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Highlights

- Helicopter TEM (HTEM) used to delineate fresh and saline water in the uppermost ~200 m of the aquifer of the Okavango Delta
- Widespread freshwater infiltration from the surface into the background saline aquifer
- Salt fingers from saline islands detected in freshwater regions
- Groundwater flow models made to help interpret the HTEM inversion results

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Using helicopter TEM to delineate fresh water and salt water zones in the aquifer beneath the Okavango Delta, Botswana

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ABSTRACT

The Okavango Delta is a vast wetland wilderness in the middle of the Kalahari Desert of Botswana. It is a largely closed hydrological system with most water leaving the delta by evapotranspiration. In spite of this, the channels and swamps of the delta remain surprisingly low in salinity. To help understand the hydrological processes at work, we reanalyzed a previous inversion of data collected from a helicopter transient electromagnetic (HTEM) survey of the entire delta and performed an inversion of a high resolution dataset recorded during the same survey. Our results show widespread infiltration of fresh water to as much as ~200 m depth into the regional saline aquifer. Beneath the western delta, freshwater infiltration extends to only about 80 m depth. Hydrological modeling with SEAWAT confirms that this may be due to rebound of the regional saltwater-freshwater interface following the cessation of surface flooding over this part of the delta in the 1880s. Our resistivity models also provide evidence for active and inactive saltwater fingers to as much as ~100 m beneath islands. These results demonstrate the great extent of freshwater infiltration across the delta and also show that all vegetated areas along the delta's channels and swamps are potential locations for transferring solutes from surface water to an aquifer at depth.

KEYWORDS

freshwater infiltration; saltwater infiltration; saltwater fingering; hydrogeophysics; TEM

1. INTRODUCTION

In the middle of the semi-arid sparsely populated interior basin of southern Africa, the Kalahari Desert, lies a sprawling perennial wetland – the Okavango Delta. It is a vast wilderness area of great importance for both vegetation and wildlife, including the typical African megafauna [1]. The relatively small local population of about 2,500 people [2], living almost entirely along the delta's western and southern edges, is dependent on the water of the delta for its livelihood, which is based largely on livestock grazing and tourism. All of this is made possible by the startling fact that the surface water of the Okavango Delta remains fresh in spite of only 1.5% of the delta's inflow exiting as outflow and evapotranspiration otherwise being the dominant mechanism of water removal [3], which would usually lead to a water body becoming progressively salty, such as the Caspian Sea, Dead Sea or Great Salt Lake. In contrast to the essentially endorheic system of the Okavango Delta, the only other inland delta of this scale, the Inner Niger Delta of Mali, experiences outflow equal to about half of its inflow [4], which removes dissolved salt from the delta.

In order to help understand this balance between fresh and saline water in the Okavango, various studies have dealt with the delta's flooding dynamics [5, 6] (and references contained within), whereas others have investigated the concentration of hypersaline water at the delta's ubiquitous salt islands [7-10]. Although the flow of fresh water and the accumulation of solutes at salt islands are well observed at the surface, very little is known about the distribution and movement of fresh and saline water within the underlying aquifers.

To investigate this as well as the hydrogeology of the Okavango Delta in general, a helicopter transient electromagnetic (HTEM) survey of the entire delta was commissioned by the Botswana government in 2007. TEM in general is an ideal non-intrusive method of investigating electrically conductive near surface materials [11]. Although traditionally used for mineral exploration with relatively easily identifiable high-contrast targets (e.g. ore bodies), HTEM has been applied more and more for surveys with lower-contrast targets such as hydrogeological mapping [12-17]. HTEM has the

advantage over ground-based investigations of being able to uniformly and efficiently record soundings over extensive inaccessible areas [12], such as the swamps of the Okavango Delta. With the exception of an earlier airborne survey conducted along a strip at the distal end of the delta's fan [18], no other geophysical surveys have targeted the hydrogeology of any extensive area of the delta, let alone the entire delta itself. Of particular interest when planning the survey were the perennial floodplains and salt islands spread throughout the middle of the delta. For this reason, an additional high resolution area was recorded over a cluster of salt islands located within permanent floodplains. After providing background information on the hydrogeology of the Okavango Delta, the HTEM dataset and modeling techniques used, we first reexamine an inversion model of the regional HTEM dataset that was previously used to interpret the geological evolution of the delta [19]. Here we closely evaluate this model in terms of groundwater infiltration. We then present an inversion model of the high resolution HTEM dataset, which targets the subsurface hydrology of salt islands. We then employ groundwater flow modeling to create 2D profiles of solute concentrations under hypothetical conditions related to freshwater and saltwater groundwater infiltration. These are then utilized to help interpret the inversion results. Finally, we discuss the implications of our findings on the long-term balance and dynamics of fresh and saline water in the Okavango Delta.

2. THE OKAVANGO DELTA

The Okavango Delta (technically a megafan) of northern Botswana is composed of an upstream linear zone known as the panhandle (approximately $100 \text{ km} \times 15 \text{ km}$; Fig. 1), which sharply transitions to the downstream fan (approximately $150 \text{ km} \times 150 \text{ km}$; Fig. 1) composed of many distributary channels, which have a typical depth of 4-5 m [20], and perennial and seasonal swamps. The surface topography of the delta is very flat and the elevation drop over the $\sim 250 \text{ km}$ from the top of the panhandle to the bottom of the fan near the town of Maun amounts to only $\sim 60 \text{ m}$ [21, 22].

2.1. Geologic setting

The fan of the Okavango Delta occupies a SW-NE graben believed to be a southwestern extension of the East Africa Rift System (EARS) [23-25]. It is likely still active as evidenced by recent seismicity [26] and fault escarpments as high as 15 m along its southeastern edge. The graben is bounded on the NW by the Gumare fault, which exhibits about 50 m of normal displacement, and the Kunyere and

Fig. 1. Inset in lower left indicates the location of the Okavango Delta within Southern Africa. The Okavango River flows from the northwest out of Angola and Namibia and enters the Okavango Delta in Botswana at the panhandle, which conveys the water to the fan where it is then distributed among channels and swamps toward the city of Maun. The boundary faults of the graben occupied by the fan

are indicated as: G – Gumare fault, K – Kunyere fault, Th – Thamalakane fault. The latter two faults contain rivers by the same names. The Thaoge channel was the primary distributary channel in the fan until about 1883. Yellow borders delineate the boundaries of the regional and high resolution 1 (HR1) helicopter transient electromagnetic (HTEM) survey areas. The regional survey lines were flown in a SW-NE direction with the exception of two cross lines that are shown in Figure 6. Other features of interest include: BR – Boteti River, LN – Lake Ngami, OR – Okavango River, SS – Selinda spillway. Thamalakane faults in the SE, which together have about 300 m of normal displacement [19, 27, 28]. The NW-SE-oriented panhandle is likewise bounded by faults.

The delta is located within the middle of the Kalahari Basin as defined by the extent of Kalahari sediments [24]. This sediment group is characterized by up to 450 m of late Cretaceous to Recent unconsolidated sands, gravels, and clays that owe their existence to aeolian, fluvial, and lacustrine depositional processes and rest unconformably on Paleozoic or Precambrian basement [29]. In the vicinity of the Okavango Delta, the thickness of Kalahari sediments lies in the range of about 50 – 300 m [3, 30].

The pattern of sedimentation and drainage around the Kalahari Basin is partly the consequence of bulging and rifting associated with the break-up of Gondwana [31]. Subsequent planation and crustal uplift further sculpted the area [32]. In particular, uplift along the SW-NE Kalahari-Zimbabwe axis cut off the Okavango River from the Limpopo River to form the southern boundary of the basin [23, 33], which is proposed to have occurred in the late Paleogene [34]. The impoundment of the proto Okavango River led to the formation of Paleo Lake Makgadikgadi, which existed at various levels until as recently as about 8500 ka [35]. At some point, the paleo lake covered an area in excess of 90,000 km² (larger than any lake today) and included the present location of the Okavango Delta [19, 36]. Consequently, the Okavango Delta is underlain by intercalated silt/clay lacustrine sediments, which themselves are underlain by sands or gravels of a Paleo Okavango Megafan [19, 36]. Reduction of water flow into the basin due to climate change and/or river redirection [33, 37] heralded the end of Paleo Lake Makgadikgadi and led to the formation of the modern Okavango Delta (megafan). The

delta itself is composed of fluvial Kalahari sand and aeolian deposits interspersed with calcretes and silcretes precipitated from solutes in the vadose zone [5].

2.2. *Surface hydrology*

Although referred to as a desert, much of the Kalahari is actually semi-arid. The Okavango Delta receives 467 mm of precipitation annually ($6260 \text{ Mm}^3/\text{yr}$) [6], which occurs almost entirely during the southern summer months (Dec.-Mar.). Precipitation is much greater to the north in Angola (up to 1300 mm/yr), which gives rise to the Okavango River and its tributaries. This supplies an annual flood to the delta that peaks in the upper panhandle in April and constitutes the bulk of water influx to the delta ($8780 \text{ Mm}^3/\text{yr}$) [6]. Due to the low topographic gradient of the delta and its complex pattern of channels and swamps, this flood takes about four months to reach the distal end of the fan. At the peak of the flood, the 3300 km^2 of permanent wetlands are augmented with an additional $3300 - 10,400 \text{ km}^2$ of seasonal wetlands [6]. Depending on the amount of floodwater and preexisting groundwater storage, surface flow out of the delta may occur, which on average amounts to about 2% of the water influx ($260 \text{ Mm}^3/\text{yr}$) [6]. Such outflow occurs predominantly via the Thamalakane and Boteti Rivers toward the Makgadikgadi Pans or via the Kunyere River to Lake Ngami, although the Selinda spillway may also carry water to the east during exceptionally high floods (Fig. 1). Evapotranspiration is responsible for most of the remaining water outflux from the delta ($14,920 \text{ Mm}^3/\text{yr}$) [6].

Distributary flow across the fan takes place in several main channels that leak water into the adjacent swamps. The pattern of distributaries is in an on-going state of change as a consequence of the very low gradient of the fan [21, 22]. Channels are believed to aggrade with bedload and eventually become choked with vegetation, which leads to channel abandonment and redirection of water elsewhere [38]. However, it has also been suggested that seismic activity may play a role in the shifting of channels [39]. One of the most significant channel abandonments ever observed in the delta is that of the Thaoge channel (Fig. 1) located in the western side of the fan. From as late as the time of the first European visits to the region until about 1883, the Thaoge channel carried the bulk of water through the delta and filled Lake Ngami. By 1884 it had become blocked in the upper part of the fan, cutting

off flow to the lower part of the delta and Lake Ngami [39]. As a consequence, water flow shifted toward channels in the central and eastern fan, where most of the flow occurs today.

Essentially all surface flow into the delta is supplied by the Okavango River. It is very fresh (40 mg/l total dissolved solids (TDS)) and carries a sandy bedload derived from the Kalahari sands of its catchment area in Angola [40]. In spite of evapotranspiration removing about 95% of the water from the delta, water flowing out of the delta remains surprisingly fresh (~80 mg/l TDS) [3]. In total, only 5% of inflowing dissolved solids exit the delta by surface flow [29]. The remainder accumulates at islands as a consequence of transpiration of water by vegetation and capillary evaporation [6].

2.3. *Groundwater*

The Okavango Delta can be regarded as a freshwater anomaly superimposed on a regional saline groundwater environment. That is, less dense fresh water is mounded on top of denser saline water [3]. In fact, a rule of thumb of some safari lodge operators within the delta, where groundwater information is very sparse, is for drinking water wells to be drilled no deeper than 25 m, since a well of greater depth may yield saline water. At the drier distal end of the delta, where fresh water is generally confined to channels, the solute content of the uppermost aquifer appears to be more heterogeneous, as found by an extensive groundwater study around the city of Maun [41]. In addition to drilling some 70 wells, this study made use of the helicopter transient electromagnetic (HTEM) method in mapping aquifers [18], the same method used in our study. Additional groundwater wells and water samples exist along the thinly populated western edge of the delta.

Within the generally freshwater environment of the Okavango Delta, isolated pockets of highly saline water of around 20 g/l [5] may develop beneath islands scattered throughout the fan. As already mentioned, trees and other robust vegetation living on islands draw in and consume groundwater, leaving behind the dissolved solutes [7, 42]. The accumulation of salt causes islands to expand laterally and leads to a radial zonation of vegetation with more salt-tolerant species located toward the interior. The center of some islands is barren due to salt concentration being too high for any vegetation type. Such white “salt islands” ringed with vegetation are ubiquitous throughout the fan.

Furthermore, once the saline water beneath an island achieves a sufficient density, it may overcome the evapotranspiration-induced upward flow of fresh groundwater and sink to depth [8, 10].

Depending upon the strength of the horizontal hydraulic gradient, the sinking saline water may either form a vertical salt density finger or a laterally dispersed plume [9]. Indeed, a salt finger has been imaged beneath an island in the Okavango Delta using borehole-to-surface electrical resistance tomography (ERT) [10].

Lateral groundwater outflow from the Okavango Delta is unknown but is not considered significant on the basis of comparisons of dissolved isotopes in the shallow groundwater [3, 43].

3. HTEM DATA

3.1. TEM/HTEM overview

The transient electromagnetic (TEM) method measures the time-dependent response of the earth to an electromagnetic pulse induced by a varying current in a transmitter coil at the surface [11].

Immediately after the transmitter current has been turned off, a receiver coil measures the changing magnetic flux of induced eddy currents that diffuse downward and outward into the ground. These data are recorded at distinct times (up to 20 ms), which are known as time gates. The rate of change of the magnetic flux is dependent on the electrical resistivity of the materials in the ground as well as instrument-related factors. Depending on the magnetic moment of the transmitter coil and the resistivity of the ground, the depth of investigation of the TEM method can reach several hundred meters.

In general, crystalline rock is more resistive than sand, which is again more resistive than silt or clay. In unconsolidated sediments with low clay content, the resistivity of a formation is dominated by that of its pore fluid [44]. In this case, lower resistivity is indicative of more saline water and vice versa. Although the resistivity ranges of fresh and saline aquifers and sand and clay sediments can vary greatly [45], borehole investigations at the distal end of the delta near the city of Maun found the following relationship [41]:

- 20-200 Ωm : clean sands
- 10 – 20 Ωm : interlayered sand and clay
- <10 Ωm : clay or saline water in a clean sand matrix

The use of a helicopter to make measurements while towing the TEM transmitter and receiver coils (HTEM) allows for the rapid and efficient recording of many thousands of TEM soundings over an extensive area. This has the added advantage of being able to record over a large regular grid while avoiding the difficulties associated with accessing remote protected areas.

3.2. *HTEM survey of the Okavango Delta*

In 2007, the Botswana Department of Geological Survey commissioned an HTEM survey of the entire Okavango Delta to aid in understanding the hydrology of the delta. The survey pursued the goals of meeting the government's obligations to provide a database for international efforts in protecting the delta as well as identifying potential freshwater resources for communities along the delta's western edge. The HTEM method was very well suited for this task given the vastness and remoteness of the delta.

The survey included a regional area flown in a SW-NE direction with 2-km line spacing covering the entire delta and a much smaller, 5 km $\times$ 5 km high resolution area (HR1) flown in a N-S direction with 50-m line spacing centered over a cluster of salt islands in the eastern fan (Fig. 1). The survey benefited from very low electromagnetic noise as a consequence of the delta being almost entirely uninhabited by people. The commercial VTEM instrument was used, which has the advantage of a high transmitter moment (440,000 Am^2) allowing for great depth penetration. Podgorski et al. [46] provide details on the survey.

4. METHODS

4.1. *HTEM data processing and inversion*

Data processing and inversion were previously conducted for the regional area [19] and now for the first time for HR1. Prior to being able to invert the HTEM data for a reliable resistivity model of the

ground, a series of standard and novel processing steps were applied as described in Podgorski et al. [46]. The regional survey data had been inverted using a pseudo-2D laterally constrained inversion (LCI) scheme, whereby resistivities and layer depths of 1D models are constrained to those of adjacent models along a flight line [47]. For the much smaller HR1 area, we used a pseudo-3D spatially constrained inversion (SCI) scheme that employs resistivity and thickness constraints across as well as along lines [48]. The SCI approach had also been tested on the much larger regional data set but was not deemed worthwhile given only very minor improvements at a significantly greater computational cost. That is, the constraints provided by the 2-km line spacing of the regional survey (versus 50-m line spacing of HR1) had only a very minimal effect on models across lines. Furthermore, the regional inversion did not suffer from artificial lineations that had been initially apparent in the high resolution survey [46].

Analysis of a multi-layer inversion with fixed layer boundaries indicated that 4 layers would adequately capture the variability in resistivity throughout the survey area, although for much of the area only 3 layers would have sufficed. Selecting 4 layers therefore allowed for all important features to be depicted while also keeping the model relatively simple. Although not reported by the contractor flying the survey, a measurement error of 5% was assumed for the purpose of the inversions. The start models and root-mean-square (RMS) statistics of model fit to HTEM sounding data are listed in Table 1.

Table 1. Inversion details for the regional and HR1 survey areas.

Survey area	Regional	HR1
Layers	4	4
Start model depths (m)	30, 100, 200	25, 70, 200
Start model resistivities (Ωm)	50, 40, 50, 40	50, 50, 50, 50
% RMS	1.5 ± 1.5	1.3 ± 0.4

4.2. Groundwater flow modeling

The program SEAWAT was used to simulate variable density groundwater flow [49] to compare with and help interpret the inversion models of HTEM data. A forward Euler finite difference method with the generalized conjugate gradient solver was used to simultaneously solve the equation for variable density groundwater flow in porous media:

$$-\nabla \cdot (\rho \vec{q}) + \bar{\rho} q_s = \rho S_p + \frac{\partial \rho}{\partial t} + \theta \frac{\partial \rho}{\partial C} \frac{\partial C}{\partial t} \quad (1)$$

where:

ρ is fluid density,

$\vec{q}$ is specific discharge,

$\bar{\rho}$ is density of water from a source or sink,

q_s is volumetric flow rate per unit volume of aquifer from sources and sinks

S_p is specific storage,

P is pore fluid pressure,

C is solute concentration,

θ is porosity,

t is time,

and the solute transport equation [50]:

$$\frac{\partial C}{\partial t} = \nabla \cdot (D \cdot \nabla C) - \nabla \cdot (\vec{v} C) - \frac{q_s}{\theta} C_s \quad (2)$$

where:

D is the hydrodynamic dispersion coefficient,

$\vec{v}$ is the fluid velocity,

C_s is the solute concentration of water from sources or sinks

The infiltration of freshwater was modeled with 60 mg/l TDS (average concentration of fresh water entering and exiting the delta) [3] in a 2D section by imposing a relative head of +1 m to a 200-m-wide strip representing a freshwater channel on top of the regional saline aquifer. This elevated head was adjoined by two 100-m-wide sections of -1 m head representing the upward pull of transpiration by lush flanking vegetation. This setup ensures that the positive head of filtrating fresh water is offset by an equivalent amount of evapotranspiration, which is consistent with the delta essentially being a closed system. Cells on the surface greater than 100 m away from the +1 m head of fresh water were set to inactive to represent a lack of interaction with the aquifer by the predominant sparse vegetation of the arid landscape further away from channels. The models were assigned a horizontal hydraulic conductivity of 4 m/day and four different conceivable concentrations of the background saline aquifer (120 mg/l, 2600 mg/l, 8000 mg/l, 16,000 mg/l). The uppermost 60 m of the models were assigned a vertical/horizontal hydraulic conductivity anisotropy ratio of 0.5 and a porosity of 0.3 to correspond to sandy sediments [10]. Four different combinations of anisotropy and effective porosity were used to represent varying clay content below 60 m [51], which is the upper extent of a continuous layer of sandy clay found in a borehole in the middle of the fan [52]. Each model was run for a simulation time of 1000 years to produce long-term steady-state solute transport conditions, which is a conservative estimate given that channels may shift position on a shorter time scale [38]. The parameters used in all SEAWAT models are listed in Table 2.

Table 2. List of model parameters used in SEAWAT for modeling the infiltration of fresh water, rebound of saline water, and salt fingering.

Parameter	Value (freshwater infiltration)	Value (saline aquifer rebound)	Value (salt fingering)	Unit
Domain size (length x/height z) (cells x/cells z)	3000/200 (300/20)	3000/200 (300/20)	800/100 (160/100)	(m) (cells)
Cell size (horizontal/vertical)	10/10	10/10	5/1	(m)
Lateral boundary conditions	n/a	Fixed head, fixed concentration	Fixed head, fixed	

			concentration	
Simulation time	1000	100	2000	(years)
Prescribed head at recharge/transpiration zone	+1 for 200 m in center of profile -1 for 100 m on each side	-1 for 400 m in center of profile	n/a	(m)
Horizontal K	4	4	2	(m/day)
Anisotropy ratio (K_z/K_x)	0.5 (0-60 m depth); 0.5, 0.05, 0.005, 0.0005 (>60 m depth)	0.5 (0-60 m depth); 0.5, 0.05, 0.005, 0.0005 (>60 m depth)	0.1	
Effective porosity	0.3 (0-60 m depth); 0.30, 0.23, 0.16, 0.10 (>60 m depth)	0.3 (0-60 m depth); 0.30, 0.23, 0.16, 0.10 (>60 m depth)	0.3	
Initial aquifer concentration	120, 2600, 8000, 16000	(values following freshwater infiltration)	60	(mg/l)
Vertical dispersivity	10	10	1	(m)
Horizontal dispersivity	1	1	0.1	(m)
Effective molecular diffusion coefficient	10^{-8}	10^{-8}	10^{-8}	(m ² /sec)
Courant number	1	1	0.1	
Peclet number	1	1	1	

Since elevated head from perennial flooding has not existed in the western fan since about 1884, the denser water of the background aquifer could here be rebounding and displacing previously infiltrated fresh water back toward the surface where it is transpired by vegetation. To model the effect of transpiration at an abandoned channel, a head of -1 m was applied for 100 years across the 400-m-wide active surface of the freshwater infiltration models. The 2D concentration distributions after 1000 years of freshwater infiltration (60 mg/l TDS) were used as initial conditions. The same four anisotropy combinations as for freshwater infiltration were used as were typical values for the molecular diffusion coefficient (10^{-8} m²/sec) and longitudinal dispersion (0.1 of the transport distance, or 10 m) [53]. The Courant number was set to 1, which then dictated the transport step size. The vertical Peclet number (the ratio between grid spacing and longitudinal dispersivity) equals one, which

is well within the Peclet criteria of ≤ 2 for avoiding problems with numerical stability and dispersion [54]. All model parameters are listed in Table 2.

SEAWAT was also employed to model the effect of transpiration-induced solute accumulation on an island (parameter list in Table 2). Evapotranspiration of 0.8 mm/year was first run continuously for 2000 years on a freshwater aquifer (60 mg/l) to model the effect of the accumulation of solutes in the near surface. To test the effect of the extreme case of the surface becoming devoid of all vegetation due to over-salinization, the same model was first run for 200 years, after which evapotranspiration was then turned off (see Results for details).

5. RESULTS

5.1. Inversion of HTEM datasets

The depth of investigation limit of the inversions is about 300 m, which is based on the greatest depth that the models are sensitive to, although this is quite variable. Although the noise level was uniformly low throughout the survey area, the resistivity encountered by soundings greatly influenced the depth to which a signal could penetrate. For example, an eddy current will propagate relatively slowly through a strongly conductive layer, thereby reducing or preventing any signal from underlying layers. This affected the resolution of resistivity of a deep 100-km-wide resistive layer [36]. However, this was not an issue for the present paper which focuses on aquifers at shallower depths.

Figure 2 contains resistivity depth slices of the regional survey area from the surface to 200 m depth. The uppermost 20 m are dominated by moderate resistivities in the range of 15 – 40 Ωm , which is indicative of the fresh water of the delta and/or saturated sand of the surrounding area. At greater depths, the fan becomes more conductive due to the presence of saline water and lacustrine clay layers from Paleo Lake Makgadikgadi [19, 36]. From 60 – 200 m depth, a layer of high resistivity with a convex outward edge expands with depth to the southeast across the upper fan. Most of this feature has been interpreted to be a Paleo Okavango Megafan composed of freshwater-bearing sediments [19, 36], the extent of which is indicated in Figures 2a and 2l. The remainder of this resistive feature to the

southwest was determined to be basement rock on the basis of seismic reflection and refraction experiments [19].

Figure 2. Resistivity model depth slices to 200 m depth of previously conducted inversions of the regional HTEM data set with a new color scale to highlight freshwater infiltration. (b) Surface resistivity ranges between 15 – 100 Ωm representing fresh water and dry sands. (c) Below 20 m depth, low resistivity (blue, $<4 \Omega\text{m}$) occurs outside of the fan, whereas the resistivity beneath the fan generally remains moderate (green/yellow, 10 – 40 Ωm) with increasing depth. (h) – (l) From 120 – 200 m depth, a more resistive feature ($>200 \Omega\text{m}$) spreads outward from the northwest of the fan. It has been shown to be composed of freshwater sediments and to represent a Paleo Okavango Megafan, the maximum extent of which is indicated in (a) and (l). (f) and (g) indicate the location of profiles shown in Figures 5 and 6 respectively. An area of fresh water (moderate resistivity) in the southeastern fan deflects southward between 140 – 200 m depth (i) – (l) over what appears to be a clay lens.

5.2. Freshwater infiltration

Areas of moderate resistivity of 10 – 40 Ωm on the surface of the fan resemble the pattern of channels in the delta. Close inspection indeed shows a strong correlation. In Figure 3, four strips of moderate resistivity within a low resistivity background at the fan's distal end coincide with four channels on the surface, indicating

Figure 3. Evidence of freshwater infiltration from channels. (a) Within the 30 m resistivity depth slice, moderate resistivity (10 – 40 Ωm , green/yellow) at channels at the distal end of the fan (b) contrasts sharply with low resistivity of the regional saline aquifer (<4 Ωm , blue). (c) A vertical profile shows fresh water (moderate resistivity) present as deep as ~90 m beneath the water channels, whereas the area of moderate resistivity between channels is confined to the uppermost 10 m.

that the areas of moderate resistivity are due to fresh water infiltrated from the channels. Water samples from about 45 boreholes along the western edge of the fan provide confirmation [55].

Samples taken from 89 – 92 m depth in five boreholes in close proximity to each other highlight the relationship between resistivity and salinity (Fig. 4). Three of these samples come from a pocket of moderate resistivity (14 – 17 Ωm) that is contiguous with the dendritic pattern of moderate resistivity

at the surface. The other two samples come from the surrounding area of low resistivity ($\sim 2 \Omega\text{m}$). The solute concentrations of the three samples from the moderately resistive area are between 350 – 911 mg/l TDS, which is fresh water. Conversely, the solute concentrations of the two samples from the conductive area are 1820 mg/l and 2000 mg/l TDS, which is brackish water. Water samples from the other boreholes show a similar relationship. Therefore, the dendritic pattern of moderate resistivity extending across the surface of the fan and becoming more featureless at depth indicates freshwater infiltration from channels and swamps.

Figure 4. Comparison of resistivity model and groundwater samples. (a) Moderate resistivity with a dendritic pattern (green/yellow) contrasts with low resistivity (blue) in the 50 m depth slice. (b) Water samples from five boreholes spanning one of these resistivity boundaries in the southwestern corner of the fan. Samples from the three boreholes in the northeast of this area have solute concentrations of

350 – 911 mg/L TDS at 82 – 88 m depth and correspond to a pocket of moderate resistivity at 80 m depth (c). Saline water samples from the two boreholes in the southwest of this area (1820 – 2000 mg/L at 95 m depth) come from the broad area of low resistivity ($< 4 \Omega\text{m}$, blue) around the edges of the fan (d).

Freshwater infiltration extends from the surface to a maximum of ~ 200 m depth (15 – 40 Ωm areas in Fig. 2). It is more or less uniform across the channels and swamps of the fan from the surface to 40 m depth. At greater depths, fresh water is absent in the west, whereas it exists to as much as 200 m depth in the east. The displacement of the underlying saline aquifer by infiltrating fresh water can be explained in terms of elevated freshwater head, that is, fresh water in channels and swamps that is mounded on top of the regional saline aquifer.

Two resistivity profiles across the lower part of the fan that show significant lateral variability in the thickness of the moderately resistivity surface layer match well with water inundation as depicted by darker colors in the satellite image of the surface (Fig. 5; profile locations in Fig. 2f). Particularly striking is the thin surface layer of moderate resistivity in the far southwest and northeast of the profiles outside of the active fan, which contrasts with much thicker moderate surface resistivity within the active fan (middle of profiles). This appears to extend with decreasing resistivity to about 200 m depth between about km-75 and km-150 along the profile in Figure 5a. Further along between km-150 and km-175, freshwater infiltration appears to be abruptly stopped at ~ 150 m depth, which could be due to an impenetrable clay layer. Figure 5b extends the profile of Figure 3c laterally with the degree of freshwater infiltration being greater to the northeast where more surface flow occurs.

Figure 6 contains two vertical resistivity profiles along the general direction of surface water flow (and increasing sediment thickness) as indicated in Figure 2g. These show the uppermost 10 – 40-m-thick layer of moderately resistive freshwater sands overlying a more conductive 20 – 80-m-thick layer that is underlain by either basement rock or freshwater sediments of the paleo megafan. The resistivity of the intermediate conductive layer is about 10 Ωm higher in the northeastern profile (Fig. 5a) than in the southwestern one (Fig. 5b). This again appears to be due to freshwater infiltration as evidenced by the darker colors of the satellite image (surface inundation) along each line. Note the downward bulge

of the moderately resistive surface layer at the distal end of the delta in Figure 5b. This corresponds to the point where the profile crosses a river channel.

Figure 5. Inversion models of the two SW-NE lines crossing the fan of the Okavango Delta as indicated in Figure 2g. Along-line strips from the satellite image in Fig. 2a are also shown, whereby dark and light colors represent inundated and dry regions, respectively. (a) – (b) In both Line 3 and Line 4, there is a strong correlation between surface water (darker color in satellite image) and a thicker surface layer of moderate resistivity. In the middle of both profiles, the resistivity of intermediate-depth layers is higher (6 – 20 Ωm) than that of adjacent sections ($<4 \Omega\text{m}$) due to the increased presence of perennial water.

Figure 6. Inversion models of the two NW-SE lines crossing the fan of the Okavango Delta as indicated in Figure 2f. Along-line strips from the satellite image in Figure 2a are also shown, whereby

dark and light colors represent inundated and dry regions, respectively. (a) Line 3 passes close to the middle of the fan and shows freshwater sands at the surface underlain by a layer of lower resistivity representing the infiltration of fresh water into a clay-rich environment, which is then underlain by the Paleo Okavango Megafan and basement rock. Dashes indicate the uncertain boundary of the paleo megafan. (b) Line 4 lies within the western fan. Note the correlation of light/dark areas on the satellite image with thinner/thicker surface layers of moderate resistivity.

Groundwater flow modeling of freshwater infiltration produced 16 different models (Fig. 7) that were compared with the patterns of freshwater infiltration as depicted by the resistivity cross-sections (Fig. 5). This allowed for an assessment of what combination of salinity/clay content may be present. These models show that the vertical extent of freshwater infiltration is inversely proportional to both clay content (lower vertical hydraulic conductivity) and the solute concentration of the background aquifer. Lower vertical hydraulic conductivity tends to cause the freshwater plume to spread horizontally, whereas higher background solute concentration makes the leading edge of the plume become narrower. Tests showed that varying the effective porosity in the range of 0.1 – 0.3 affected the extent of the freshwater infiltration at a given time by ~5% but did not affect the shape of the infiltration pattern.

Figure 7. Modeling of freshwater infiltration with SEAWAT to distinguish among possible aquifer conditions. Both the solute concentration (c) of the background aquifer and vertical to horizontal hydraulic conductivity (K_z/K_x) and porosity (n) were varied below 60 m, the latter to represent varying clay content. A surface head of +1 m was applied to a 200 m-wide strip in the center of the model representing freshwater infiltration (60 mg/l TDS) at a channel. This was flanked by a head of -1 m along 100 m to each side, which corresponds to an equivalent removal of water by evapotranspiration. Cells at the surface greater than 100 m away were set to inactive (grey) to represent a lack of interaction with the aquifer by the sparse vegetation found away from channels. The different parameter combinations were run for 1000 years.

Figure 8 shows the 16 models of freshwater rebound using only negative surface head representing the rebounding of fresh water. These models show a greater decrease in freshwater plume size with higher solute concentration of the background aquifer. Clay content appears to have little effect on the response of a net negative head on an existing freshwater plume.

Figure 8. Modeling with SEAWAT of the transformation to a freshwater plume once the surface channel is abandoned. Concentrations from the models of Figure 7 were used as initial conditions, and the models were run for 100 years with a head of -1 m to represent evapotranspiration along the entire active width of 400 m. The grey-colored cells along the surface are inactive.

5.3. Saltwater infiltration

Embedded within the broad area of moderately resistive fresh water at the surface are small pockets of low resistivity, particularly beneath islands. Figure 9 shows resistivity depth slices from a part of the HR1 area that contains several white salt islands of about 500 – 1000 m in size. From the surface to 20 m depth, there is a low resistivity anomaly of about 8 Ω m beneath three salt islands in the north of the area. At 40 – 100 m depth, much weaker low resistivity anomalies are found beneath each island. At the island indicated in the south of the area, the situation is reversed. Resistivity beneath this island from the surface to 20 m depth is consistent with that of the surrounding area, whereas a low-

resistivity anomaly occurs at 40 – 100 m depth. Both cases may be examples of saltwater fingering from the surface.

Figure 9. Resistivity depth slices in an area of salt islands within high resolution area 1 (HR1). (a) Selected salt islands are indicated with x's. (b) – (c) Low resistivity anomalies from the surface to 20 m depth are associated with the three marked islands in the north and likely represent saltwater fingering. (d) – (g) Between 40 – 100 m depth, low resistivity anomalies are no longer present beneath the three northern islands, although a low resistivity anomaly appears beneath the southern island.

Another example of a low resistivity anomaly and possible sinking of saline water beneath an island is shown in Figures 10 and 11. In this case, the island is much larger (several kilometers wide) and, although not as obviously “salty” or white as the islands in the HR1 area, still exhibits a marked decrease in vegetation toward the middle of the island. It also contains a significant low resistivity anomaly. The resistivity of the island from the surface to 20 m depth is considerably lower than that of the surrounding area, which contains fresh water and/or sand. At 40 – 60 m depth, the anomaly at the island becomes less resistive and appears to connect with the larger regional low resistivity zone to the east. The low resistivity anomaly beneath the island weakens at 80 m and is no longer present at 100 m. This resistivity pattern is also visible in four profiles across the island (Fig. 11).

Figure 10. Low resistivity anomaly beneath a large island in the eastern fan (a) – (b). The SW-NE-oriented rectangle is the area shown in Figure 10. (c) – (d): The island is less resistive than its surroundings from the surface to 20 m depth. (e) – (g): At 40 – 80 m depth, resistivity beneath the island further decreases and appears connected with the low resistivity of the regional saline aquifer to the east ($<4 \Omega\text{m}$, blue). (g) – (h): Between 80 – 100 m, the low resistivity anomaly beneath the island has similar values to those of the infiltrated fresh water that is dominant in the area.

Figure 11. Vertical resistivity profiles across an island in the eastern fan as shown in Figure 9. The extent of the island is indicated by black bars on the satellite image and profiles. (a) The low resistivity anomaly is confined to the surface across the northwestern end of the island. (b) – (c) In the midsection of the island, low resistivity at the surface is concentrated along the edges of the island and connects with a broad continuous zone of low resistivity extending to ~100 m depth. This appears to be due to the sinking of concentrated salt. (d) The low resistivity anomaly at ~40 – 80 m depth extends to the northeast away from the island, which appears to be connected with the regional saline aquifer as in Figure 10e – g.

In the SEAWAT simulations of the accumulation and sinking of saline water, it is seen that a significant concentration of salt accumulates in the middle of the island by 200 years (Fig. 12a) and overcomes the force of upward flowing groundwater to form a salt finger by 300 years (Fig. 12b). From 500 to 2000 years (Fig. 12c – e), salt continues to accumulate at the surface and is transported to depth via the salt finger. Vegetation can no longer grow (and transpire water) at toxic salinity levels such as those typically occurring in the middle of salt islands in the Okavango Delta. To test the effect on the development of a salt finger in the extreme case of a complete stop of evapotranspiration, the same model was run with evapotranspiration being turned off after 200 years (Fig. 12f), which is the time at which solute concentrations have considerably increased but not yet formed a salt finger. By

300 years, a more dispersed salt finger still forms, and between 500 and 2000 years (Fig. 12h – j) solutes continue to be transported from the surface to depth.

Figure 12. SEAWAT models of the development of a salt finger by evapotranspiration (ET) of 0.8 mm/day in a fresh groundwater environment of 60 mg/l. Two scenarios were run: (a) – (e) evapotranspiration allowed to run continuously for 2000 years, and (f) – (j) evapotranspiration initially run for 200 years and then turned off for the following 1800 years to represent the extreme case of the accumulation of salt at the surface inhibiting the growth (and associated transpiration) of all plant life.

6. DISCUSSION

The correspondence of subsurface moderate resistivity in the inversion model (15 – 40 Ω m) with locations of surface channels (Figs. 2, 3, 5, and 6) and of moderate and low resistivity with fresh and saline groundwater samples, respectively (Fig. 4), clearly indicates that the Okavango Delta produces a freshwater anomaly penetrating to various depths in a generally saline groundwater environment. It

is noteworthy that fresh water has infiltrated down to as much as 200 m in the eastern fan, as indicated by the inversion (Figs. 2l and 5a).

Groundwater flow models of freshwater infiltration into a background aquifer with varying salinity and clay content (Fig. 7) has allowed for a general assessment of these parameters in the regional aquifer in the Okavango Delta by comparing these models with an inversion profile from the lower fan (Fig. 5b). This profile shows plumes of freshwater infiltration that decrease in width with depth. The freshwater infiltration flow models that come close to this pattern are those with a background solute concentration of 8,000-16,000 mg/l and a ratio of vertical to horizontal hydraulic conductivity between 0.05 and 0.005 (below 60 m depth). Lower solute concentrations (≤ 2600 mg/l) and/or lower hydraulic conductivity anisotropy (0.5) result in freshwater infiltration that does not taper with depth. Lower solute concentrations and a greater anisotropy ($K_z/K_x < 0.5$) produce a plume that spreads laterally with depth. Finally, a combination of high solute concentrations (≥ 8000 mg/l) and high anisotropy (0.0005) almost entirely impedes the downward flow of fresh water, which is generally not observed at shallow depths in the fan. A notable exception to this is a ~20-km-wide strongly conductive area indicated as a clay lens in Figures 2l and 5a that sharply truncates well-defined freshwater infiltration from above.

Overall, the model with a background concentration of 8000 mg/l and K_z/K_a ratio of 0.05 below 60 m depth best resembles the freshwater plumes of Figure 5b. Using the estimated effective porosity corresponding to this level of clay content (0.23), the total volume of fresh water from the surface to 200 m depth based on locations of moderate resistivity is estimated to be $\sim 475,000 \text{ Mm}^3$, which is roughly 30 times the annual freshwater influx to the delta. When considering the range of effective porosity modeled (0.1 – 0.3), the estimates of fresh water from the surface to 200 m depth vary from $\sim 207,000$ to $620,000 \text{ Mm}^3$.

Because the resistivities of the 4-layer inversion models ultimately represent averages of each layer and the inversion method is inherently non-unique [56], these resistivity values need to be treated with caution, in particular with regard to the resistivity of intermediate layers (e.g., ~50 – 150 m depth in Figures 5 and 6). The reason for this is that dozens of borehole logs from just outside the survey area

[41], as well as one inside [52], indicate the presence of numerous intercalated clay layers with varying amounts of clay within this depth range. Since the inversion cannot distinguish among many relatively thin layers, an average for the entire section ultimately results. That is, the actual resistivity of a single layer in the model is not necessarily homogeneous but rather may have a high degree of small-scale variation as would be expected from the presence of many interbedded sand and clay layers. The infiltration of fresh water at least into the top of such a model layer would then have the effect of raising the overall resistivity of the layer (e.g., moderate resistivity of the middle sections of Figs. 5a – b and 6a). At the same time, the net hydrological effect of this package of clay layers must be such as to prevent infiltration of saline water into the underlying proposed Paleo Okavango Megafan. This is a reasonable hypothesis given the understanding that the clay layers were deposited during successive stages of Paleo Lake Makgadikgadi, whereby the lower part of this section would have been more frequently inundated by the lake and therefore contain greater amounts of lacustrine clay.

Another reason for the lower resistivity of the infiltrated fresh water is interaction with the surrounding saline water. The water at the surface is generally fresher than that at depth. For example, the solute content of the freshwater samples from the southwestern fan (Fig. 4) is considerably higher (350 – 911 mg/l) than that of the surface water (~60 mg/l). Mechanical dispersion (but not molecular diffusion) has a noticeable effect in the models, such that the elevated solute concentration of infiltrated fresh water can be attributed to mixing primarily by advection and dispersion.

That the depth to which fresh water infiltrates is significantly less in the western fan than in the eastern fan is best explained by the abandonment of the western Thaoge channel and adjacent swamps in 1884. Following this removal of surface freshwater head, the surrounding denser saline water appears to have been rebounding to a new pressure equilibrium between saline and fresh layers of water due to the gradual consumption of the latter by deeply rooted vegetation. Hydrological modeling (Fig. 8) confirms that such a rebound of saline water and retraction of fresh water are to be expected under these conditions.

Superimposed on the freshwater infiltration is saltwater infiltration on a much smaller scale from dry land, particularly from islands within the swamps. This is seen clearly beneath the salt islands in the HR1 area (Fig. 9), where salt fingers have developed down to 20 m as represented by zones of low resistivity beneath the area's northern islands. The low resistivity anomaly beneath the southern island in Figure 9 may be an inactive salt finger that had previously transported solutes to depth and subsequently allowed fresh water to reoccupy the shallowest subsurface. Due to lateral spreading of the TEM signal with depth, the resistivity and width of a salt finger will tend to be overestimated. That is, if anything a salt finger may actually be somewhat narrower and more conductive than it appears in an inversion profile. Furthermore, the depths of salt fingers as imaged with these constrained 1D inversions can be regarded as conservative estimates. Hydrological modeling confirms that highly concentrated saline water at the surface can sink to depth in situations where evapotranspiration is still taking place (Fig. 12a-e) as well as after evapotranspiration and associated solute accumulation may have ceased (Fig. 12f-j). Although this is a simplified model that does not consider the precipitation of solutes at the surface, it shows that a salt finger can still form once a high concentration of solutes has developed beneath an island, even if the supply of solutes via evapotranspiration were to cease. In any case, both scenarios indicate that an area of low resistivity (high solute concentration) extending beneath an island and surrounded by higher resistivity (e.g. fresh water) has likely formed as a salt density finger.

A similar situation may be the case at the significantly larger island in Figures 10 and 11. In particular, surface resistivity is higher beneath the center of the island than along the edges, although a low resistivity zone does exist at 40 – 80 m depth below the middle of the island. As with the southern island in HR1, this may represent saline water that had previously accumulated at the surface and sunk to depth. This is plausible given the greater size and presumably age of the island, whereby the island would have existed for a sufficiently long period of time for this event to have occurred (i.e., hundreds of years). In any case, the edges of the island clearly contain zones of low resistivity that connect at ~40 m depth with the low resistivity area beneath the center of the island. Close inspection of satellite imagery shows highly concentrated vegetation along a narrow strip at the island's edge and sparse vegetation with barren patches in the middle, which is consistent with the zonal vegetation of the

archetypal Okavango Delta island. It is also noteworthy that the saline water beneath this island is linked with the regional saline aquifer to the southeast at 60 – 80 m depth. This would therefore represent a case of island-derived saline water feeding into the regional aquifer, and it may indicate the presence of a regional hydraulic gradient toward the northeast.

Once transported to depth, solutes may mix into the surrounding fresh water and increase its salinity (and conductivity). In addition to this, a significant amount of solutes remains as precipitates on the surface causing a net mass flux to islands and creating salt flats which grow laterally as vegetation migrates outward. A summary of the basic processes of freshwater infiltration from channels/swamps and the concentration of saltwater and infiltration at islands is depicted in Figure 13.

Figure 13. Sketch of the two basic processes of surface water infiltration in the Okavango Delta.

Perennial fresh water in the main channels and swamps produces a pressure head on the regional saline aquifer allowing fresh water to infiltrate and displace the underlying denser water. Vegetation along the water's edge consumes some of this infiltrated water by transpiration to the atmosphere.

Within the zone of freshwater infiltration and evapotranspiration, concentrated saline water in the center of islands can achieve a sufficient density to overcome the upward flow force of evapotranspirative flux and sink as a salt finger to either connect with the regional saline aquifer or disperse within the infiltrated fresh water.

7. CONCLUSION

Inversion of a high-quality HTEM data set collected over the Okavango Delta has shed new light on its subsurface hydrology. This is the first time that a comprehensive image of the distribution of groundwater salinity has been obtained across the entire delta. This was made possible firstly by the ability of the HTEM method to record a large uniformly-spaced data set over this otherwise remote inaccessible area. Particularly advantageous to this survey was the very low electromagnetic noise of the region, which allowed for a very high signal-to-noise ratio in the TEM recordings with correspondingly exceptional resolution. Up to now, information sources on the hydrogeology of the delta have come only from boreholes concentrated along the delta's western and southern edges or from very sparsely scattered geophysical surveys within the delta. These data could only hint at the groundwater salinity and active processes that this HTEM survey has uncovered.

Laterally and spatially constrained inversions of the HTEM data have produced electrical resistivity models of the regional and HR1 survey areas to a maximum of ~200 m depth. Comparison with surface inundation and existing groundwater samples as well as the use of hydrological modeling helped clarify the origin of the resistivity values (i.e., lithology or pore fluid), which otherwise possess a degree of ambiguity. Toward the surface, most of the resistivity variability in the 1 – 40 Ωm range is due to differences in salinity of the groundwater, although an overall decrease in resistivity observed below about 20 – 50 m depth is also due to increased clay content.

A significant discovery of this study is the extent to which fresh water infiltrates the subsurface. Our inversions show fresh water penetrating to as much as 200 m depth, although this depth is uncertain given the limited number of model layers and increase of clay content with depth. More infiltration is present beneath the eastern fan than beneath the western fan. This appears to be the consequence of the cessation of flow in the Thaoge channel of the western fan in 1884, which led to a shift of regular flooding to the eastern fan. It highlights the rebound of the regional saltwater-freshwater interface as fresh water is transpired by vegetation.

Salt density fingering was detected through low resistivity anomalies in the inversions, thereby providing further evidence that this process is active throughout the Okavango Delta. The examples shown include a cluster of typical small islands covered with thick salt deposits as well as a much larger, sparsely vegetated island also exhibiting a characteristic ring of dense vegetation along its edge. Some of these low-resistivity anomalies extend from the surface to about 40 m depth, whereas others are disconnected from the surface. The latter case appears to represent salt density fingers that have detached from the surface after surface salt accumulation ceased or was reduced by decreased transpiration. In one of the examples, infiltrated saline water appears to connect with the regional saline aquifer.

These results demonstrate that all islands and dry areas along the delta's edge have the potential to remove salt from the surface by salt density fingering due to high solute concentrations caused by vegetation. Dry areas that do not appear particularly saline may have been locations of previously active salt density fingering. To quantify the impact of density fingering in removing solutes from the surface, an estimate of the rates at which solutes accumulate and sink to depth would be needed. These could be multiplied by the area of dry land within and bordering the delta in which vegetation accesses infiltrated fresh water. Adjustments would need to be made for seasonally flooded areas, where solute accumulation may proceed at a slower rate.

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