Electrical Resistivity Characterization of Anisotropy in the Biscayne Aquifer

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Abstract

Electrical anisotropy occurs when electric current flow varies with azimuth. In porous media, this may correspond to anisotropy in the hydraulic conductivity resulting from sedimentary fabric, fractures, or dissolution. In this 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. The observed electrical anisotropy was used to estimate hydraulic anisotropy (ratio of maximum to minimum hydraulic conductivity) which ranged from 1.18 to 2.83. The largest values generally were 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 higher values of anisotropy found on the ridge may be due to increased dissolution rates of the oolitic facies of the Miami formation limestone compared with the bryozoan facies to the west. The predominate trend of minimum resistivity and maximum hydraulic conductivity was E-W/SE-NW beneath the ridge and E-W/SW-NE farther west. The anisotropy directions are similar to the predevelopment groundwater flow direction as indicated in published studies. This suggests that the observed anisotropy is related to the paleo-groundwater flow in the Biscayne Aquifer.

Introduction

The Biscayne Aquifer of SE Florida is a carbonate aquifer which exhibits karst features, and secondary porosity is due to the presence of touching vugs, conduits, and solution holes (Cunningham et al. 2006). Dissolution can alter the direction and velocity of groundwater 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. 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 determine 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 ft²/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 to 57,000 ft²/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 (DC) 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 this can occur in the presence of steeply dipping beds or fractures (Taylor and Fleming 1988; Lane et al. 1995; Boadu et al. 2005; Ramanujam et al. 2006). Therefore, electrical anisotropy measurements can provide a useful analog for estimating the maximum 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 the principal components of the hydraulic conductivity tensor. These estimates provide an alternative assessment tool for characterizing the directional properties of the aquifer at the regional scale.

**Geology and Hydrogeology of the Biscayne Aquifer**

The Biscayne Aquifer is a Pleistocene unconfined carbonate aquifer located in SE 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 1). A detailed discussion of the geology and the hydrogeology of the Biscayne Aquifer can be found in studies such as 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. 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 Costal 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) due to 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 not be representative of the true 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). The horizontal and vertical permeability both increase with increases in porosity. Hydraulic conductivities in excess of 3.5 cm/s (10,000 ft/d) 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 that formation (Fish and Stewart 1991).
Electrical Anisotropy

Electrical anisotropy occurs in the subsurface as a result of current flowing differently in one direction relative to another. This 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 \( \theta \) (Figure 2). The resistivity across the bedding plane (transverse resistivity, \( \rho_t \)) is normally greater than resistivity parallel to bedding (longitudinal resistivity, \( \rho_l \)). The anisotropic media is characterized by the coefficient of anisotropy, \( \lambda \), and the mean resistivity, \( \rho_m \).

\[
\lambda = \sqrt{\frac{\rho_t}{\rho_l}} \\
\rho_m = \sqrt{\rho_t \rho_l}.
\]

The coefficient of anisotropy, \( \lambda \), for a homogeneous anisotropic geological unit is always greater than 1 as 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 3). The square array has several advantages over the linear array, including faster setup 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, \( \theta_0 \), the apparent resistivity, \( \rho_a \), varies in an elliptical-like pattern given by Habberjam (1972),

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

where \( \theta \) is the minimum (longitudinal) resistivity direction.

Numerous studies have used measurements of electrical anisotropy to characterize steeply dipping fractures and beds in the subsurface (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 preferred direction of dissolution cavities resulting in anisotropy in the hydraulic conductivity.

Data and Method of Analysis

Data for this study were collected over a 3-year period from December 2008 to January 2012 at 13 sites in eastern Miami Dade County (Figure 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. Because of space limitations, the sites were restricted to public parks and other areas where open space are 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; Austin, Texas) Super-Sting R1/IP 28-electrode resistivity imaging system. This 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 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 \( \sqrt{2} \). 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. This procedure produced a set of anisotropic soundings which provided a measure of 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). While the exact relationship between the measurements and depth requires numerical modeling, this 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, \( \rho_m \), coefficient of anisotropy, \( \lambda \), and direction of minimum resistivity, \( \theta \), in Equation 3 using an iterative nonlinear 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 nonlinear inverse models, this 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 \( \rho_m, \lambda, \) and \( \theta \) 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.

### Results

In total, 100 sets of azimuthal measurements were conducted 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.

Examples of typical survey results are shown in Figure 5. Figure 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 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 3 deviates from a pure ellipse (Figure 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 two distinct directions of fracture orientation and in the absence of a defined fracture the true minimum resistivity direction obtained from Equation 3 is
Figure 5. Examples of polar plots of apparent resistivity plotted against azimuth (°). The thick solid line is the best fitting apparent resistivity ellipse obtained from Equation 3. Locations of sites are shown in Figure 1.

Figure 6. Distribution of the coefficient of anisotropy (λ) measurements for all the sites and square sizes in the study.

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 1.12 with a mean of 1.06 and a mode of 1.03 (Figure 6). At one site, West Perrine Park (WP, Figure 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 to 23 m (effective depth, Ze, of 5 to 11 m).

The direction of minimum resistivity, θ, shows that anisotropy on the regional scale, exhibits a range of different orientations (Figure 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 8). In general, the coefficient of anisotropy was marginally higher for sites on the Atlantic Coastal Ridge compared with those close to the Everglades. For example, at a square size of 11.3 m (Ze, 5.6 m), the magnitude of anisotropy for sites on the ridge ranges from 1.07 to 1.08 while those behind the ridge are 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 11.3 m (Ze, 5.6 m) some sites trend NE while others trend SE or SSE.

Three of the sites (BL, PL, and SC; Figure 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 to 8 m (Ze, 2 to 4 m), the minimum resistivity trends in the SE direction whereas for square sizes greater than 11 m (Ze ≥ 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 required detailed modeling but this analysis is beyond the scope of this paper.

Hydraulic anisotropy

Electric current flow and groundwater flow have obvious analogies through the mathematical similarities of Ohms law to Darcy law (Ahmed et al. 1988). Hence, the electrical anisotropy results may be interpreted as anisotropy in the 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. Empirical relationships between electrical properties and hydraulic conductivity may be derived 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 Soupios et al. (2007) to accurately estimate aquifer hydraulic conductivity and porosity from surface geoelectrical measurements.

Koukadaki 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,

$$K = a_2 \times \frac{\gamma}{\mu} \left( \frac{\rho_0}{\rho} \right)^{(b_2/m)} \tag{4}$$

where $m$, $\alpha$, $a_2$, and $b_2$ are empirically derived constants which depend on rock type, $\rho$ and $\rho_0$ are the formation and the fluid resistivities, and $\gamma$ and $\mu$ are the specific weight and viscosity of the fluid, respectively. The two principal components of the hydraulic conductivity tensor, $K_{hi}$ and $K_{hl}$ are then calculated by substituting the transverse and longitudinal resistivities, $\rho_t$ and $\rho_l$ into (Equation 4). These values are then divided to obtain the hydraulic anisotropy $\Psi_h$,

$$\Psi_h = \frac{K_{hl}}{K_{hi}} = \left( \frac{\lambda^2}{\lambda^m} \right)^{(b_2/m)} \tag{5}$$

where $\lambda$ is coefficient of electrical anisotropy defined in Equation 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),

$$k = a_2 \phi^{b_2}. \tag{6}$$

In this 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). This analysis derived estimates for $a_2$ and $b_2$ of $4.22 \times 10^{-06}$ and 5.69 respectively. A cementation factor, $m$, of 1.7 and a tortuosity factor, $\alpha$, of 1 were assumed (Kwader 1985). These parameters and the average electrical anisotropy were used to calculate $K_{hi}$, $K_{hl}$, and $\Psi_h$ at each site.

The calculated values for $K_{hi}$, $K_{hl}$, and $\Psi_h$ for each site are shown in Table 1 and Figure 9. A mean
Table 1
Summary Hydraulic Parameters at Each Site Calculated from Electrical Resistivity

<table>
<thead>
<tr>
<th>Site</th>
<th>Coefficient of Anisotropy (λ)</th>
<th>Hydraulic Conductivity $K_{ht}$ (min) m/s</th>
<th>Hydraulic Conductivity $K_{hl}$ (max) m/s</th>
<th>Hydraulic Anisotropy $\psi_h$</th>
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</thead>
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<tr>
<td>BL</td>
<td>1.06</td>
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<td>1.51</td>
</tr>
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<td>CO</td>
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<td>0.057</td>
<td>0.097</td>
<td>1.71</td>
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<tr>
<td>DF</td>
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<td>0.079</td>
<td>0.094</td>
<td>1.19</td>
</tr>
<tr>
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<td>1.03</td>
<td>0.029</td>
<td>0.035</td>
<td>1.19</td>
</tr>
<tr>
<td>MC</td>
<td>1.09</td>
<td>0.028</td>
<td>0.039</td>
<td>1.78</td>
</tr>
<tr>
<td>ML</td>
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<td>0.069</td>
<td>0.092</td>
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</tr>
<tr>
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</tr>
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<td>0.229</td>
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</tr>
<tr>
<td>PL</td>
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<td>0.032</td>
<td>0.044</td>
<td>1.39</td>
</tr>
<tr>
<td>SC</td>
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<td>0.057</td>
<td>0.087</td>
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<tr>
<td>WL</td>
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<td>0.133</td>
<td>0.167</td>
<td>1.26</td>
</tr>
</tbody>
</table>

Notes: $K_{ht}$ and $K_{hl}$ are the transverse and longitudinal horizontal hydraulic conductivity respectively. Geographic locations of sites are shown in Figure 1.

Discussion

The coefficient of electrical anisotropy obtained from the study generally ranged from 1.01 to 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 ranged 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 terrains. At one site (WP, Figures 1 and 4) the anisotropy was as high as 1.36. This may be due to large solution cavities and is the subject of further investigation.

The electrical measurements in this study were used to estimate the hydraulic anisotropy in the aquifer. Hydraulic anisotropy is highest on the Atlantic Coastal Ridge and generally decreases to the west (Figure 9). This suggests that the geology and topography are the important factors in the development of the anisotropy. 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 1), is located near a 3 m high bluff which contains cross beds dipping 20° to 30°. The strike of these beds, 70° ± 2°, is similar to the direction of minimum resistivity, 60° ± 5°, 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 paleogroundwater 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 the high both E/SE toward Biscayne Bay and W/SW toward the Everglades (Parker et al. 1955). When drainage canals were constructed across the ridge, the hydrostatic head decreased allowing regional groundwater flow toward the E/SE. Fenemna et al. (1994) modeled the groundwater flow under predevelopment conditions. Beneath the Atlantic Coastal Ridge, the flow was highest and was oriented toward 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 toward the W and SW and was smaller in

Figure 9. Map showing the directional mean (Davis, 1986) of minimum resistivity and the mean horizontal anisotropy averaged from all the array sizes at each site. The size of the arrow is proportional to the magnitude of hydraulic anisotropy.
magnitude. This is consistent with most of the observed anisotropy to the west of the ridge (Figure 9). This suggests a causal mechanism linking the direction of the minimum resistivity and maximum hydraulic conductivity to dissolution caused by the paleo-groundwater flow.

Summary

Over 100 azimuthal square array resistivity measurements were collected in eastern Miami-Dade County, Florida, 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 magnitude and direction of the principal components of the hydraulic conductivity tensor.

The coefficient of electrical anisotropy ranged from 1.01 to 1.36 with a modal value of 1.03. The estimated 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 toward the SW-NE and E-W. In both cases, the magnitude and direction of the anisotropy are 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.

This study shows that azimuthal resistivity surveys can provide reasonable estimates of hydraulic anisotropy 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 for quantitative measurement in regional hydrological studies.

Acknowledgments

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Supporting Information

Additional Supporting Information may be found in the online version of this article:

Figure S1. Maps of other array sizes.

References


