

“EXPLORING THE NATURE OF FREE CONVECTION IN A SABKHA WITH
ELECTRICAL IMAGING AND HYDROLOGICAL MODELING”

By

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ABSTRACT

“EXPLORING THE NATURE OF FREE CONVECTION IN A SABKHA WITH ELECTRICAL IMAGING AND HYDROLOGICAL MODELING”

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Free convection is an important process that has applications in many disciplines of earth science and is a factor in a variety of environmental issues. Although the study of free convection has been historically limited to theory, laboratory experiments, numerical modeling, and secondary field evidence, recent studies have used electrical resistivity geophysical methods to capture direct evidence of this phenomenon in a natural field setting. However, there is still very little known about the transient characteristics of free convection in a field setting.

This thesis uses geophysics and hydrological modeling to investigate the occurrence, initiating mechanisms, and transient nature of free convection in a coastal sabkha in the United Arab Emirates. The electrical resistivity data shows that free convection in the sabkha is an episodic process that is likely caused by the dissolution of the halite crust by precipitation and its subsequent infiltration to the aquifer. Hydrological modeling of the system supports precipitation as the initiating mechanism and allows for the characterization of the growth and decay of the convective fingers.

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LIST OF SYMBOLS

Symbol	Parameter	Unit
q	Water flux	m/d
K	Hydraulic conductivity	m/d
dh	Change in head	m
dl	Change in length	m
g	Acceleration of gravity	m/s ²
k	permeability	m ²
β	Linear expansion coefficient	-
ΔC	Change in concentration	kg/m ³
H	Aquifer thickness	m
θ	Porosity	-
ν ₀	Kinematic viscosity	m ² /s
μ ₀	Dynamic viscosity	kg/m•s
ρ	Density	kg/m ³
ρ ₀	Initial density	kg/m ³
ρ _s	Infiltrating density	kg/m ³
ρ _u	Upwelling density	kg/m ³
Δρ	Change in density	kg/m ³
D _L	Coefficient of molecular diffusion	m ² /s
D _D	Dispersion coefficient	m ² /s
u	Darcy velocity	m/s
p	Fluid pressure	Pa
D	Vector in direction in which gravity acts	m
Q _m	Mass source term	kg/(m ³ •s)
θ _s	Fluid volume fraction	-
c	Concentration	kg/m ³
c ₀	Initial Concentration	kg/m ³
c _s	Infiltrating Concentration	kg/m ³
c _u	Upwelling Concentration	kg/m ³
τ	Tortuosity	-
L	Aquifer thickness	m
q _u	Upwelling flux rate	mm/yr
U _c	Convective velocity	m/d
α	Dispersivity	m

INTRODUCTION

Free Convection

Free convection is the mixing of fluid driven by buoyancy forces. It occurs when temperature and salinity related changes create a density inversion that causes the system to become critically unstable. A density inversion is a light fluid overlain by a dense fluid. When the contrast between the fluids is great enough, the less dense fluids rise and higher density fluids sink at rates faster than they diffuse. Free convection appears in nature in the form of convection cells or as descending fingers of dense fluid.

Free convection is important because it causes mixing of solutes at faster rates than advection and diffusion, over larger spatial scales, and involving larger volumes of materials. It applications in many disciplines including: atmospheric sciences, limnology, oceanography, geophysics, hydrogeology, and astrophysics (Diersch and Kolditz, 2002). Movement by free convection plays an important role in a variety of environmental areas of concern such as seawater-groundwater interactions, contaminant migration, flow through formations at radioactive waste disposal sites, and the salinization of groundwater in arid and semi-arid regions (Simmons, 2005; Bauer et al., 2006).

Most free convection has been studied through theory (Rayleigh, 1916; Horton and Rogers, 1945; Lapwood, 1948), laboratory experiments (Schneider, 1963; Elder 1967), and numerical modeling (Wooding, 1957; Elder 1967). Identification of free convection in field settings has been limited to secondary evidence such as: unusual tritium distributions (Wood et al., 2002), measured density inversions in groundwater, reversals in hydraulic gradients, high fluid fluxes, theoretical considerations (high Rayleigh number), modeling of a system showing

convection, and salt deficits in saline lakes (Simmons, 2005). This secondary field evidence, while supporting that free convection was possible or that mixing was taking place at rates faster than diffusion and dispersion, did not confirm that free convection was actually occurring. Direct evidence in a field setting has been identified as one of the missing links in the study of free convection (Simmons et al., 2005). In addition, the transient characteristics of free convection in field settings are poorly known. In recent years, electrical geophysical methods have been used to capture direct field evidence of this phenomenon (Bauer et al., 2006; Van Dam et al., 2009; Stevens et al., 2009).

Previous Studies

Introduction. There have been three major studies using geophysics to image free convection in the subsurface. In each study, free convection was driven by density differences caused by variations in solute concentrations and the fingering was imaged using electrical resistivity. Electrical resistivity measurements were employed based on the relationship between salinity variations, fluid electrical conductivity, and bulk electrical resistivity.

Okavango Delta. Free convection has been hypothesized as a possible cause of a salt deficit in the surface water of the Okavango Delta of Botswana (Gieske, 1996). A majority of the water in the system is lost to evapotranspiration from islands between the delta channels. This evapotranspiration leads to the accumulation of solutes near the surface of the freshwater aquifer creating a density inversion (Bauer et al., 2006).

Bauer et al. (2006) used a combination of surface and borehole electrical resistivity measurements to map the distribution of salinity under two islands in the delta. Under one island, they were able to capture evidence of a single convection finger. Through hydrodynamic

modeling they determined that the instabilities required for free convection would take 100 to 150 years of evapoconcentration and that the saline finger they imaged was likely over 200 years old (Zimmerman et al., 2006).

This was the first field evidence capturing an apparent saline finger using electrical resistivity. Hydrodynamic modeling was consistent with their geophysical findings. However, with the development of fingering in the Okavango Delta taking 100's of years to occur, testing and calibrating models that simulate the transient nature of free convection is nearly impossible.

Padre Island National Seashore. The tidal flats on Padre Island, Texas, have a low hydraulic gradient, undergo evapoconcentration, and are occasionally inundated by saline water from Laguna Madre making them a likely location for free convection to occur (Stevens et al., 2009). Stevens et al. (2009) used a variety of secondary evidence such as hydraulic gradients, sampling data showing occasional density inversions, and theoretical calculations to show the potential for free convection at their field site.

A 3D electrical resistivity survey was used to capture evidence of saline fingering beneath Padre Island. Stevens et al. (2009) employed a surface electrode array with a 14 X 6 grid of electrodes with 2 meter spacing.

The secondary evidence presented by Stevens et al. (2009) all suggests that free convection is indeed taking place. However, the plumes of high salinity and density inversions that they identified in their inverted 3D resistivity results all occur along the edges and corners of their data plots, where because of the way ERI is measured, these features must have been extrapolated to areas where there are little or no data points.

The United Arab Emirates. Wood et al. (2002) suggested that density driven convective flow was taking place in a sabkha in the United Arab Emirates (UAE) based on measured density inversions and a distribution of elements consistent with vertical convective mixing. The sabkha makes an ideal natural laboratory for the study of free convection in a field setting because the aquifer sands are very homogeneous, meaning any variation measured in resistivity will be due to variations in solute concentration.

Van Dam et al. (2009) used electrical resistivity imaging (ERI) at the sabkha to document the existence of fingering. They used two perpendicular 2D surface arrays with 84 electrodes each with 1 meter spacing and one array with 84 electrodes with a 0.5 m electrode spacing.

Van Dam et al. (2009) were able to, in contrast to the other two studies, image a pattern of fingering over a large spatial scale similar to patterns shown in laboratory and modeling studies of free convection. This study, however, is limited in the fact that it is just a single snapshot in time and tells us very little of the characteristics of the processes involved in free convection.

Objective and Approach

The objective of this thesis is to investigate the occurrence and transient nature of free convection in the UAE coastal sabkha. It will be testing the hypothesis suggested in Van Dam et al. (2009) that free convection in the sabkha is episodic by nature and caused by precipitation-induced dissolution of the halite crust and subsequent infiltration to the aquifer.

The first part of the thesis is an introduction to the study area and is accompanied by a description of the setting, climate, and a detailed characterization of the physical, hydraulic, and electrical properties of the capillary zone. It will then compare the data from Van Dam et al.

(2009), which captured evidence of free convection in 2008 to data collected during the following two years at the same site to test if free convection in the sabkha is episodic in nature rather than continuous based on proposed models for creating the density inversion. The next section will employ a hydrologic model based on measured field values to characterize the transient nature of free convection. The hydrologic model will be used for testing the hypothesis of the mechanism for the onset of convective fingering, the speeds at which the fingers grow, and the rates at which they decay.

STUDY AREA

Introduction

A sabkha, which means “salt flat” in Arabic, is a supratidal environment where evaporite minerals precipitate displacively within the sediments in the capillary zone (Boggs, 2001). The study area is a sabkha located about 40 km southwest of the city of Abu Dhabi along the coast of the Arabian Gulf in the United Arab Emirates (Figure 1). The field site consists of ~11.5 m of sand from sand dunes reworked by sea level rise underlain by Miocene carbonates. The sands are essentially homogeneous (Wood et al. 2002), except for in the top meter of the sabkha, which is characterized by authigenic evaporites. The topography of the sabkha is relatively flat with a gradient of 1 m per 10 km toward the coast (Sanford and Wood, 2001).

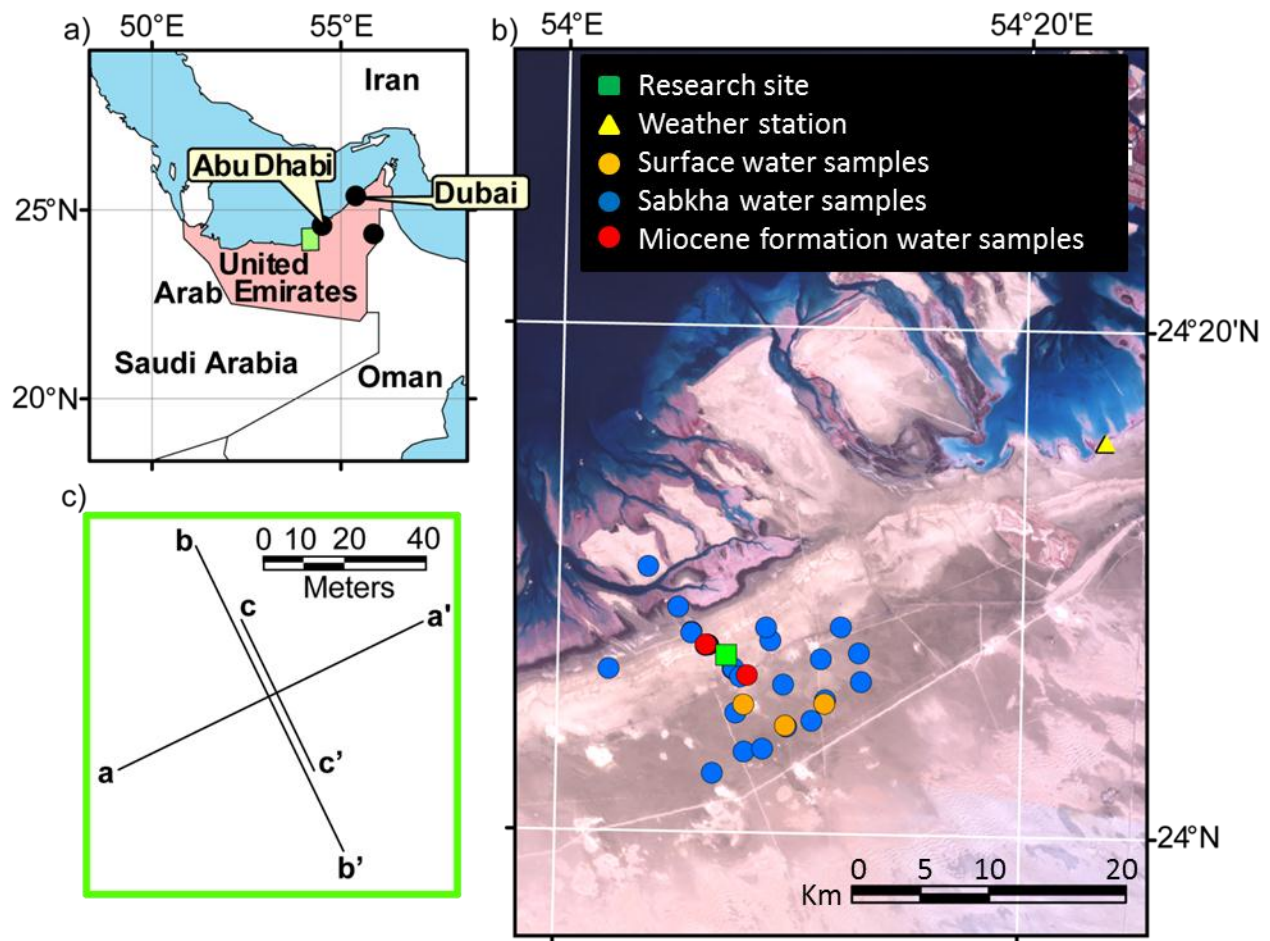


Figure 1. Location of the study area. Regional map (a), geophysical research site, water well and weather station locations (b), and the layout of geophysical survey lines (c) (adapted from Van Dam et al., 2009). For interpretation of the references to color in this and all other figures, the reader is referred to the electronic version of this thesis.

Groundwater Properties

The region around the study area contains a series of groundwater wells that extend to various depths throughout the sabkha aquifer and into the formation below (Figure 1). Additionally, surface water samples were taken where ponding occurred after a rainfall event (Wood et al., 2002). The water in the sabkha was found to have an average solute concentration of 286,415 mg/L and a density of 1.1858 g/cm^3 (Wood et al., 2002). The Tertiary formation below has a TDS (total dissolved solids) of 101,453 mg/L and a density of 1.0753 g/cm^3 .

(densities are for 20°C) (Van Dam et al., 2009). Water and solute budgets for the sabkha have been calculated by Wood et al. (2002). They determined that the majority of the solutes in the sabkha aquifer have come from the slow upwelling (around 4 mm/yr) of saline water from the Tertiary formation below. Due to the low hydraulic gradient the solutes have a residence time in the system of ~26,000 years. Water in the system, however, has a residence time of only ~50 years. The bulk of the water flux in and out is from evaporation and precipitation. Surface water samples had an average density of 1.2675 g/cm³ and ranged in solute concentration from 292,487-536,969 mg/L. The high concentration and density of the surface water is due to dissolution of the dry halite crust (Wood et al., 2002).

The lateral water flux through the sabkha is calculated using Darcy's law:

$$q = -K(dh/dl) \quad (1)$$

where q is the flux of water (m/d), the negative sign signifies flow from higher to lower potential, K is the hydraulic conductivity (m/d), and dh/dl is the hydraulic gradient (m/m). Using a falling-head permeameter, the hydraulic conductivity was measured in the sabkha sands to be 6.57 m/d (compared to 1 m/d measured by Sanford and Wood, 2001) and the hydraulic gradient is 0.0002 (Sanford and Wood, 2001). Therefore, the lateral flux ranges from 0.073 to 0.48 m/yr. The seepage velocity is found by dividing the lateral flux by the porosity (0.38) of the aquifer, which gives a velocity of 1.26 m/yr (~3.5 mm/day).

Climate

The UAE is a hot arid sub-tropical desert. The mean minimum and maximum annual temperatures are 21.7 and 33°C respectively. Long term precipitation values (1983-2010) from a WMO weather station near the study area indicate that the average annual precipitation is around 32 mm/yr, mostly from November to April, with a large standard deviation since it is not uncommon to see multiple years with no rain (Figure 2). The pan evaporation rate of fresh water in the area is 3500mm/yr (Bottemley, 1996); however, evaporation rates from the surface of the sabkha were measured by Sanford and Wood (2001) and found to be less than 3% of pan evaporation (Table 1). Transpiration is not occurring because there is no vegetation on the sabkha.

Table 1. Water budget, adapted from Sanford and Wood (2001).

Budget component	Flux rate (mm/yr)	Ranges	Sources
Evaporation	69	50-88	Sanford and Wood (2001)
Precipitation (recharge)	32	0-132	WMO Weather station 41217
Lateral flux in	80	73-480	Sanford and Wood (2001)
Lateral flux out	80	73-480	Sanford and Wood (2001)
Upward leakage	4	4-5	Sanford and Wood (2001)
Water table depth	1 m	0.5-1.3	Wood et al. (2002); field measurement

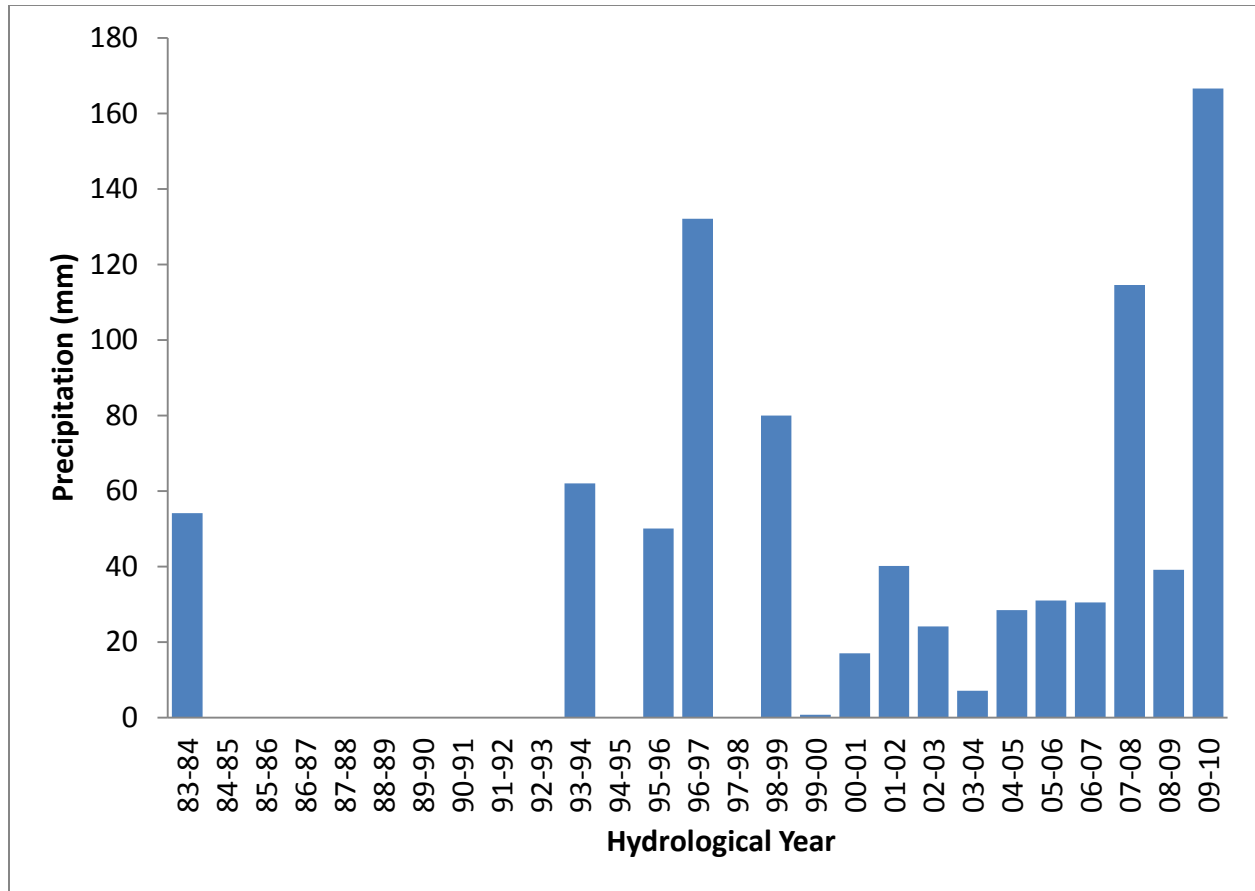


Figure 2. Annual precipitation (mm) based on hydrological year (Oct. 1- Sept. 30) from 1983-2010 from WMO Weather Station 41217 near Abu Dhabi.

Characterization of the Capillary Zone

Methods. In 2010, a detailed characterization of the capillary zone was performed. A trench was dug from the sabkha surface down to the water table (1.3 m) and a detailed description of the sediment layers was made. The electrical properties of the entire vertical column were measured in 5 cm increments using a horizontal Wenner array with an a-spacing of 5 cm (Figure 3), and the saturated hydraulic conductivities in the trench were measured for several depths (Figure 4).

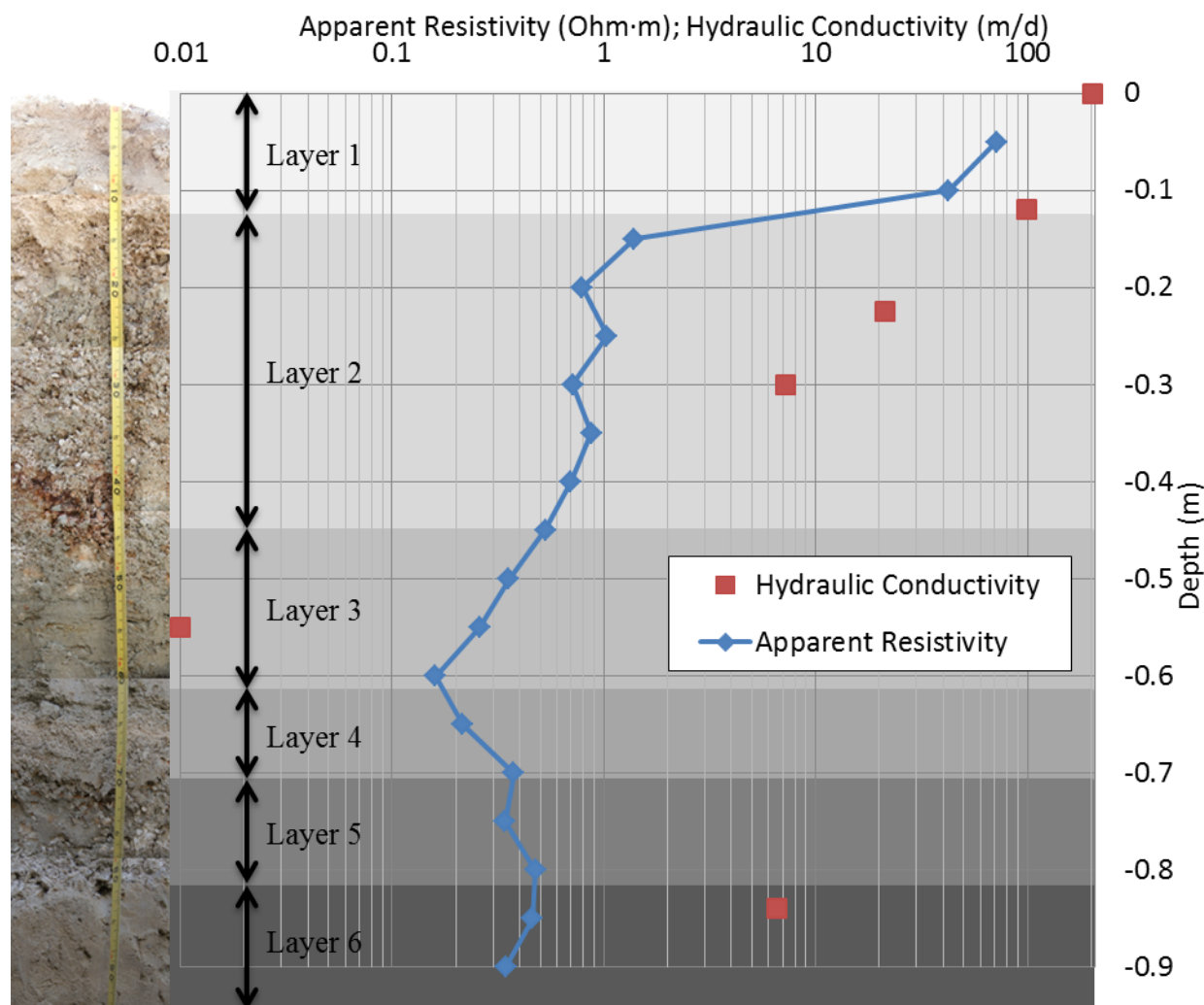


Figure 3. Cross section of trench wall (tape measure scale in cm) combined with a plot of apparent resistivity and hydraulic conductivity vs. depth with the 6 layers identified.

Physical and Electrical Properties. The sediments are classified into 6 distinct layers (Figure 3). The top 12 cm, layer 1, are characterized by a dry, friable, sand-size crust containing crystals of gypsum 5-10 mm in diameter. This layer has the highest measured electrical resistivity in the system ranging from 40-70 Ohm·m. Layer 2 extends from 12 to 45 cm and is made up of a mix of sand- and clay-size grains and contains gypsum crystals 15-20 mm in diameter. A red iron oxide band is found at 42-43 cm depth. The electrical resistivity of this layer ranges from 0.5-1.3 Ohm·m. Layer 3, between 45 and 62 cm depth is comprised of 1-5

mm thick horizontal layers of blue clay. Gypsum crystals are found throughout this layer, but are decreasing in abundance with depth. This clay layer has a very low electrical resistivity of 0.18-0.36 Ohm•m. The 4th layer is entirely made up of coarse-grained sand-size gypsum crystals, ranges from 62 to 71 cm depth, and has an electrical resistivity ranging from 0.21-0.38 Ohm•m. Below that is layer 5, 71-82 cm, which is made up of similar sized gypsum crystals but is set in a clay matrix containing large, 15-50 mm diameter, soft anhydrite nodules. This layer has an electrical resistivity between 0.32-0.48 Ohm•m. The material from 82 cm down is the homogeneous fine grain sand that makes up layer six and the rest of the aquifer, the capillary portion of which has an electrical resistivity ranging from 0.32-0.48 Ohm•m.

The top of the capillary zone in this system has a diurnal fluctuation, but remains at or near the surface (Wood, 2002). High resistivities were measured in the top 15 cm where the sediment is mostly dry during the day. Below that, resistivity gradually drops until around 60 cm as the sediments become more and more saturated. From 70 cm and down the electrical resistivity remains fairly stable at values similar to the aquifer, suggesting that this section of the capillary zone is saturated.

Hydraulic Properties. Horizontal tiers were dug in the trench at a series of depths from the surface down to 84cm. The vertical saturated hydraulic conductivity was measured at each tier using a single-ring falling-head infiltrometer (Figure 4). The infiltrometer consisted of a section of pipe 15 cm in diameter with a graduated scale on the side in centimeters. The infiltrometer was carefully inserted 5 cm into the sediment, to minimize disturbance, and then filled with 1.5 L of sabkha water. The time was recorded at every 0.5 cm of head loss (Table 2). The data was then plotted as cumulative infiltration (cm) per time (Figure 4). Once the

infiltration rate is steady, the sediment is assumed to have become saturated. A trend line was added to the linear (steady state) section of the plots. The slope of the trend line represents the field saturated hydraulic conductivity of the sediments in cm/s which was then converted to m/d (Hillel, 1982). This procedure may lead to overestimates of vertical saturated hydraulic conductivity due to the sediment cracking and creating preferential pathways when the infiltrometer was inserted (although care was taken to prevent this) and because of non-vertical flow below the base of the infiltrometer at 5 cm depth.

Table 2. Measured heads and times from a single-ring falling-head infiltration experiments.

Measured Head (cm)	Measured Time (s) at depth:					
	84 cm	55 cm	30 cm	22.5 cm	12 cm	0 cm
8.5	13	0	-	-	-	-
8	63	-	18	-	-	-
7.5	128	85920	49	-	-	-
7	191	103980	91	16	-	-
6.5	258	-	134	30	14	-
6	314	170580	175	43	17	16
5.5	388	-	219	59	21	20
5	447	250979	271	77	24	23
4.5	515	276959	324	97	28	26
4	584	-	379	114	32	28
3.5	652	-	428	134	36	30
3	726	-	493	153	40	33
2.5	778	-	535	171	44	35
2	862	-	614	195	49	37
1.5	934	-	680	219	52	39
1	999	-	754	244	60	40
0.5	1047	-	804	259	-	-
0	-	-	-	-	-	-

- Tests at the sabkha surface (0 cm) and in the loose, dry material below (12 cm) revealed a very high hydraulic conductivity of 204.42 m/d and 99.71 m/d, respectively. These high

conductivities are at least in part due to the dissolution of the soluble minerals encountered by the infiltrating fluid.

- The zone from ~0.15 m to 0.8 m depth is divided into smaller zones with distinctly different hydraulic conductivity. The upper part consists of two zones with conductivities of 21.25 m/d (22.5 cm) and 7.26 m/d (30 cm).
- The horizontally-layered clay from 45-62 cm appeared a strong barrier to water flow. The estimated hydraulic conductivity of 0.01 m/day implies that infiltrating rainwater would take up to 25 days to pass this unit. If this layer is laterally continuous (it was observed in a boring ~100 m northeast of the trench (Figure 20)) it would strongly impede infiltration.
- The hydraulic conductivity measured in the sabkha sand (84 cm) is 6.57 m/d.

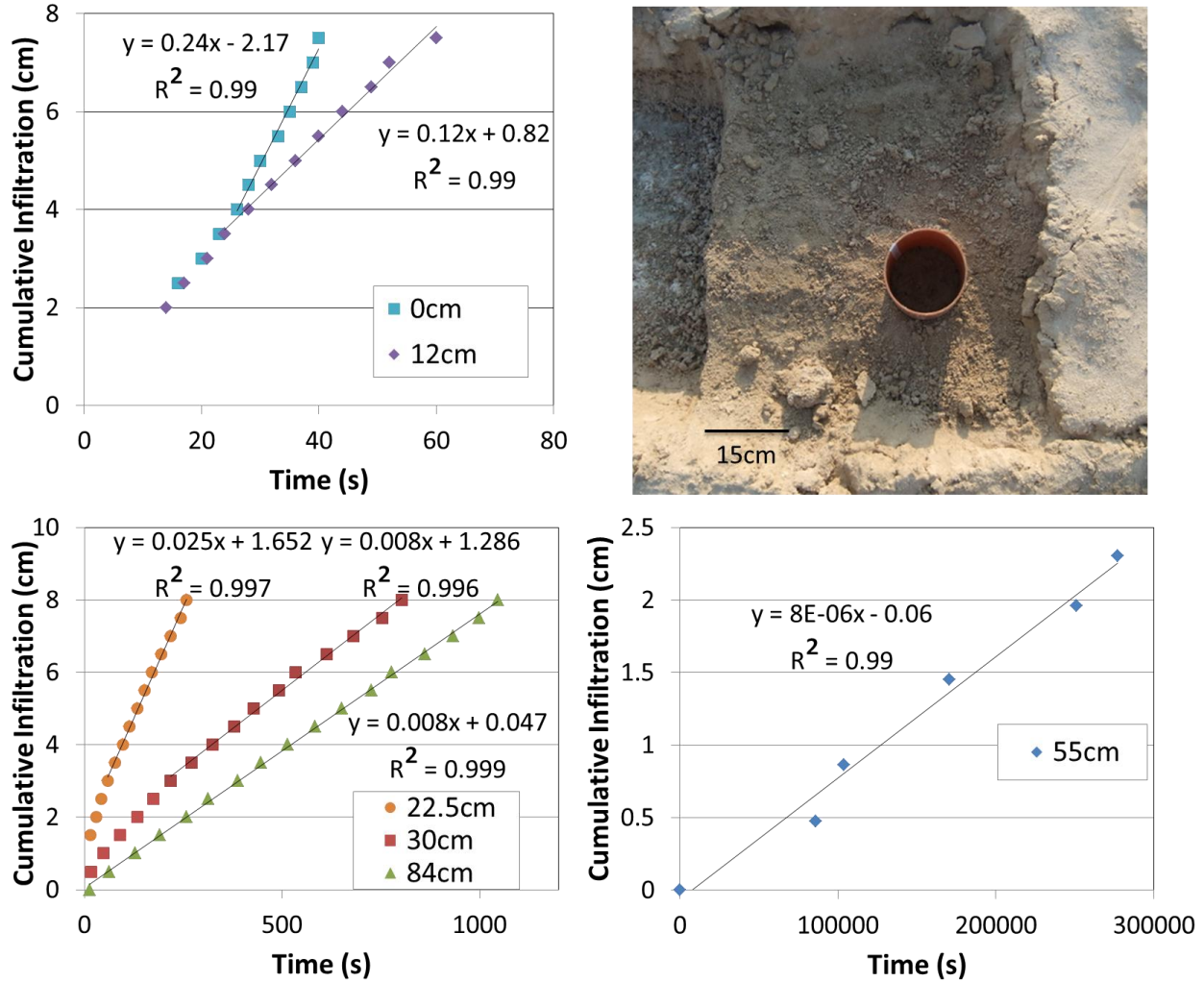


Figure 4. Picture of single-ring falling-head infiltrometer and graphs of cumulative infiltration with time for depths of 0, 12, 22.5, 30, 55, and 84 cm. Linear regression lines and equations given for steady-state conditions.

Using the sediment layer thicknesses and the measured hydraulic conductivities, it is possible to estimate the length of time it would take infiltrating water to travel from the surface of the sabkha down to the water table (Table 3). Values in the table are approximate because the depths of hydraulic conductivity measurements do not all correspond exactly with sedimentary layer boundaries (Figure 3). It is estimated that a molecule of infiltrating water will take around

593 hours (~25 days) to reach the water table. This value is dominated almost entirely by the very low hydraulic conductivity of the clay layer from 45 to 62 cm depth.

Table 3. Layer thickness and hydraulic conductivity to estimate travel time through the capillary zone.

Measurement Depth (cm)	Approximate Layer Thickness (cm)	Measured K (m/d)	Estimated Travel Time (hours)
0	12	204.42	0.014
12	10.5	99.71	0.025
22.5	7.5	21.25	0.085
30	15	7.26	0.496
55	17	0.01	590.3
84	68	6.57	2.49
Total	130		593.4

CONCEPTUAL MODELS FOR CREATING A DENSITY INVERSION

Introduction

Three major conceptual models have been proposed for the formation of a density inversion in the UAE sabkha. The first involves a precipitation-induced dissolution of the halite crust, which infiltrates through the capillary zone into the aquifer. The second proposes that evapoconcentration leads to a buildup of a high salinity layer near the surface of the aquifer (Van Dam et al., 2009). The third suggests that the slow upwelling of low-density water from the Miocene formation below the aquifer creates the inversion.

A major factor in these conceptual models is that they must create a large enough density inversion to cause the system to become unstable causing free convection to occur. Stability of layered fluids is calculated using the Rayleigh number; a dimensionless value describing the

ratio of buoyancy driven flow to dispersive and viscous forces (Rayleigh, 1916). The Rayleigh number is calculated using the equation:

$$Ra = \frac{\text{Buoyancy and gravity driving forces}}{\text{dispersive and diffusive forces}} = \frac{gk\beta\Delta CH}{\theta v_0 D_L} \quad (2)$$

where:

g is the gravitational acceleration (m/s^2);

k is the permeability (m^2);

β is the linear expansion coefficient for density;

ΔC is the difference in concentration by mass fraction = $C_{\max} - C_{\min}$ (kg/kg);

H is the thickness of aquifer (m);

θ is the porosity (dimensionless);

D_L is the fluid coefficient of molecular diffusion (m^2/s).

v_0 is the kinematic viscosity = μ_0/ρ_0 (m^2/s);

μ_0 is the dynamic viscosity ($\text{kg/m}\cdot\text{s}$);

ρ_0 is the underlying water density (kg/m^3);

To evaluate the stability of a system, a critical Rayleigh number $Ra_c = 4\pi^2$ was derived for an infinite horizontal layer with uniform porosity, no flow, and constant temperature boundaries (Lapwood, 1948). When a system's Rayleigh number is below Ra_c , it is considered stable and

diffusive forces control solute movement. A system with a Rayleigh number higher than Ra_c is likely to be unstable as buoyancy forces dominate and free convection occurs.

Precipitation

Density inversions may be created in the sabkha through the dissolution of the halite crust by rain water and the subsequent infiltration to the water table. Concentration differences measured between ponded surface water (rain water that has dissolved the halite crust) and the aquifer were found as high as 250,000 mg/L (Wood et al., 2002) with a $Ra=2.5 \times 10^5$, a value four orders of magnitude higher than the critical value for convection. For this model to work, however, there needs to be enough halite in the system for the rainwater to dissolve so that its concentration and density become high enough to create an inversion.

Ion chromatography measurements of water samples collected in 2010 show that sodium and chloride are the dominant ions in solution, making halite the most common salt to precipitate at the surface (Table 4). Sodium, the limiting ion for halite in the sabkha, has a concentration of 92,802 mg/L. If we take a 1 m by 1 m area of the surface, based on the evaporative flux rate measured to be between 50-88 mm/yr by Sanford and Wood (2001), there will be 50-88 L of water evaporated per year. The amount of halite precipitated can be estimated by taking 50 L X 92,802 mg/L = 4.64 kg Na (8.17 kg Na using 88 L of water). The equivalent mass of precipitated Cl will be 7.16 kg (12.6 kg). Dividing the mass of precipitated NaCl by the density of 2.16 g/cm^3 , we end up with 5461.0 - 9611.3 cm^3 of halite per year in the 1x1 m square, which is equivalent to a deposition rate of 5.48-9.64 mm/yr. Using the Pitzer equation in PREEQC (Parkhurst and Appelo, 1999), the estimated deposition rates, and the solubility of halite, there

would need to be between 32-57 mm/yr of precipitation to dissolve the halite at the same rate as it is being deposited. This indicates that with an average rainfall of 32 mm/yr there should be a surplus of dissolvable halite in the system of up to 4.5 mm/yr.

Table 4. Ion chromatography analysis results in ppm for 5 field samples collected from depths between 1.6 and 2.5 m (sampling sites are shown in Figure 20).

Sample	Cl	Br	NO ₃ N	SO ₄	Ca	Mg	Na	K
2-B	149786	130	40	1773	3702	8116	94166	3605
3-B	144466	124	39	1913	3638	7834	92110	3455
4-B	145117	126	39	1947	3654	7793	92156	3430
5-B	145748	133	38	1979	3734	7944	93264	3512
6-B	146679	126	39	1928	3663	7886	92315	3487
Average	146359	128	39	1908	3678	7915	92802	3498

Evapoconcentration

The second conceptual model for the development of density inversions in this system is evapoconcentration. As the water is being evaporated from sabkha, its salts are being left behind, which in turn increase the concentration of the fluids near the top of the aquifer. This progressive increase of concentration near the top of the aquifer, if occurring at a faster rate than diffusive mixing, could eventually cause the system to reach a critical concentration gradient and induce convection (Nield et al., 2008).

Upwelling

The third conceptual model for the formation of density inversions is vertical transport of less dense fluid into the aquifer from the formation below. Water with a concentration difference of approximately 185,000 mg/L is upwelling from the Tertiary formation below the aquifer at a

rate of ~4 mm/yr. The Rayleigh number for the sabkha and the underlying formation is 1.92×10^5 and is well above the Ra_c of $4\pi^2 \approx 39.5$ and considered unstable enough for convection to occur.

GEOPHYSICS

Methods

Electrical resistivity imaging is a proven method for detecting variations in groundwater salinity. Electric current is easily conducted through the pore water by electrolytic conduction (Reynolds, 1997). Conductivity of the pore water, the electrolyte, is related to its total dissolved solids (TDS) concentration. The more dissolved ions in solution the higher the conductivity. Since electrical resistivity is the inverse of conductivity, the higher the conductivity the lower the resistivity. The differences in TDS for the precipitation and upwelling conceptual models have been measured at $\sim 10^5$ mg/L. These large variations in salinity would be detected as variations in electrical resistivity.

The field site was outfitted with two perpendicular electrical resistivity lines, A and B, with 84 electrodes each, at spacing of 1m. Another 84 electrode line, line C, was set parallel to line B with an electrode spacing of 0.5 m (Figure 1C). Data were collected using two electrode configurations, dipole-dipole and pole-pole arrays. Pole-pole surveys, which are relatively fast but have limited sensitivity to lateral variations in resistivity, were used for vertical soundings. These soundings confirmed the hydrostratigraphic model for the field site (Van Dam et al., 2009).

For the purpose of this study, the dipole-dipole array (Figure 5) was chosen because of its high sensitivity to lateral variations in resistivity (Reynolds 1997). A disadvantage of this type of array is that it has a lower signal to noise ratio, requiring longer measurement times. The survey was set up using Advanced Geosciences, Inc. (AGI) SS Command Creator. The command file had a maximum n length of 8 m and a maximum dipole length of 5 m resulting in 2580 image points along each line.

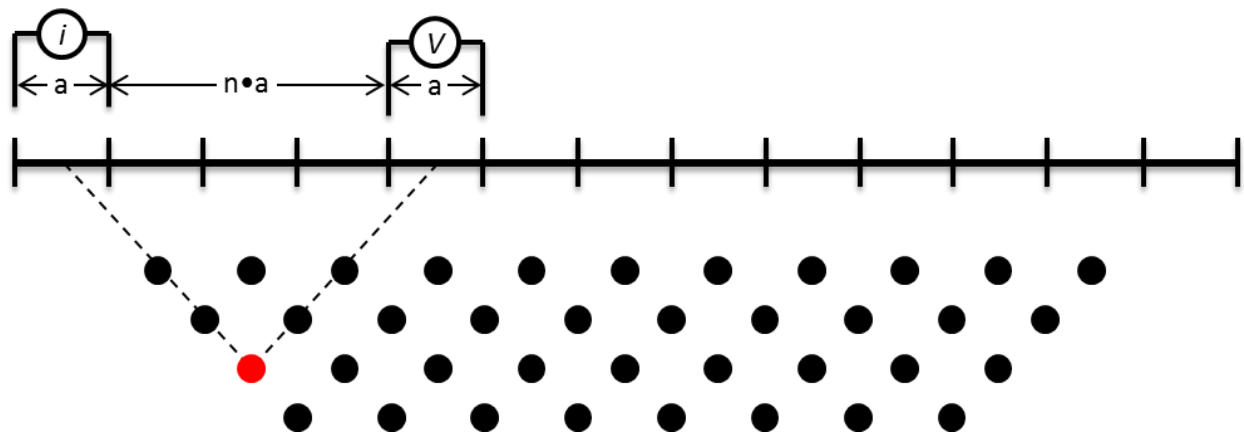


Figure 5. Illustration of dipole-dipole array, showing electrode locations, spacings, and image points.

Results

Inversions of the electrical resistivity data were performed using Advanced Geosciences EarthImager2D (Figure 6) Initial inversion settings for the software were tested and optimized using simplified models of the subsurface resistivity distribution. Initial settings and preprocessing steps are listed in Appendix A. 1D resistivity profiles were created from the 2D inversion results of lines A and B by taking the average resistivity of the middle 40m plotted versus depth (Figure 6b and 6c).

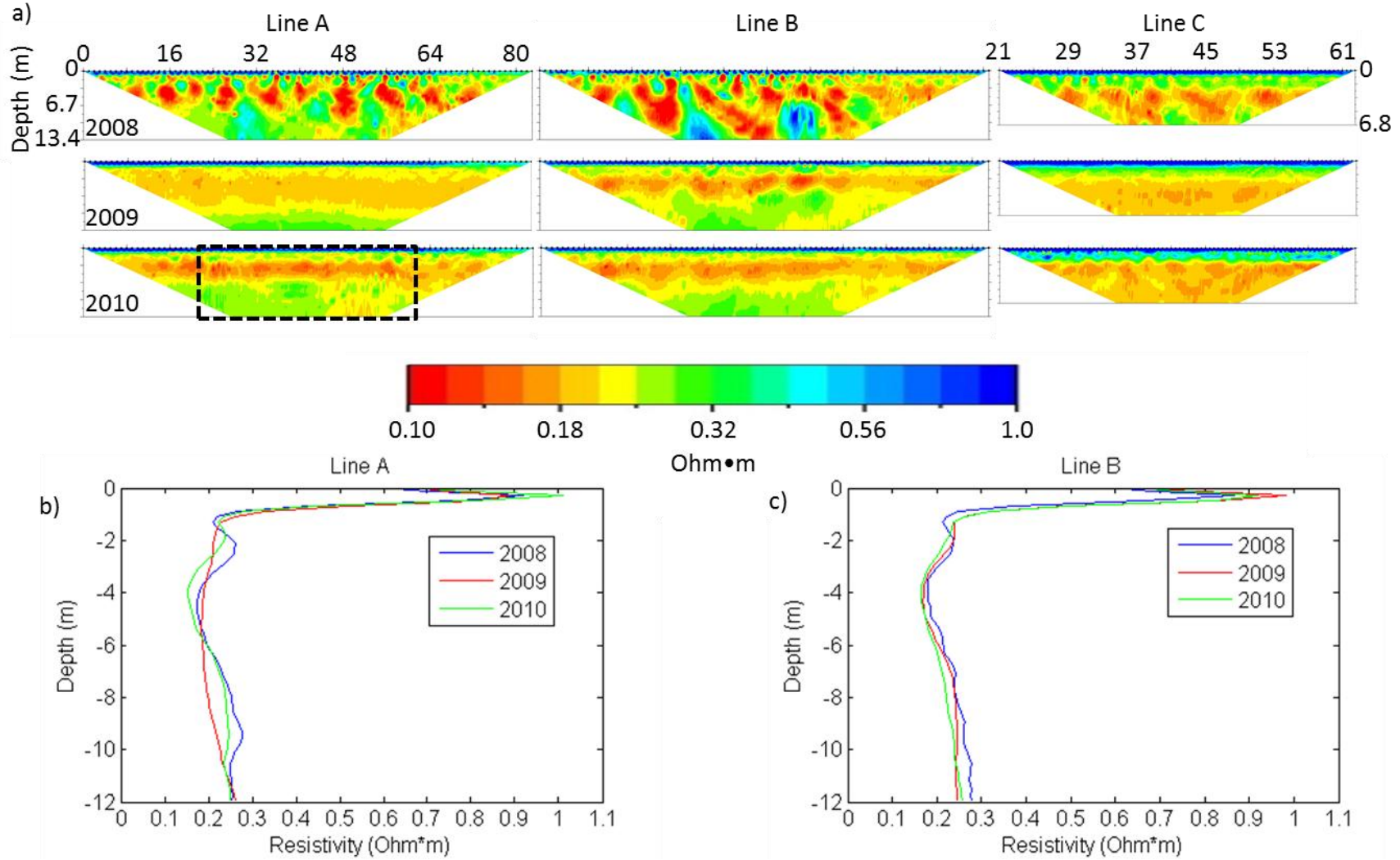


Figure 6. Inverted resistivity models for data from lines A, B, and C collected in 2008, 2009, and 2010 (a). 1D profiles of spatially averaged resistivity with depth for lines A (b), and B (c). The area used to generate the 1D profiles is marked with black dashed lines. See Fig. 1 for locations.

1D Profile Analysis

The 1D profiles illustrate that the average resistivity does not exhibit much change from year to year. This suggests that the average salinity profile stays fairly constant. The shallow values have a maximum resistivity of around 0.9-1.0 Ohm•m, which are significantly smaller than those measured in the soil pit of around 70 Ohm•m. This inconsistency is because the smallest dipole separation of 1 m in the surface resistivity survey does not have the resolution to image such shallow features.

2008 Data Analysis

The three inverted dipole-dipole resistivity lines collected in 2008 (Figure 6a) each display similar features. The top 75-80 cm shows the highest resistivity with values around 1.0 Ohm•m. This layer of high resistivity is representative of the capillary zone. Below the capillary zone the resistivity sharply drops and vertical structures of alternating resistivity of around 0.1 and 0.3 Ohm•m start to appear. Most of the vertical structures of low resistivity extend to a depth of around 7 m; however a few reach close to the bottom of the profile. The shallow vertical structures of low resistivity have a wavelength of 6.5-7 m while the deeper structures display an irregular pattern with a wavelength between 10-15 m where present.

2009 and 2010 Data Analysis

The inverted electrical resistivity profiles from the 2009 and 2010 field campaigns are all very similar (Figure 6a). Change in resistivity occurs vertically with depth and there is very little lateral variation. Like the 2008 data, the layer from the surface to 80 cm has the highest resistivity and represents the capillary zone. Below that, the resistivity decreases to a value of

around 0.25 Ohm•m for ~1 m. The next 5-6 m have the lowest resistivity with an average value of around 0.19 Ohm•m. From about 7 m depth to the bottom of the profile, the resistivity stays around 0.25 Ohm•m.

Discussion

The 2008 inverted resistivity models shows significant lateral variations in resistivity (Figure 6a), which is consistent with unstable fingering. The vertical low resistivity features appear to be high density fingers that are descending through the aquifer while the high resistivity features are ascending lighter fluids. There are no obvious vertical features and very little lateral variation in resistivity in the 2009 and 2010 data (Figure 6a). This lack of lateral variation in resistivity suggests concentrations in the groundwater are only changing vertically between strata and convective fingering is not occurring. Analysis of the geophysical data thus suggests that free convection is not occurring continuously in this system. The episodic nature of the convection along with the 1D and 2D profiles give some insight into which conceptual model is creating the density inversion.

Evapoconcentration in an arid region would likely be a continuous process of building up high concentration brines at the surface of the water table. If evapoconcentration is what triggers free convection, it is likely that there would be a layer of low resistivity just below the capillary zone present in the 2008 data or building back up by the 2010 data; however, this is not the case. Evapoconcentration was initially proposed as a conceptual model because the results published by Van Dam et al. (2009) displayed a layer of low resistivity below the capillary zone (Figure 13). The layer of low resistivity seen in their data was shown to be an artifact of the inversion process and does not actually exist (Appendix A). Additionally, the presence of a halite crust

and the high measurements of TDS in ponded rain water suggest that the bulk of the salts are being precipitated out of solution when it is evaporated instead of being left behind in the aquifer. The combination of the lack of a low resistivity layer and the presence of a halite crust suggest that evapoconcentration is not a likely mechanism for creating a density inversion.

Upwelling from the formation below the aquifer can be seen in the 1D and 2D profiles as the gradual increase in resistivity from around 4 m to 12 m depth (Figure 6a,b, and c). This gradient is present during all three years even though free convection is not, suggesting that although it may contribute to the density gradient it is likely not the mechanism that initiates convection.

Dissolution of the halite crust by rainwater that becomes saturated with solutes and infiltrates to the water table is another possible mechanism for the creation of a density inversion. Figure 7 shows daily temperature, daily precipitation in mm, and geophysical data collection periods since 2007. Two major precipitation events of 50 mm and 64 mm took place 18 weeks and 6 weeks, respectively, before the first data collection period in March 2008. Precipitation was insignificant in the year prior to the 2009 field campaign. There was a total of 18 mm of rain with the maximum event being 4 mm. Three large precipitation events of around 42 mm each occurred 47, 36, and 7 weeks before the third data collection period in 2010.

A conservative halite solute budget based on evaporative flux as described earlier shows that the two rainfall events before the 2008 data collection and the rainfall event 47 weeks before the 2010 data collection would be able to dissolve enough halite to make the system unstable. The other two large events before the 2010 data collection would likely not have enough halite to dissolve in order to reach a density inversion.

Based on the amount of dissolvable halite in the system and the timing of the precipitation events in relation to the data collection campaigns it is likely that free convection in the sabkha is precipitation-induced. A rainfall of 50 mm has the potential to become completely saturated with dissolved halite. Completely saturated rainwater at 27.4 °C (average air temperature) would have a sodium concentration of ~140,000 mg/L, which, when compared to sodium concentrations of 92,802 mg/L measured in the aquifer, illustrates the potential for a density inversion. Assuming that the rainfall event 18 weeks prior to the 2008 data collection is responsible for fingering and taking into account the ~25 days for infiltration through the capillary zone (Table 3), it can be calculated that the fingers had around 15 weeks to develop.

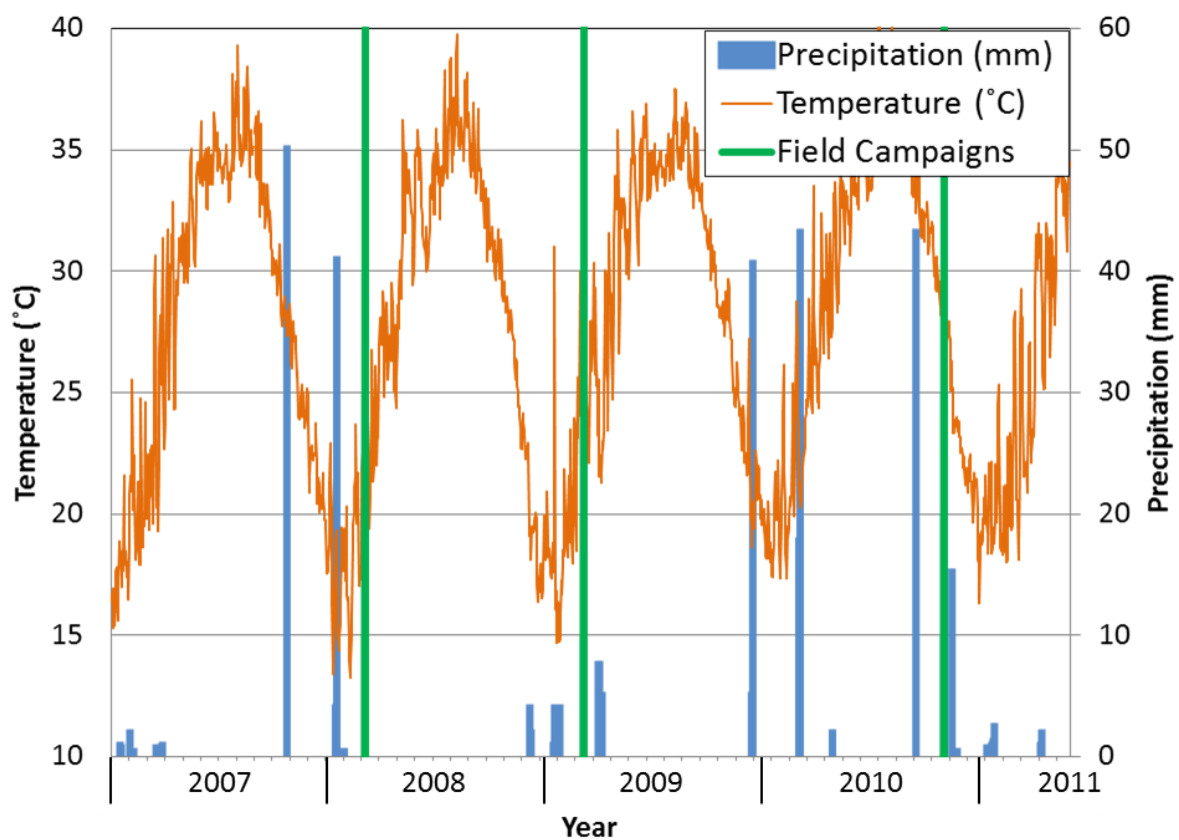


Figure 7. Graph of air temperature, daily precipitation, and data collection periods from 2007-2011.

MODELING

Introduction

The geophysics appears to demonstrate that free convection is occurring in the sabkha, but also that it is not a continuous process. Because the geophysics only offers snapshots in time, numerical modeling of the sabkha system was done to investigate the transient nature of convective fingering. Modeling was performed using COMSOL Multiphysics software (COMSOL Multiphysics, 2011).

Governing Equations

Fluid flow in a porous media is controlled in this model by applying Darcy's law:

$$u = -\frac{\kappa}{\mu}(\nabla p + \rho g \nabla D) \quad (3)$$

where:

u is the Darcy velocity or specific discharge vector (m/s);

p is the fluid pressure (Pa);

μ is the dynamic viscosity (kg/m•s);

g is the magnitude of gravitational acceleration (m/s²);

∇D is a unit vector in the direction over which the gravity acts (m).

κ is the permeability (m²); calculated by

$$\kappa = K \frac{\mu}{\rho_0 g} \quad (4)$$

K is the hydraulic conductivity;

ρ is its density (kg/m^3); calculated by

$$\rho = \rho_0 + \beta c \quad (5)$$

ρ_0 is the initial groundwater density (kg/m^3);

β is the linear expansion coefficient;

Darcy's law is then inserted into a continuity equation:

$$\frac{\partial}{\partial t}(\rho\theta) + \nabla \cdot (\rho u) = Q_m \quad (6)$$

where:

θ is the porosity;

Q_m is a mass source term ($\text{kg}/(\text{m}^3 \cdot \text{s})$).

This results in:

$$\frac{\partial}{\partial t}(\rho\theta) + \nabla \cdot \rho \left[-\frac{\kappa}{\mu} (\nabla p + \rho g \nabla D) \right] = Q_m \quad (7)$$

Since the model assumes that solute transport is taking place in a fully saturated medium and that the solutes are conservative (thus ignoring adsorption and desorption terms) transport can be modeled using a simplified advection-dispersion equation:

$$\frac{\partial \theta_S c}{\partial t} + \nabla \cdot cu - \nabla \cdot [D_D + \theta_S \tau D_L \nabla c] = 0 \quad (8)$$

where:

θ_s is the fluid volume fraction (porosity);

c is the salt concentration (kg/m^3);

u is the Darcy velocity (m/s);

τ is the fluid phase tortuosity factor $\tau = \theta^{1/3}$;

D_D is the dispersion term;

D_L is the fluid diffusion coefficient (m^2/d).

Conceptual Models

Precipitation Induced. The sabkha system was modeled in 2D using the hypothesis that free convection is precipitation-induced. For simplicity, the starting conditions of the model assume that a rainfall has already become saturated from the dissolution of the halite crust and has infiltrated through the capillary zone to the top of the water table. The model domain is divided into two layers. The bottom layer is 10m thick and represents the sandy aquifer. The top layer's thickness is dependent on the amount of precipitation that occurred divided by the porosity (0.38) so that it represents a completely saturated layer that is an extension of the aquifer. For example, a rainfall event of 0.10 m would result in a top layer with a thickness of 0.26 m. For the Darcy's law interface, the sides and bottom are considered to be no flow boundaries. Zimmerman et al. (2006) have modeled the effects of hydraulic gradients on the formation of convective fingering. Their results demonstrate that low gradients (~ 0.0001) have almost no effect on finger development for a site where the concentration difference and a Rayleigh number are each an order of magnitude smaller than that of the sabkha. Lateral flow in the sabkha based on a hydraulic gradient of 0.0002 is 1.26 m/yr and is insignificant compared to

the velocity of vertical density driven flow and will therefore not be considered in the model. Additionally, the influence on the system from the upwelling of 4 mm/yr from the formation below the aquifer is orders of magnitude smaller than that of a large precipitation event and was therefore ignored. The top boundary is treated as an atmospheric pressure boundary and the initial pressure of the entire domain is equal to

$$p_0 + \rho \cdot g \cdot L \quad (9)$$

where:

p_0 is the initial pressure (Pa•s);

ρ is the calculated density (kg/m^3);

g is the gravitational acceleration (m/s^2);

L is the domain height (m).

The solute transport calculations treat all four boundaries as no flux boundaries; therefore, no solutes may be transported across them by advection, dispersion, or diffusion. The top layer in the domain represents the infiltrating water with the starting concentration c_s and density ρ_s and the bottom layer, the aquifer, begins with c_0 and ρ_0 .

Upwelling Induced. A second model was created based on the hypothesis that upwelling of less dense fluid from the Miocene formation below the aquifer is causing free convection. The model consists of one domain, a 10 m by 30 m sandy aquifer. Just as in the precipitation induced model, the top boundary is an atmospheric pressure boundary and the side

boundaries are no flux and no flow. The bottom boundary of the model was assigned an inward mass flux based on the concentration and density of the upwelling water (Table 5) at a rate of 4 mm/yr.

Simulation Results

Precipitation Induced. The starting model assumes an infiltration layer 50 mm thick to simulate a rainfall event similar to the one that coincided with the 2008 data collection. Model parameters are based on values measured in the field in an attempt to simulate the sabkha aquifer as accurately as possible. Table 5 lists the values used in the model and their sources.

Table 5. Parameters, values, and sources for COMSOL modeling.

Parameter	Value	Unit	Source
Domain height (average sabkha depth) (L)	10	[m]	Sanford and Wood (2001)
Density of aquifer water (ρ_0)	1180	[kg/m ³]	Wood et al.(2002)
Density of infiltrating water (ρ_s)	1305	[kg/m ³]	Wood et al.(2002)
Density of upwelling water (ρ_u)	1075	[kg/m ³]	Wood et al.(2005)
Concentration of aquifer water (c_0)	275.8	[kg/m ³]	Wood et al.(2002)
Concentration of infiltrating water (c_s)	440	[kg/m ³]	Wood et al.(2002)
Concentration of upwelling water (c_u)	101.5	[kg/m ³]	Wood et al.(2005)
Dynamic viscosity (μ)	0.001	[kg/m s]	Assumed for fresh water
Hydraulic conductivity (K)	6.57	[m/d]	Measured
Permeability (κ)	6.566×10^{-12}	[m ²]	Calculated from K, μ , ρ_0 , and g
Porosity (θ)	0.38		Wood et al.(2002)

Table 5 (cont'd)

Coefficient of molecular diffusion (D_L)	1×10^{-9}	$[m^2/s]$	Simmons et al. (2001)
Dispersivity (α) (lateral and transverse)	0.08	$[m]$	Schulze-Makuch (2005)
Linear expansion coefficient (β)	0.8384		Calculated
Pressure (p)	0	$[atm]$	Assumed for sea-level
Acceleration of gravity (g)	9.81	$[m/s^2]$	Constant
Upwelling flux rate (q_u)	4.0	$[mm/yr]$	Sanford and Wood (2001)

The model simulation ran for 52 weeks. Fingers of high concentration quickly developed at the top and begin descending towards the bottom of the model (Figure 8). In the first weeks, the model developed approximately 28 small fingers. Over time, the small fingers began to coalesce into fewer and larger fingers. At the end of 10 weeks there were 18 fingers and by 25 weeks there were 13 fingers (Figure 9).

The rate of finger descent was relatively fast at first, but slowed over time due to the dispersion of the solutes (Figure 10). Eventually, as the concentration differences between the fingers and the aquifer became less and less the rate of descent became very slow. After about 30 weeks, advection slowed down to the point where there is little movement of the fingers. At this point diffusive forces are becoming dominant and the fingers begin to disappear.

Convective velocity (U_c) was calculated using the equation:

$$U_c = \frac{K}{\theta} \cdot \frac{\Delta \rho}{\rho_0} \quad (10)$$

where:

U_c is the convective velocity (m/d);

K is the hydraulic conductivity (m/d);

θ is the porosity;

$\Delta\rho$ is the density difference from initial density ($\rho-\rho_0$) (kg/m^3);

ρ_0 is the initial density (kg/m^3).

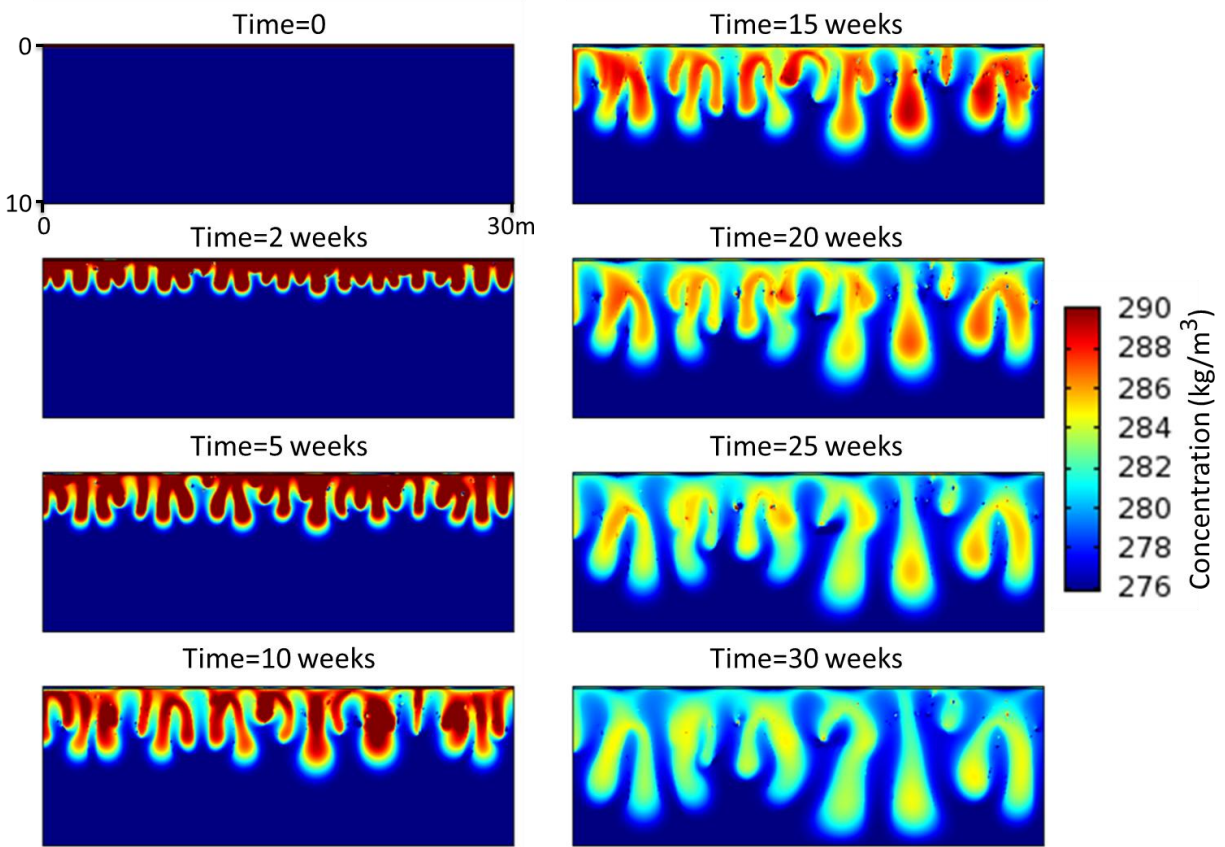


Figure 8. Simulation of groundwater convection from 50mm of precipitation at time $t=0$, $t=2$, $t=5$, $t=10$, $t=15$, $t=20$, $t=25$, and $t=30$ weeks.

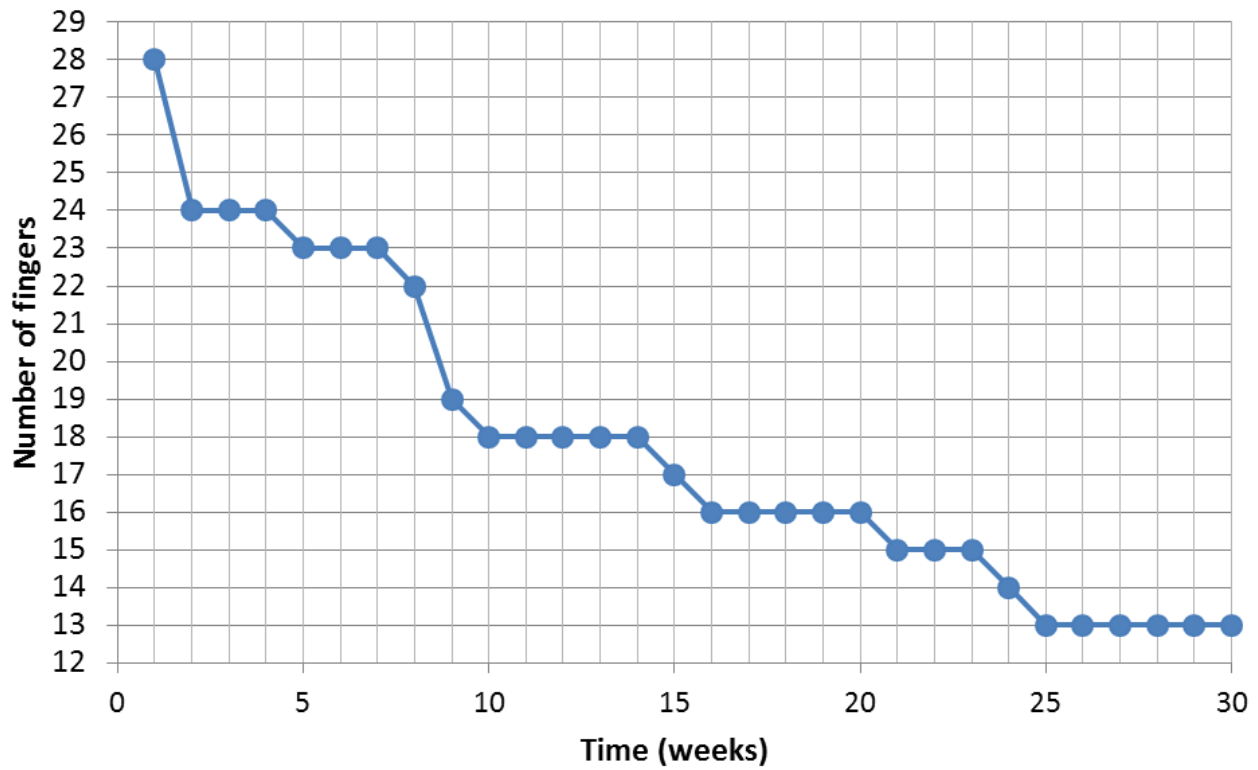


Figure 9. The number of fingers in the model over time as smaller fingers coalesce into fewer and larger fingers.

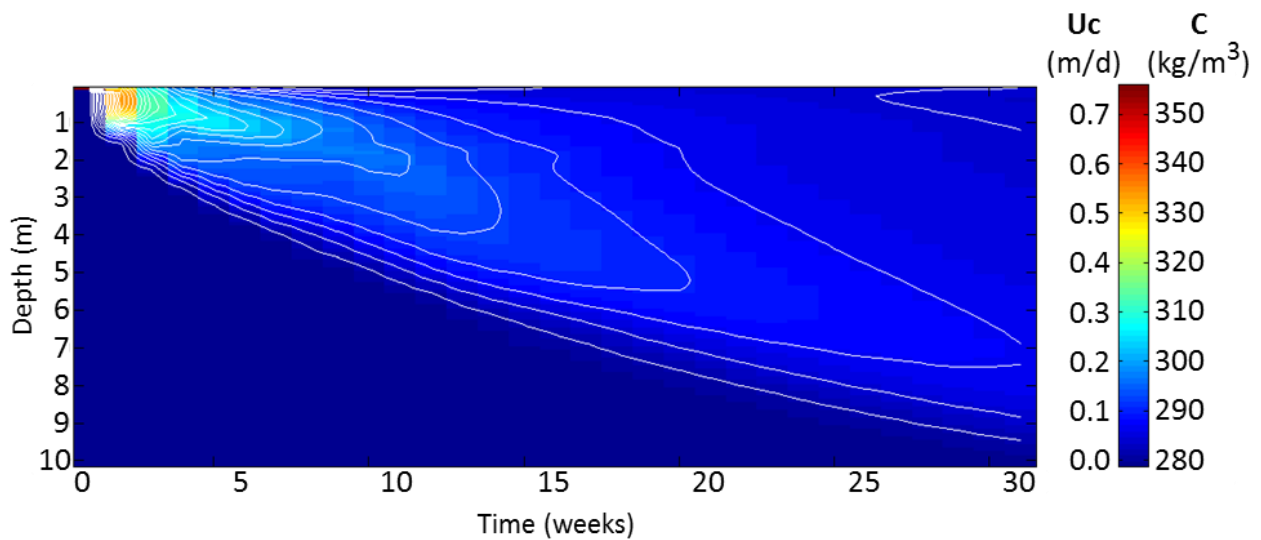


Figure 10. Convective velocity (U_c) and concentration (c) for a single saline finger ($x=21.5$) from model (Figure 8) over time.

Upwelling Induced. The simulation results show that upwelling from below the aquifer at a rate of 4 mm/yr has the potential to initiate free convection (Figure 11). A layer of low concentration fluid continuously builds up at the bottom of the aquifer. The model shows that it takes around 30 to 31 years of upwelling in order to create a layer that is unstable enough for convection to occur.

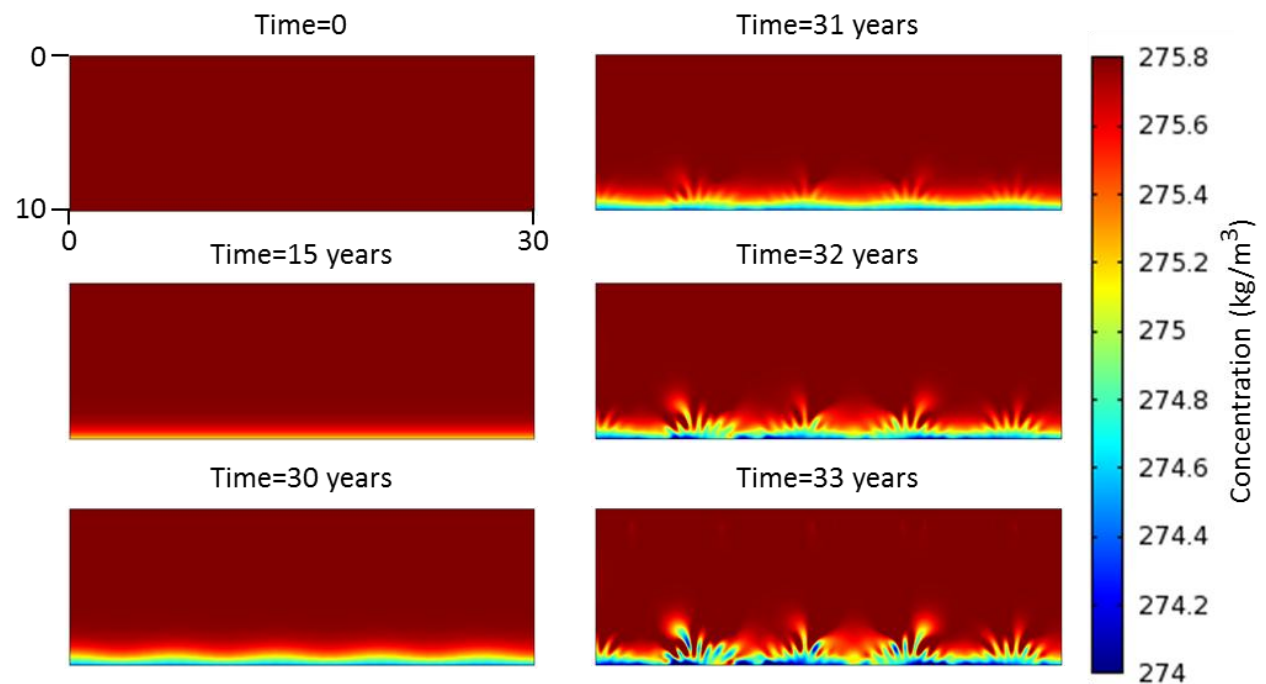


Figure 11. Simulation of groundwater convection caused by upwelling of less dense water at times $t=0$, $t=15$, $t=30$, $t=31$, $t=32$, and $t=33$ years.

DISCUSSION

The hypothesis that free convection is precipitation induced is supported by both the geophysics and the modeling. A comparison between the 2008 field data and the modeling output is shown in Figure 12. As calculated earlier, the convective fingering seen in the geophysics data is assumed to have been developing for a period of about 15 weeks, therefore, it is being compared to the model output at 15 weeks. The modeled finger lengths at 15 weeks

correlate with the features in the geophysical data. The model is not intended to be an exact replication of field data and is illustrating 2D representations of 3D features; however, it does show that fingers can occur, grow, and decay in this system on the timescales that we have observed with geophysical measurements.

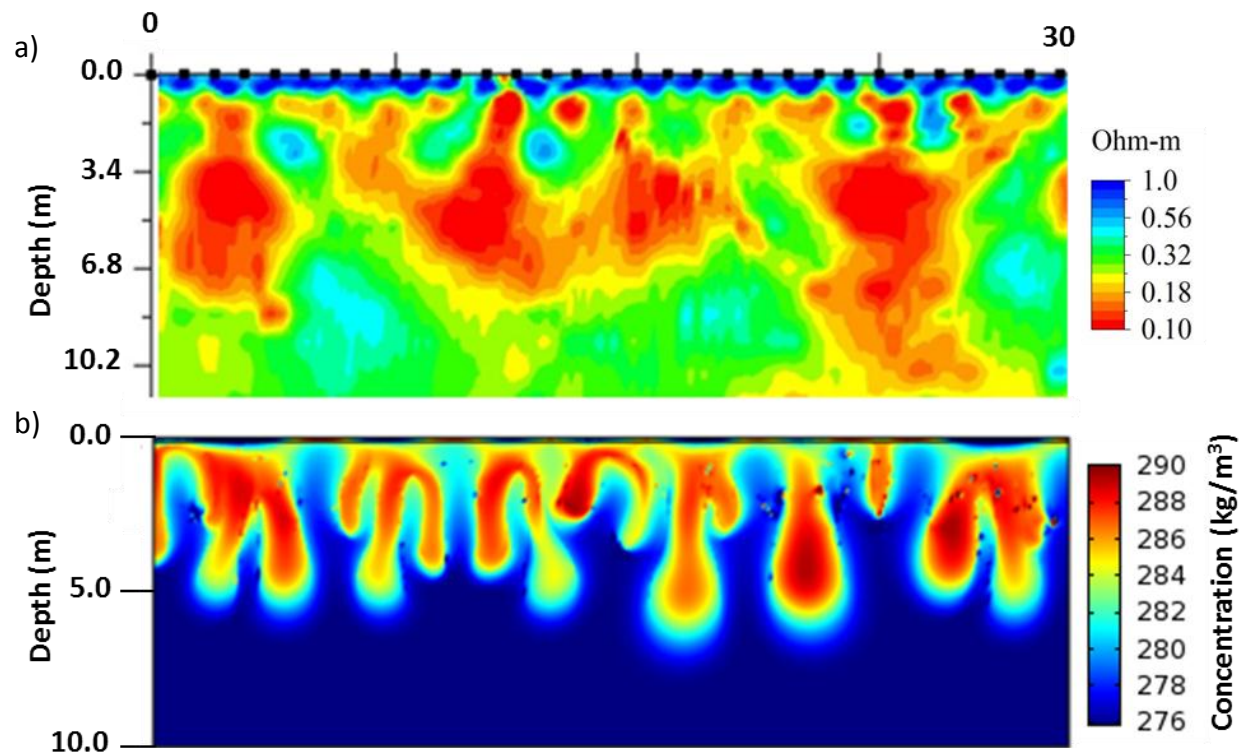


Figure 12. Comparison between 2008 field data for line A (a) and COMSOL model output at time $t=15$ weeks (b).

Secondary evidence suggests that upwelling may contribute to the density inversion, but the geophysics shows that this is likely not the main cause since convection in the aquifer is not a continuous process. Modeling has shown that upwelling is capable of creating density inversions that, given enough time, can become unstable enough to cause free convection. The amount of time it takes for upwelling fluids to make the system unstable enough for convection

make it an unlikely source due to the fact that it would most likely be erased by the faster and more frequent precipitation induced convective mixing.

Evapoconcentration was not evident in the geophysics data and as discussed earlier is an unlikely source because of the presence of the precipitated halite crust. Additionally, Appendix A shows that the array typed and inversion method used is not sensitive enough to detect a thin low-resistivity layer between the water table and the capillary zone.

CONCLUSIONS

The analysis of the electrical resistivity data from three consecutive years at a sabkha in the UAE confirms the hypothesis that free convection is not a continuous event. The episodic nature of the process, the system's high Rayleigh number, the timing of rain events, and the unlikelihood of the other conceptual models strongly suggests that the convection is precipitation-induced.

Modeling of a density inversion based on a rain event of 50 mm and measured field values demonstrates that free convection in this system is possible. The timing for the development of fingers in the model coincides with the timing of a large rain event prior to the 2008 data collection. The decay rate of the fingering and lack of rainfall explains why fingers were not apparent in subsequent data collections.

APPENDICES

APPENDIX A

Preprocessing

Testing for and Removing Artifacts

The inverted resistivity data presented in Van Dam et al. (2009) contains two potential artifacts. The first is a layer of very low resistivity directly below the capillary zone. The second potential artifact is the low resistivity value found at the ground surface, where the resistivity has been measured to be much higher (Figure 3). This appendix will describe the testing process to see if these features were artifacts, and if so, how they could be removed.

The inverted resistivity images of the 2008 data presented in Van Dam et al. (2009) show a layer of high resistivity just below the surface with a layer of very low resistivity directly below it near the water table (Figure 13). Initial data analysis of the 2009 data showed a similar low-resistivity layer, but without the fingering observed in the 2008 data. This layer of low resistivity has been hypothesized to be the accumulation of dense water through evapoconcentration and possibly the driving force of the free convection. Alternatively, it has been suggested that this layer is an artifact of the inversion process caused by the sudden change from the high resistivity capillary zone into the low resistivity aquifer.

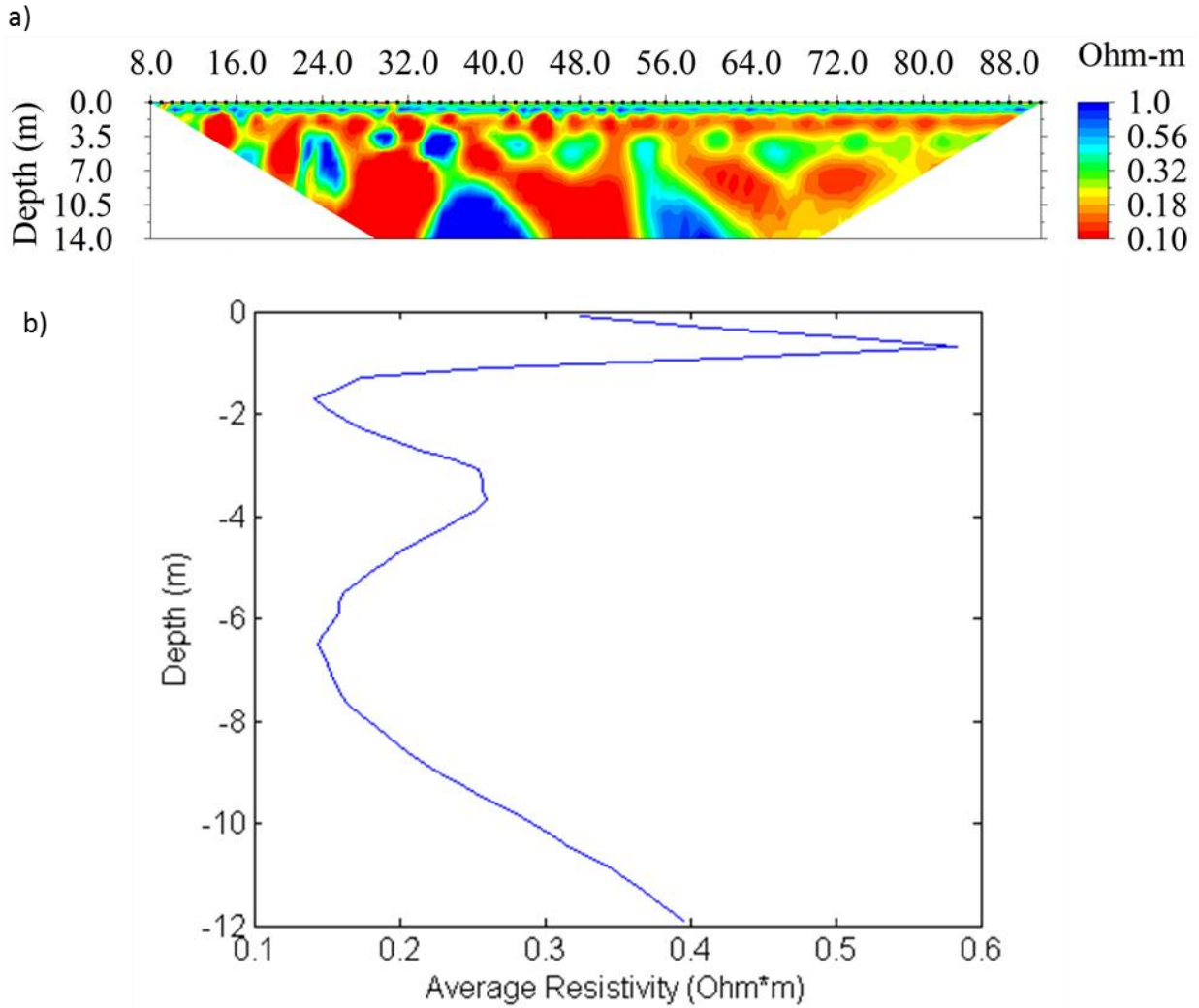


Figure 13. Cross section of inverted resistivity for line B using inversion settings from Van Dam et al. (2009) showing layer of low resistivity below the capillary zone (a) and 1D profile (b) showing average resistivity (Ohm•m) with depth. See Fig. 6 for area used in averaging.

Testing of this layer as a potential artifact was done by creating simple two and three layer models, forward calculation of the synthetic resistivity measurement data (using the same electrode configurations and array combinations), and subjecting them to the same inversion methods as in Van Dam et al (2009). The two layer model contains only the highly resistive capillary zone and the low resistivity aquifer. The three layer model has a layer of even lower

resistivity between the capillary zone and the aquifer to represent a zone of higher salinity due to evapoconcentration (Figure 14).

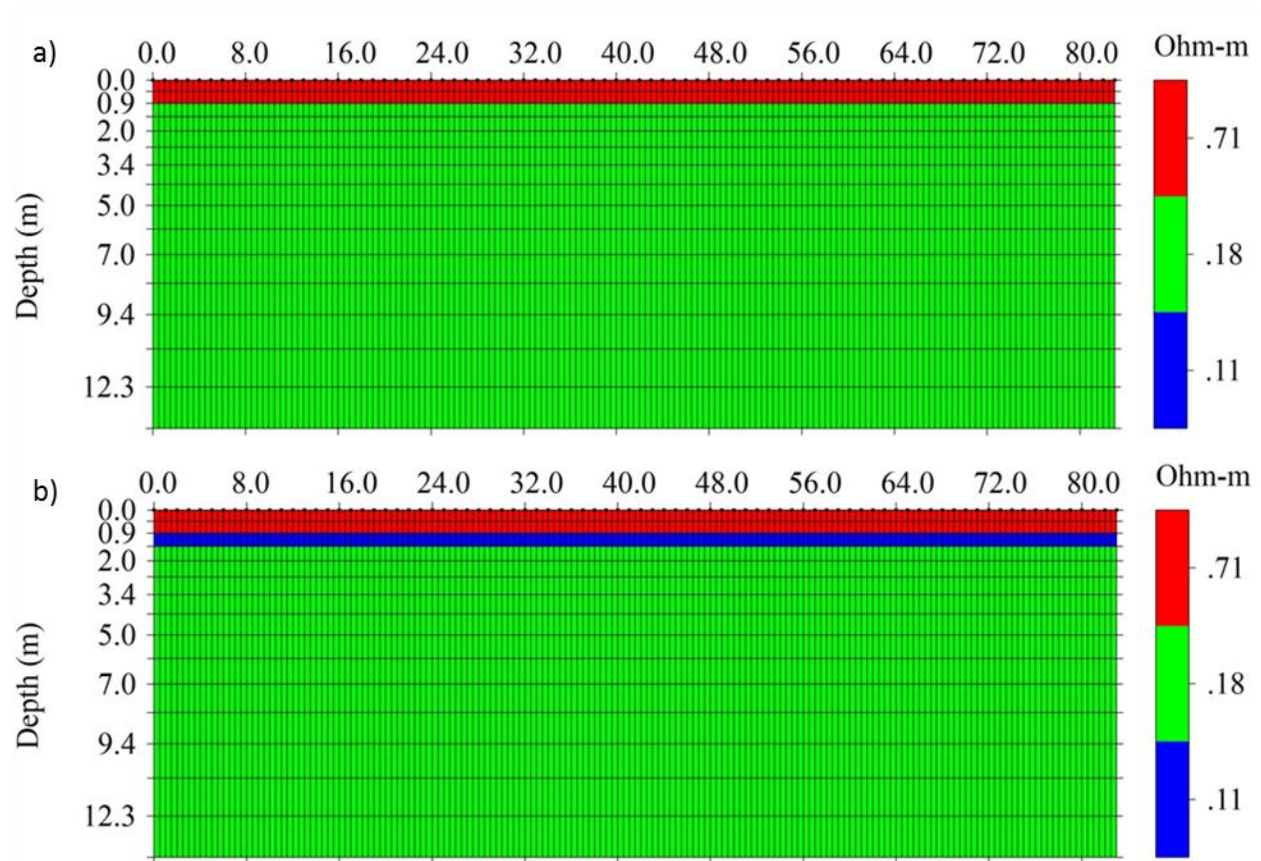


Figure 14. Cross sections of 2-layer (a) and 3-layer (b) models. Both models have a background resistivity of 0.18 Ohm•m, a capillary zone with a resistivity of 0.71 Ohm•m. The 3-layer model has an intermediate layer of low resistivity (0.11 Ohm•m) just below the capillary zone.

For both the two and three layer models, the inverted data show a low resistivity layer followed by a thin layer of high resistivity before returning to the background resistivity value (Figure 15). The occurrence of these layers in both models suggests they are indeed artifacts created during the inversion process. These results, however, do not rule out the possibility of the existence of these layers in the sabkha. Therefore, the next step was to adjust the inversion process in order to remove the artifacts from the models and then use the updated inversion methods to reprocess the data.

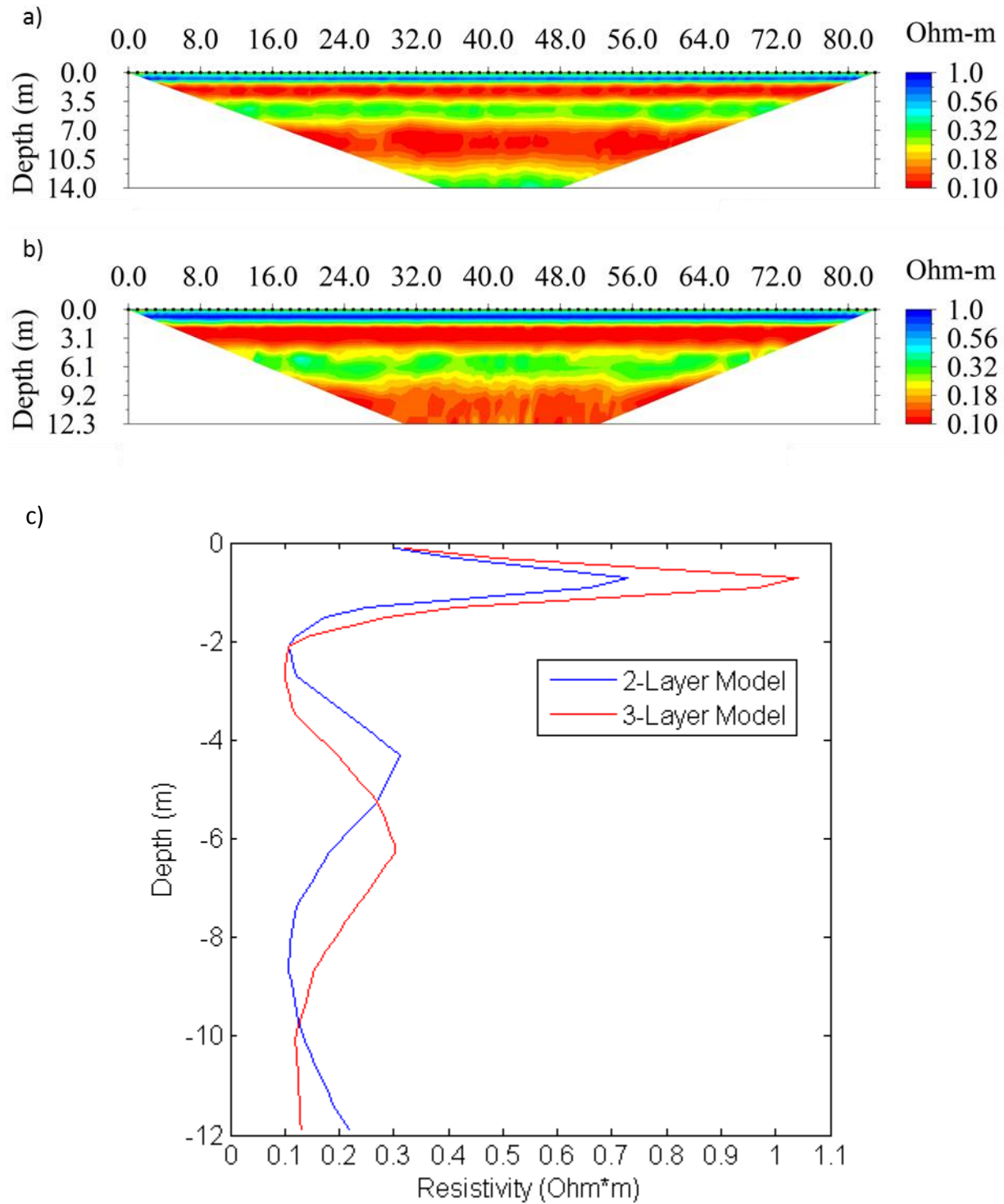


Figure 15. Cross sections of inverted resistivity outputs from the 2-layer model (a) and the 3-layer model (b) both showing a zone of low resistivity below the capillary zone. 1D profile (c) showing average resistivity for middle 40 m of 2- and 3-layer inverted models. See Fig. 6 for area used in averaging. See Fig. 14 for the starting model.

Since the artifacts were likely caused by the way the inversion software handled the sharp boundary from high to low resistivity, the first step that was taken was to refine the mesh discretization. In the initial processing of the data (Van Dam et al., 2009), the number of mesh divisions between electrodes was set at 2. The finer the mesh setting the more accurate the forward models become, but computing time is longer. Since computing time was not an issue, the number of mesh divisions was increased to 8, the highest recommended by the software.

To compare the results of the two different meshes, an average resistivity value was calculated at 20 cm intervals over the middle 40 meters of the inverted resistivity sections. These were then plotted in 1D profiles of depth versus resistivity (Figure 16c and d). When the mesh is set at 2 divisions it creates a sinusoidal type curve that shows high amplitude changes in resistivity with depth, which are reflected in the inverted sections as the layers of low and high resistivity. With the number of mesh divisions set at 8 the 1D profile still show minor sinusoidal curves, however their amplitude is muted and the values are much closer to the expected background values. These results also demonstrate that the method (or array type) used is not sensitive to the thin low-resistivity layer ($0.11 \text{ Ohm}\cdot\text{m}$) in Figure 14b.

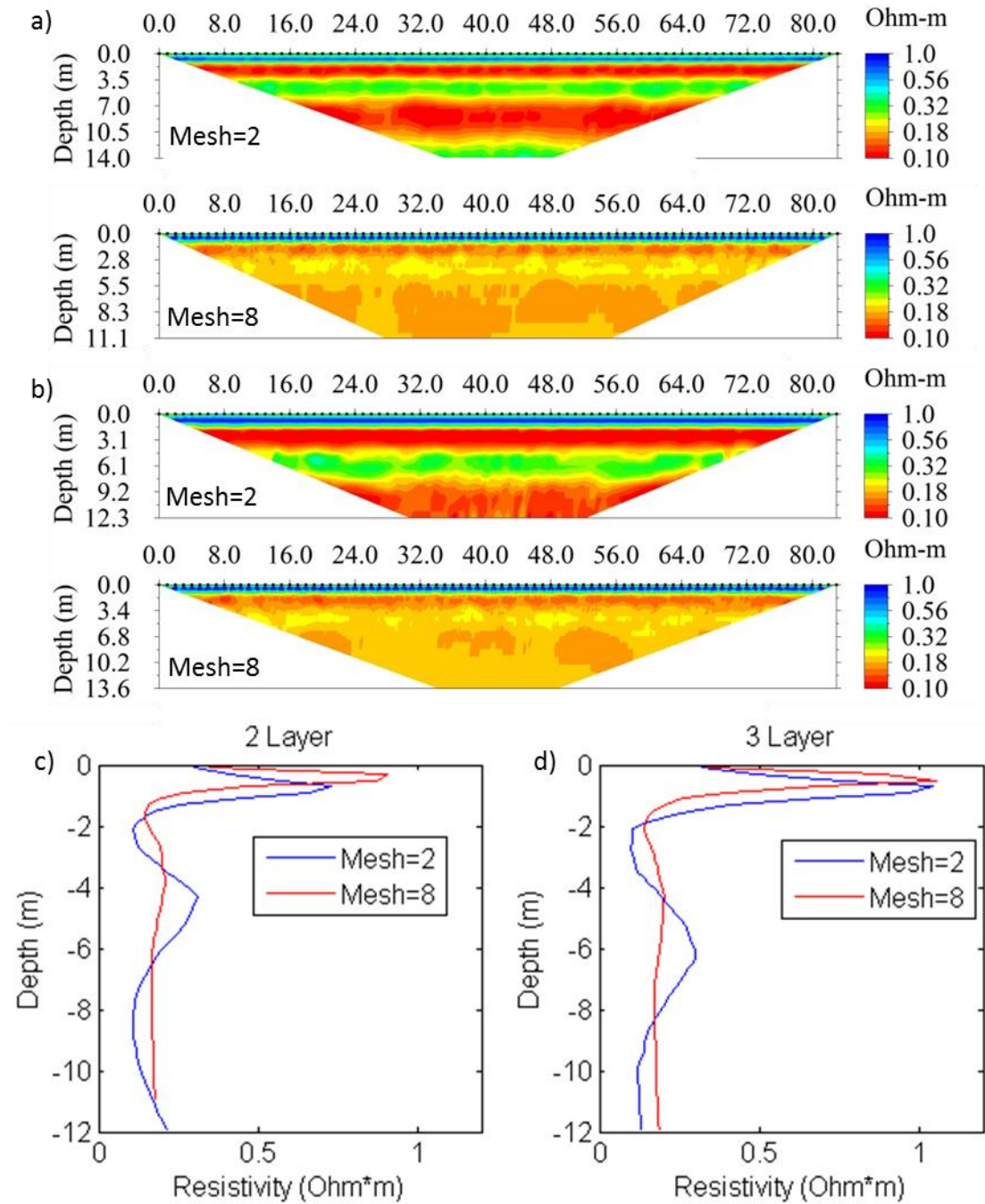


Figure 16. Cross sections of inverted resistivity for 2 layer (a) and 3 layer (b) models for mesh settings of 2 and 8 divisions between electrodes respectively. 1D profiles of spatially average resistivity for 2 layer (c) and 3 layer (d) models. See Fig. 6 for area used in averaging.

The 1D profiles indicate another artifact caused by the original inversion settings that are not consistent with the models or the field site. The inverted resistivity value at the ground

surface is calculated to between 0.2-0.3 Ohm•m while the model's surface layer is set at 0.71 Ohm•m (Figure 16c and d). This error is likely caused by the inversion starting model being set to “average apparent resistivity” with a value of 0.23 Ohm•m which is normally reserved for borehole or noisy surface data sets. With this setting, the inversion process begins with a model that is entirely one apparent resistivity, 0.23 Ohm•m in this case, which it then uses to create the first forward model. During the iteration process a new forward model is created from the previous model and it is possible that the artifact is a numerical edge effect at the surface caused by the starting model.

To investigate if the starting model was the cause of the artifact, the model was run again using “Pseudosection” as the starting model. This setting uses an interpolated pseudosection of apparent resistivity based on the input data as the starting model. 1D profiles created from the new inversion outputs show the resistivities at the surface have a value constant with the models (Figure 17).

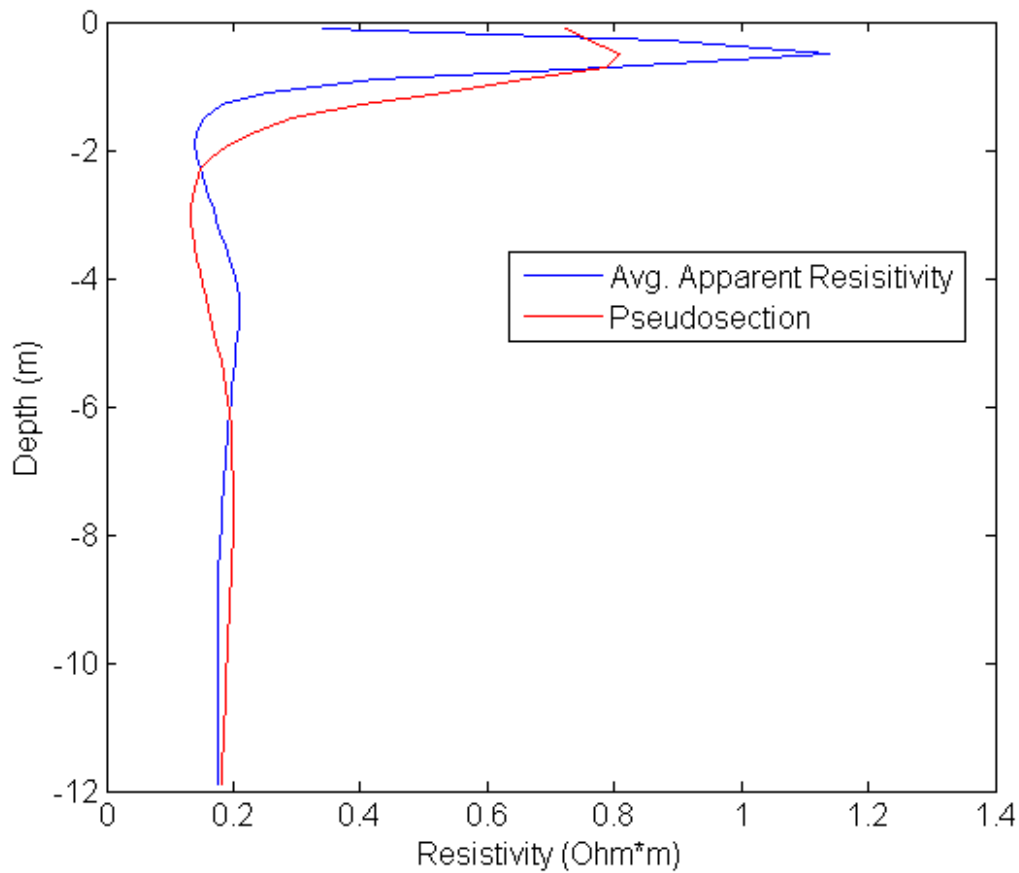


Figure 17. 1D profiles of the two different starting models showing each model's influence on the resistivity at 0 m depth. See Fig. 6 for area used in averaging.

This analysis of the inversion settings has shown that artifacts were created in the way the data was initially processed. By changing the mesh size and using a more suitable starting model, the artifacts have been minimized or removed. For processing the 2008, 2009, and 2010 data in this thesis, the number of mesh divisions will be set at 8 and the starting model will be the interpolated pseudosection. All other settings will be the same as those used in Van Dam et al. (2009).

Inversion Settings

Minimum Voltage (mv) = 0.02
Minimum V/I (Ohm) = 2E-5
Minimum apparent resistivity (Ohm•m) = 0.01
Maximum apparent resistivity (Ohm•m) = 10000
Maximum repeat error (%) = 5
Maximum reciprocal error (%) = 5
Remove negative apparent resistivity in ERT data: No
Keep All Data (no data removal): No
Inversion Method: Smooth model inversion
Vertical axis: Positive Upward
Y Coordinate = Depth
Min electrode spacing X (m) = 0.01
Min electrode spacing Z (m) = 0.01
Forward Modeling Method: Finite element method
Forward system solver: Cholesky decomposition method
Boundary condition type: Dirichlet
Number cells or elements between two electrodes = 8
Lower-layer-thickness / Upper-layer-thickness = 1.1
Depth of Inverted Model / Depth of Pseudosection = 1.1
Max number of iteration of nonlinear inversion = 8
Stop RMS error = 3%
Minimum error reduction between two iterations = 5%
Stop at Max number of iterations: Yes
Stop when RMS is small enough: Yes
Stop when RMS can not be reduced: No
Res Data reweighting: No
Use Reciprocal Error: No
Stop when L2 norm is small enough: Yes
Initial Lagrange multiplier or roughness factor = 10
Roughness conditioner = 0.2
Starting model: pseudosection.
Minimum resistivity = 0.1 Ohm•m
Maximum resistivity = 100000.0 Ohm•m
Number of elements combined horizontally = 1
Number of elements combined vertically = 1
Vertical / Horizontal roughness ratio = 0.5
Estimated noise of resistivity data = 3%
Initial damping factor of resistivity = 10
Starting iteration of quasi Newton method = 21
IP inversion method: No IP Inversion

APPENDIX B

Error Analysis

The raw apparent resistivity data from the ERI survey conducted in 2008 shows evidence of at least two bad electrodes (Figure 18a). Bad electrodes are easily identified in the raw data by diagonal trends of anomalous data (see Figure 5). The inversion software is capable of removing erroneous data points and does so by deleting the bad image point. Even though these bad image points were removed, the inverted resistivity cross section from the 2008 data line b-b' (Figure 18, b), shows feature that resemble the ones created by the bad electrodes.

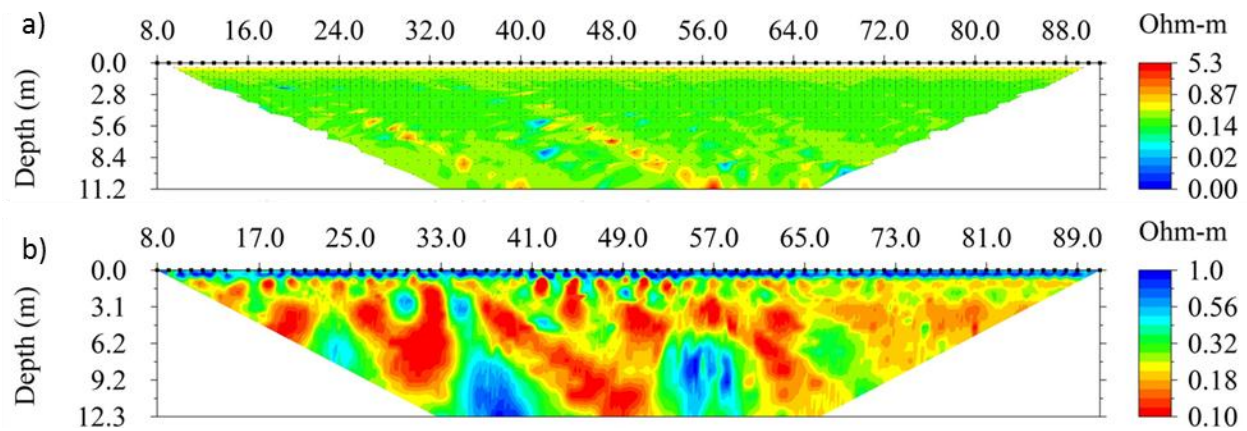


Figure 18. Pseudosection of measured apparent resistivity (a) and Inverted resistivity cross section (b) for the 2008 data line b-b'.

To test if deleting multiple image points from the apparent resistivity pseudosection could still have an effect on the inverted data, a two layer model (figure 14a) was used and 2 bad electrodes were simulated using the data from the 2008 b-b' line. The modeled apparent resistivity pseudosection (Figure 19a) shows the same erroneous diagonal lines as the 2008 data (Figure 13a). The data were then inverted using the same method as the 2008 data (Figure 19b).

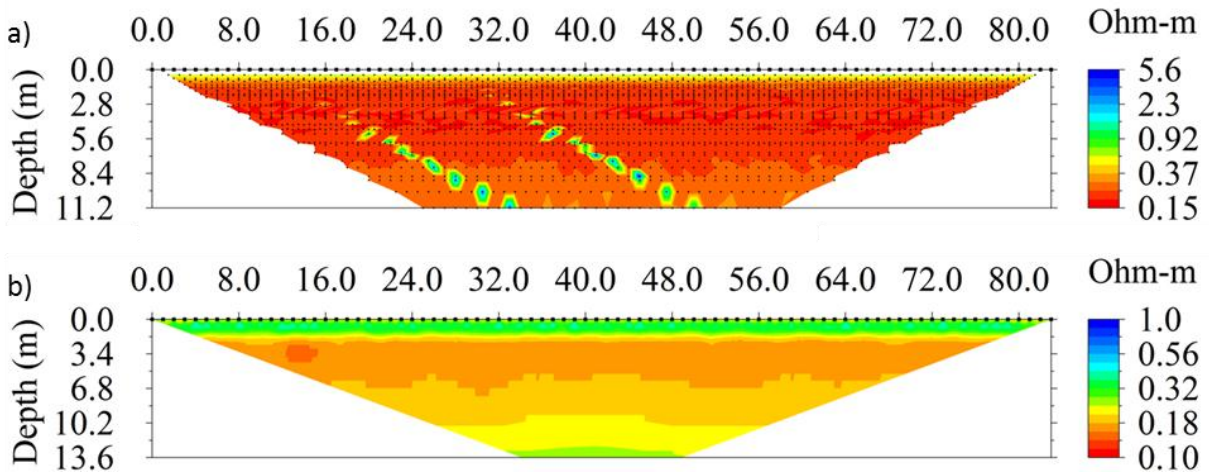


Figure 19. Pseudosection of calculated apparent resistivity (forward calculation) with 2 bad electrodes (a) and modeled inverted resistivity cross section (b).

Analysis of these results shows that there is indeed no effect caused by deleting multiple image points from bad electrodes. This confirms that the features present in the 2008 resistivity cross section of line b-b' are the result of free convection and not an artifact creating during inversion.

APPENDIX C

2010 Sabkha Stress Test Experiment

In the 2010 visit, an 11.5 by 12 m grid of 240 electrodes with 0.5 m electrode spacing and two 8-electrode borehole arrays with 0.25 m electrode spacing were installed at the field site to monitor an infiltration experiment using dipole-dipole configurations. Over the course of 15 days, ~6350 liters of saline water were extracted from the sabkha 110 m southwest of the study area. Over a 6-day period the sabkha water was then sprayed evenly over a 10 m diameter circle, centered in the study area (Figure 20), resulting in approximately 80 mm of infiltration. The electrical resistivity monitoring system was programmed to perform daily measurements of the study area and then upload them for processing using a wireless USB modem. In addition to ERT monitoring, six groundwater sampling wells were installed, four inside the infiltration area and two outside, at depths between 1.6 and 2.5 m. Groundwater samples were collected from the wells daily throughout the experiment and then periodically over the next three months.

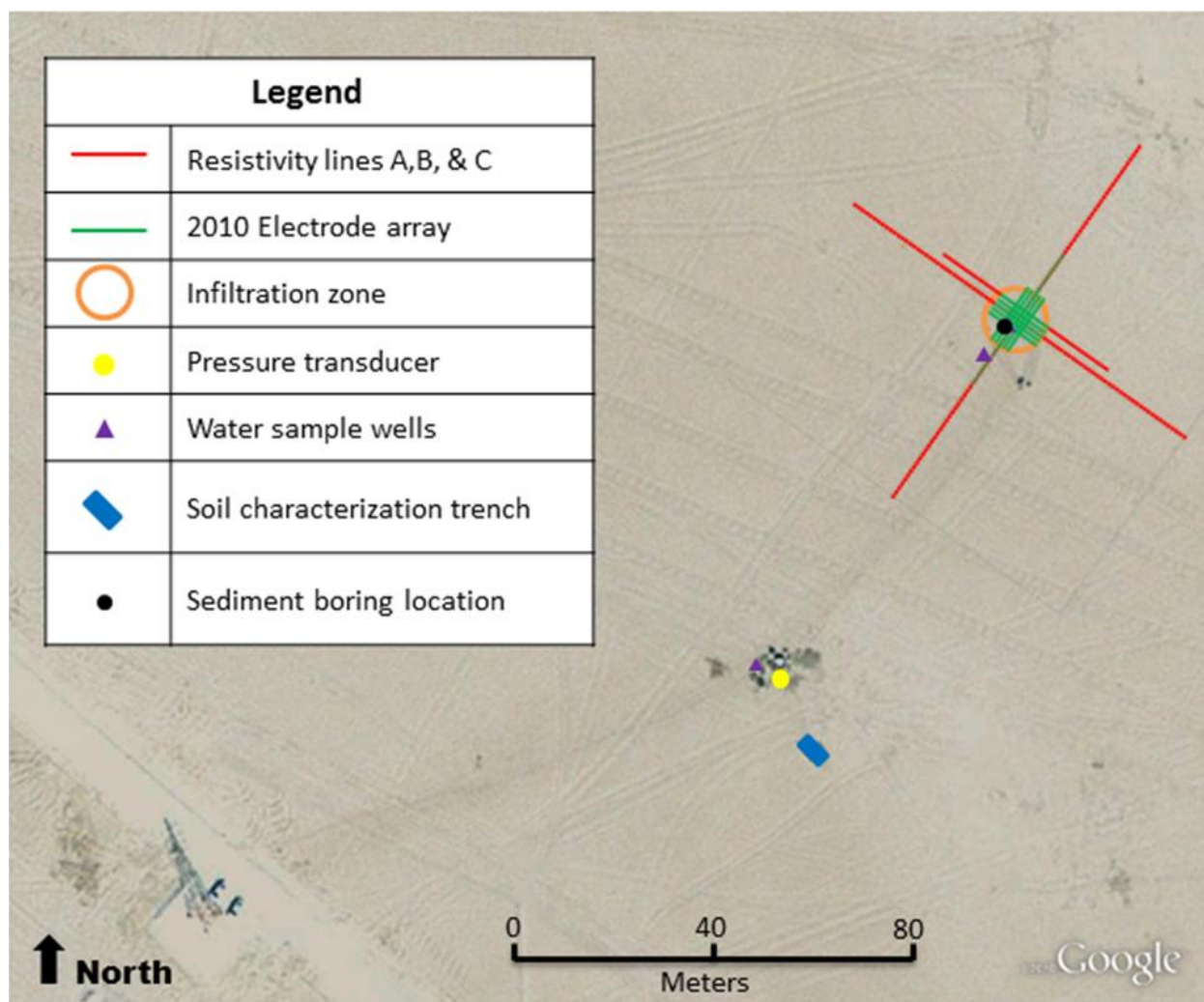


Figure 20. Map of the study area showing the 2010 electrode arrays, infiltration zone, and pressure transducer (background image from Google Maps taken 11/7/2010.)

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