

CO

DF

A = 4 m
 $\lambda = 1.08 \pm 0.002$
 $\theta = 169 \pm 1^\circ$
 $\rho = 116 \pm 1 \Omega/\text{m}$
 $R^2 = 0.75$

A = 11.3 m
 $\lambda = 1.02 \pm 0.002$
 $\theta = 147 \pm 2^\circ$
 $\rho = 97 \pm 1 \Omega/\text{m}$
 $R^2 = 0.09$

A = 22 m
 $\lambda = 1.02 \pm 0.001$
 $\theta = 137 \pm 3^\circ$
 $\rho = 109 \pm 1 \Omega/\text{m}$
 $R^2 = 0.22$

A = 8 m
 $\lambda = 1.01 \pm 0.001$
 $\theta = 1 \pm 6^\circ$
 $\rho = 93 \pm 1 \Omega/\text{m}$
 $R^2 = 0.06$

A = 16 m
 $\lambda = 1.01 \pm 0.002$
 $\theta = 131 \pm 2^\circ$
 $\rho = 103 \pm 1 \Omega/\text{m}$
 $R^2 = 0.15$

A = 32 m
 $\lambda = 1.02 \pm 0.001$
 $\theta = 118 \pm 2^\circ$
 $\rho = 111 \pm 1 \Omega/\text{m}$
 $R^2 = 0.37$

EC

A = 5 m
 $\lambda = 1.02 \pm 0.001$
 $\theta = 42 \pm 1^\circ$
 $\rho = 156 \pm 1 \Omega/\text{m}$
 $R^2 = 0.61$

A = 11 m
 $\lambda = 1.03 \pm 0.001$
 $\theta = 53 \pm 1^\circ$
 $\rho = 142 \pm 1 \Omega/\text{m}$
 $R^2 = 0.64$

A = 22 m
 $\lambda = 1.04 \pm 0.001$
 $\theta = 63 \pm 1^\circ$
 $\rho = 138 \pm 1 \Omega/\text{m}$
 $R^2 = 0.76$

A = 8 m
 $\lambda = 1.02 \pm 0.001$
 $\theta = 54 \pm 1^\circ$
 $\rho = 156 \pm 1 \Omega/\text{m}$
 $R^2 = 0.59$

A = 16 m
 $\lambda = 1.03 \pm 0.001$
 $\theta = 50 \pm 1^\circ$
 $\rho = 134 \pm 1 \Omega/\text{m}$
 $R^2 = 0.88$

A = 32 m
 $\lambda = 1.03 \pm 0.001$
 $\theta = 30 \pm 1^\circ$
 $\rho = 135 \pm 1 \Omega/\text{m}$
 $R^2 = 0.82$

MC

ML

NT

PP

PL

SC

SL

WP

WL

APPENDIX B.

Matlab code for non-linear least square

```

% Loading files
% Input from Table
load birdlake_all.txt; % load the data
t =birdlake_all(:,1); %t=theta, the independent variable is in column 1
rho0 =birdlake_all(:,2); %resistivity the dependent variable is in column 2 @4.0m

disp('4m spacing')
rhoobs =birdlake_all(:,2);
% Extract intial values from the data set.
[rhoobsmin, irhoobsmin] = min(rhoobs); %
theta1 = t(irhoobsmin);
if theta1>pi
    theta=theta1-pi;
else
    theta = theta1;
end

theta;
rhom=mean(rhoobs); % mean resistivity
n=sqrt((max(rhoobs)/min(rhoobs))); % Anisotropy estimate
P0=[rhom n theta];

% Inverse Modeling
for i=1:100
    Kfact = 1/(2 - sqrt(2));
    %calculate forward problem at current values of p
    P=[rhom;n;theta];
    % Intial calculation
    A = (2./ ( sqrt( 1 + (n*n - 1) .* cos(theta-t) .* cos(theta-t) ) ));
    B = (1 ./ ( sqrt( 2 + (n*n - 1) .* (1 + sin(2*(theta-t))) ) ));
    C = (1 ./ ( sqrt( 2 + (n*n - 1) .* (1 - sin(2*(theta-t))) ) ));

    rhoacalc= Kfact*rhom*(A-B-C);

    %evaluate partial derivatives at current values of p

    %drhoa/drhom
    drhoadrhom = Kfact.*(A-B-C);

    %drhoa/dn
    A1 = (((cos(theta-t) .* cos(theta-t)).*(A).^3));
    B1 = ((1 + sin(2*(theta-t)).*(B).^3));
    C1 = ((1 - sin(2*(theta-t)).*(C).^3));

    drhoadn = (rhom*Kfact*n.*(-A1+B1+C1));

    %drhoa/dtheta
    A2=(((cos(theta-t).*sin(theta-t).*(n*n - 1)).*(A).^3));
    B2= (((cos(theta-t) .* cos(theta-t)).*(n*n - 1).*B.^3));
    C2= (((cos(theta-t) .* cos(theta-t)).*(1-n*n ).*C.^3));

    drhoadtheta=(rhom*Kfact.*(A2-B2-C2));

    %Fill Z the matrice
    Z=[drhoadrhom drhoadn drhoadtheta];

    %caluculate data pertubation
    dr=rhoobs- rhoacalc;

    %invert for parameter pertubations
    dp=pinv(Z)*dr;

    %adjust parameters
    P=P+dp;
    rhom=(P(1,:));

```

```

n = (P(2,:));
theta = (P(3,:));

%display current values and test for convergence
%disp(sprintf('%10.3f',P));

% Convergence criteria
rp=dp./P;
conv = sqrt(rp'*rp);
if conv < 0.005
    break
end
iter=i ;
if ( iter == 100 );
    disp(' !! Maximum Number of Iterations Reached Without Convergence !!')
    stop = 1;
end;
end
% --- End of Main Loop
% --- convergence achieved, find covariance and confidence intervals

disp(' Number of iteration to converge')
disp(iter+1)% Number of iteration to converge

% Goodness of fit and parameter uncertainties
% standard error of parameters
Npnt = length(rhoobs);
N=Npnt-3; % degree of freedom n-m
Ra=(rhoobs- rhoacalc);
vp= (1/N) * ((Ra'*Ra));%% Variance

Cv=inv(Z'*Z).*(1./vp) ; % parameter correlation matrix

sigma_p =sqrt(diag(Cv)); % standard error of parameters

%RMS Root mean Square of the model
rms= sqrt((sum(rhoobs(:) - rhoacalc(:)).^2)/(length(rhoobs)));
rms_percent=rms*(100/P(1));

%converting angle to degrees
Ang_P=P(3)*180/pi;
Ang_sigma_p =(sigma_p(3))*180/pi ;

% standard error of the fit
sigma_R=sqrt(diag(Z*Cv*Z')); %
sigma_rms= sqrt(sigma_R); % not factoring in the rms
RMS_fit= sqrt(mean(sigma_rms.^2)); %

%Estimating chi_square
Nfit = length(P);
weight_sq = (Npnt-Nfit+1)/(Ra'*Ra) * ones(Npnt,1);
Chi_sq = Ra' * ( Ra.* weight_sq ); % rule of thumb chi_sq ~ deg of freedom

% errors associated with parameters
error1 = rms;
error2=rms_percent*P(2)/100;
error3=(rms_percent*P(3)/100)*180/pi;

% R_sq = R-squared coefficient of multiple determination
R_sq = corrcoef([rhoobs rhoacalc]);
R= R_sq(1,2).^2;

% display results
disp(' rhomean +- error anisp +- error theta +- error RMS percent RMS_fit R_sq ')
disp(' -----')

```

```

disp([ P(1)      error1          P(2)   error2      Ang_P  error3   rms_percent   RMS_fit   R ])

%plot and check
figure(1)
subplot( 2,1,1)
y=birdlake_all(:,1);
t =0:0.1:6.3;%
rhom=((P(1,:)));
n=((P(2,:)));
theta =((P(3,:)));

A = (2./ ( sqrt( 1 + (n*n - 1) .* cos(theta-t) .* cos(theta-t) ) ));
B = (1 ./ ( sqrt( 2 + (n*n - 1) .* (1 + sin(2*(theta-t))) ) ));
C = (1 ./ ( sqrt( 2 + (n*n - 1) .* (1 - sin(2*(theta-t))) ) ));

rhoacalc= Kfact*rhom*(A-B-C);

maxrho=max(rhoobs);
h=polam(t,rhoacalc,maxrho,'r');
view(90,-90) ;
set(h,'LineWidth',3)
hold on
h=polam(y,rhoobs,maxrho,'+');
set(h,'LineWidth',6)

```


VITA

ALBERT YEBOAH-FORSON

EDUCATION

- 2013 Doctoral Candidate, Geosciences
Florida International University
Miami, Florida
- Research: Geophysical characterization of hydrogeological parameters in surficial aquifer systems.*
- 2007 M.S Applied Environmental Geoscience
Eberhard Karls University of Tübingen
Germany
- Research: Contaminant fate and transport effects on soil and groundwater in aqueous environments.*
- 2003 B.S. Physics (Honors)
Kwame Nkrumah Univ. of Science and Technology
Ghana

PUBLICATIONS PRESENTATIONS AND ABSTRACTS

Peer Review

- Yeboah-Forson, A., and Whitman, D. (2013). Electrical Resistivity Characterization of Anisotropy in the Biscayne Aquifer,. Ground water (accepted)
- Yeboah-Forson, A., Comas. X. and Whitman, D. (2013). Determination of anisotropic karst features in the Biscayne Aquifer using electrical resistivity imaging and Ground Penetration Radar Geophysical Methods (In review for submission)
- Yeboah-Forson, A., and Whitman, D. (2013). Hydrogeophysical Flow Analysis of Snapper Creek Well Field (Manuscript in preparation)
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