Submarine groundwater discharge and solute transport under a transgressive barrier island

Tyler B. Evans¹ and Alicia M. Wilson¹

1. Abstract

Many recent investigations of groundwater dynamics in beaches employed groundwater models that assumed isotropic, numerically-convenient hydrogeological conditions. Real beaches exhibit local variability with respect to stratigraphy, sediment grain size and associated topographic profile, so that groundwater flow may diverge significantly from idealized models. We used a combination of hydrogeologic field methods and a variable-density, saturated-unsaturated, transient groundwater flow model to investigate SGD and solute transport under Cabretta Beach, a small transgressive barrier island seaward of Sapelo Island, Georgia. We found that the inclusion of real beach heterogeneity drove important deviations from predictions based on theoretical beaches. Cabretta Beach sustained a stronger upper saline plume than predicted due to the presence of a buried silty mud layer beneath the surface. Infiltration of seawater was greater for neap tides than for spring tides due to variations in beach slope. The strength of the upper saline plume was greatest during spring tides, contrary to recent model predictions. The position and width of the upper saline plume was highly dynamic through the lunar cycle. Our results suggest that field measurements of salinity gradients may be useful for estimating rates of tidally and density driven recirculation through the beach. Finally, our results indicate that several important biogeochemical cycles recently studied at Cabretta Beach were heavily influenced by groundwater flow and associated solute transport.

2. Introduction

Surface water-groundwater interactions at the land-sea interface drive significant chemical reactions and contribute dissolved metals, carbon and nutrients to the coastal ocean (Bowen et al., 2007; Burnett et al., 2003, 2001; Charette and Sholkovitz, 2002; D’Elia et al., 1981; Johannes, 1980; Krest, 2000; Paytan et al., 2006; Whiting and Childers, 1989). A major portion of these surface water-groundwater interactions occur in sandy beaches (Bokuniewicz et al., 2004; Li et al., 1999), which occupy about 75% of the world’s ice-free coastlines (Brown and McLachlan, 2002). *Heiss and Michael* [2014] reviewed the four major driving forces for groundwater flow through beaches: (1) an inland hydraulic gradient and associated discharge of fresh groundwater (Glover, 1959; Kim and Hwang, 2002; Taniguchi and Iwakawa, 2004) seaward of a saline circulation cell (Boufadel, 2000; Michael et al., 2005; Robinson et al., 1998); (2) convective mixing due to density gradients between fresh and saline groundwater (Cooper, 1959; Groen, 2002; Kohout, 1960); (3) tidal pumping (Abarca et al., 2013; Li et al., 2000; Robinson et al., 2007a, 2007b; Sun, 1997; Vandenbohede and Lebbe, 2006); and (4) wave setup and swash (Bakhtyar et al., 2013; Heiss et al., 2014; Longuet-Higgins, 1983; Robinson et al., 2014; Sorensen, 2006; Xin et al., 2010; Fig. 1). These forcing factors for groundwater flow in the beach are non-linear in time and space (King, 2012; Xin et al., 2015, 2014). This combination of driving forces in permeable beach sediments leads to rapid flow ($0.1 - 10 \text{ m d}^{-1}$), which drives chemical transformations and exports dissolved constituents to the ocean (Huettel et al., 2014; Slomp and Van Cappellen, 2004).

Previous studies have reported a wide range of biogeochemical reactions in beach sediments. The mixing and chemical exchange that occurs in coastal aquifers, termed the subterranean estuary (Moore, 1998), is significant and has garnered much scientific interest.

Charette and Sholkovitz [2002] observed iron oxide coated sands in the subterranean estuary at the freshwater-saltwater interface, where reduced pore water containing dissolved Fe(II) was in contact with oxidized seawater. These iron oxide coatings were shown to prevent phosphorus from discharging to the coastal ocean through adsorption, exemplifying the importance of groundwater mixing at the freshwater-saltwater interface for geochemical processes. Ullman *et al.* [2003] showed that beaches serve as reservoirs for particulate matter and remineralized reactive organic matter from the coastal ocean, which drives higher rates of nutrient fluxes from the beach than could be sustained by upland discharge alone. Furthermore, fluxes of dissolved organic matter and nutrients from the subterranean estuary to the coastal ocean are sensitive to tides (Santos *et al.*, 2009). Schutte *et al.* [2015] hypothesized that a zone of rapid nitrogen cycling and subsequent nitrous oxide production in a beach in southern Georgia was controlled by periodic spring tide inundation. At the same field site, a confined aquifer system under the beach was hypothesized to facilitate methane export to the coastal ocean, but flow rates in this aquifer and the integrity of the confining unit were relatively unknown (Schutte *et al.*, 2016). The hydraulic feasibility of these proposed groundwater dynamics in the beach requires further investigation. These biogeochemical processes, like those in other subterranean estuaries, are likely transport-limited and may be sensitive to salinity, which indicates the need for a better understanding of groundwater flow below beaches.

Groundwater salinity is simple field measurement and can be used to provide insight into the flow dynamics beneath the beach. In some beaches, two separate saline circulation cells develop: (1) The classic freshwater-saltwater interface (FSI) (Cooper, 1959; Ghyben, 1889; Glover, 1959; Henry, 1959; Herzberg, 1901); and (2) the upper saline plume (USP) (Boufadel, 2000; Lebbe, 1999; Robinson *et al.*, 2006; Fig. 1). The FSI is present in all beaches and is the salinity

configuration that forms at equilibrium between flowing land-derived fresh groundwater and circulating saline groundwater (Cooper, 1959; Kohout, 1964). Density-driven recirculation (DDR) of seawater into the beach aquifer is driven by density gradients that form in the freshwater-saltwater transition zone, the width of which is controlled by dispersion. The USP develops through tidally driven recirculation (TDR) of seawater into the beach above a body of discharging fresh groundwater that exits near the elevation of low tide (Boufadel, 2000; Robinson et al., 2006). The shallow beach aquifer fills with seawater during rising tide and discharges during low tide after mixing with land-derived fresh groundwater. After the initial discovery and description of the USP by *Boufadel* [2000] and *Robinson et al.* [2006], the established convention was that most beach aquifers contained a USP. Recent studies have shown that the USP may be transient in coastal aquifers and its presence is highly dependent on lunar cycles, fresh groundwater head, beach slope and grain size (Abarca et al., 2013; Evans and Wilson, 2016; Heiss and Michael, 2014). The distribution of salinity in the beach is a direct consequence of the flow dynamics within the beach and can potentially be used to estimate rates of seawater recirculation (Evans and Wilson, 2016)

Evans and Wilson [2016] proposed a field measurement called the saline plume salinity gradient (SPSG) that was shown to vary with rates of TDR and DDR in theoretical studies. The SPSG is calculated by determining the concentration gradient between two points perpendicular to the shore, 1 meter below land surface. The first measurement is made at the approximate center of the USP, and the second is collected near the center of the freshwater tube directly seaward. These locations roughly correlate with mean high and mean low water. Stronger USPs had higher SPSGs. Beaches with the strongest USPs had the highest rates of DDR (Evans and Wilson, 2016). TDR was more complex and was dependent on beach slope, tidal amplitude and

inland hydraulic head (Robinson et al., 2007b). These hydraulic parameters can be simultaneously expressed in a nondimensional ratio of the width of the intertidal zone to tidal propagation distance (Li et al., 2000; Parlange et al., 1984; Robinson et al., 2007b):

$$\varepsilon = A \cot(\beta) \lambda \quad (1)$$

where A is tidal amplitude, β is beach slope and λ is the tidal propagation distance:

$$\lambda = \sqrt{\frac{n_e \omega}{2KH}} \quad (2)$$

where n_e is effective porosity, ω is tidal period, K is hydraulic conductivity and H is aquifer depth. This analytical index does not include the effects of the vadose zone on the tidal propagation distance in a beach. As described in *Kong et al.* [2015] the presence of an unsaturated zone above the unconfined beach aquifer has the potential to increase the tidal propagation distance. As a result, our calculations of ε are conservative estimates. In beaches with moderate slopes (0.01 to 0.1), TDR decreased with increasing values of ε (Evans and Wilson, 2016). More interestingly, SPSG, TDR and DDR are all related to ε , which can be calculated for any beach. *Evans and Wilson* [2016] hypothesized that rates of TDR and DDR could be estimated using only a measured SPSG and calculated value for ε . Additionally, these theoretical models predicted that SPSG would be higher for lower tidal amplitudes, indicating that neap tides should support weaker USPs than spring tides. Rates of seawater recirculation should be higher during spring tides (Evans and Wilson, 2016). Although these relationships are valid in theoretical beaches, the viability of using a measured SPSG to estimate rates of TDR and DDR in a real beach is unknown. Further study using numerical models calibrated to a real beach is required to test these hypotheses.

Most numerical studies of groundwater flow and solute transport processes in beaches assume homogenous, isotropic sand to depths as great as 30 m below land surface (Abarca et al.,

2013; Ataie-ashtiani et al., 1999; Boufadel, 2000; Evans and Wilson, 2016; Heiss and Michael, 2014; Lebbe, 1999; Michael et al., 2005; Robinson et al., 2007b, 2006; Vandenbohede and Lebbe, 2006). The use of homogenous models to describe groundwater flow in beaches ignores interbedded sediments present in a large portion of the world's beaches. In addition, many of these models used a constant single slope for the intertidal zone of the beach. Real beaches have variable topography due to the dynamic impacts of waves and currents. We hypothesize that the inclusion of real-world heterogeneity in beach groundwater flow models will highlight important deviations from these theoretical models and chose a transgressive barrier island beach as a site to test this hypothesis and the potential for estimating TDR and DDR from the SPSG.

We chose a transgressive barrier island beach as a test beach because barrier islands cover 49% of the coastline of passive continental margins and are characterized by complex, heterogeneous stratigraphy (Glaeser, 1978). These environments are extremely dynamic, and their morphology is controlled by wave action and tidal energy (Hayes, 1979). As a result, the beach profile across a barrier island is non-uniform. These beaches are also commonly underlain by low permeability, silty mud from ancient marsh platforms, which could drive anisotropy in groundwater flow. We hypothesized that inclusion of a non-uniform beach profile, subsurface heterogeneity and real tidal conditions would highlight significant deviations from theoretical conceptual models for beach groundwater flow and the USP. This hypothesis was tested using a combination of piezometer installation and hydraulic head monitoring, electrical resistivity tomography, and a two dimensional, variable-density, saturated-unsaturated, transient groundwater flow model.

3. Site Description

Cabretta Island is a small Holocene transgressive barrier island on the seaward side of Sapelo Island, Georgia (Fig. 2). The island experiences a semidiurnal tide with a range of ~2.5 m and average annual rainfall of 130 cm (Wilson et al., 2011). For the month of June 2012, mean sea-level was 0.24 m, MHW was 1.24 m and MLW was -0.93 m. The island is approximately 250 m in width by 2.1 km in length, and reaches a maximum elevation of ~3 m (MSL). The intertidal zone of the beach on the eastern side of the island has a slope of approximately 0.025. As described in *Wilson et al.* [2011], a transect extending from a tidal creek to the ocean was vibracored to depths of 3-4 m to determine the local stratigraphy in January 2008 (Fig. 3). On the marsh side of the island, a 1-2 m thick confining unit of marsh mud and silt was underlain by a permeable confined aquifer of fine sand. The forested upland, which was comprised of fine sand, made up the unconfined surficial aquifer. This aquifer was underlain by a low permeability confining unit made of interbedded silt and clay at approximately 2.5 m below land surface (Fig. 3). On the beach side, the sediments were fine sands with irregularly distributed mud and silt layers. Significant erosion of sand from the beach (~ 2 m) over this 4 year period was observed while surveying elevations in the summer of 2011. Additional vibracores were collected in the beach in February 2012. These cores revealed a confining unit made of interbedded silts and clays at approximately 3 m below land surface on the beach. Below this confining unit was a confined aquifer comprised of fine sand. We focus this study on flow and solute transport from the forested upland, through the beach, to the ocean.

3.1. Biogeochemistry and Hydrogeology of Cabretta Island

Several investigations at Cabretta Island have been published in the last 6 years, but none were specific to groundwater flow in the beach. *Anderson* [2010] developed a groundwater flow model for the Cabretta Island transect, but groundwater flow dynamics and solute transport in the

beach were simplified and the underlying low permeability mud layer was unknown at that time. *Wilson et al.* [2011] presented the effect of Tropical Storm Fay on groundwater dynamics at Cabretta Island but focused on the marsh side of the island. *Schutte et al.* [2015, 2016] focused on nitrogen and methane cycling on the beach side of Cabretta Island. They showed that both processes were largely groundwater transport-limited reactions. A simplified groundwater flow model was used to calculate tide-averaged flow velocities needed to calculate N₂O and methane fluxes from the beach. That model used a sinusoidal tide and annual average rainfall rate. The current study aims to quantify seawater recirculation and to investigate the configuration of the freshwater-saltwater in the beach. A more detailed numerical model was required to accomplish these goals.

4. Methods

4.1. Field Methods

Elevation surveying was conducted in early June 2012 to collect a beach topographic profile. Precipitation records were also obtained from the nearby Sapelo Island National Estuarine Research Reserve System (NERRS) weather station located at Marsh Landing. Tide water levels were obtained from the nearest NOAA tide gauge (Ft. Pulaski) from <http://tidesandcurrents.noaa.gov>. Ft. Pulaski is approximately 40 miles northeast of Sapelo Island and tidal records had a discrepancy of about -0.14 m and 27 min between our site at Cabretta and the tide gauge (Wilson et al., 2015).

As described in detail in *Wilson et al.* [2011], seven piezometer nests were installed along the transect in the summer of 2008 (Fig. 2). Piezometers were constructed using Schedule 40 PVC (3.2 cm) and Schedule 40 PVC slotted well screen (0.15 mm). The deeper wells had screened intervals of 30 cm and the shallow wells were 15 cm. Internal casings (2.5 cm) were installed in

each well to limit wellbore storage and exchange across the screen during the tidal cycles (Wilson et al., 2011). At nest TT4, two piezometers were installed that were screened at depths of approximately 2 and 4 m below land surface (Fig. 3). TT5 was located at the top of the dunes when installed, reaching approximately 5 m below land surface. At the top of the beach, nest TT6 is comprised of 3 piezometers screened at depths 1, 3 and 5 m below land surface. The final piezometer nest (TT7) was installed near the low tide mark on the beach, comprised of two piezometers screened at 1 and 3 m below land surface respectively (Fig. 3). Four years of sustained beach erosion (2008-2012) led to the loss of TT6-1, TT6-3, TT7-1 and TT7-3. Three additional piezometers were installed in the beach in 2012: TT5a, TT7a and TT7b (Fig. 3).

Pressure loggers were installed in the wells to record water level and temperature data at 20 min intervals. A barometer was also installed at the site to correct water levels for barometric pressure. For this study, we focused on pressure data for wells TT4-2, TT5a-3, TT7a and TT7b from June 2012. Groundwater salinities were collected and compiled for all the wells at Cabretta Beach on a regular basis and are presented in *Schutte et al.* [2016].

We conducted a Wenner array resistivity 2D profile along the transect at Cabretta beach from TT3b to seaward 133 m in August 2012. Electrical resistivity is sensitive to groundwater salinity and is useful for delineating the freshwater-saltwater interface in coastal settings (Choudhury et al., 2001). The survey was conducted using an AGI Super Sting R1 automatic switchbox and a waterproof cable with 28 electrodes and 3 m electrode spacing. We measured apparent resistivities along a transect 133 m long (1 roll forward) to a target depth of approximately 9 m below land surface. The survey was conducted during low tide. The roll forward was overlapped on the mid beach by 30 m in order to increase the signal to noise ratio. Electrical resistivity is affected by soil saturation and differences in geologic materials.

Approximately 14% of the measured apparent resistivity dataset was either negative or anomalously high. Low signal-to-noise ratios in those areas were likely caused by signal attenuation in highly conductive saline pore water (beach) near the surface in the former case, and signal dampening in highly resistive dry quartz sand (upland) in the latter. These data points were discarded before performing the inversion. Inverse modeling was conducted using AGI EarthImager 2D to convert measured apparent resistivities to a 2D resistivity section (AGI, 2009).

4.2 Numerical Methods

Simulations of tidally influenced groundwater flow were conducted using SUTRA, which is a model for saturated-unsaturated, variable-density groundwater flow with solute transport (Voss and Provost, 2002). The governing equation for the calculations used in SUTRA is a form of the Richards equation. We modified this equation to account for changes in overlying stress associated with tidal loading for the inundated section of the beach (Wilson and Gardner, 2006):

$$\nabla \cdot [K(\Psi)\nabla h] = S_w S_s \frac{\partial h}{\partial t} + \varphi \frac{\partial S_w}{\partial t} - \alpha_s S_w \frac{\partial \sigma_T}{\partial t} \quad (3)$$

where K is hydraulic conductivity, Ψ is negative pressure head, h is hydraulic head, S_w is water saturation, φ is porosity, σ_T is total stress and S_s is the specific storage,

$$S_s = \rho g (\alpha_s + \varphi \beta) \quad (4)$$

where ρ is the density of water, g is gravity, α_s is sediment compressibility and β is fluid compressibility.

The model domain was constructed to simulate groundwater flow from the forested upland seaward through the beach (Fig. 4). The domain was 300 m in length, and extended from TT4, across the beach, to the nearshore environment. The surface of the domain was created

using a topographic profile developed from elevation surveys, and it ranged in elevation from 2.75 to -2.5 m mean sea-level (MSL). The domain extended down through the subsurface to -10 m (MSL). Domain boundaries were chosen to ensure simulation results in the area of interest below the beach were not influenced by boundary artifacts.

A two-dimensional, finite element, irregular mesh with 16,058 nodes and 15,272 elements was used for all subsequent simulations. Elements ranged from a maximum size of 0.56 m² to a minimum of 0.01 m². Element size was refined near the surface, where flow rates were the greatest and salinity changed rapidly, to ensure the Peclet and Courant criteria were met. Saturation fronts contained at least 5 nodes to prevent problems with numerical convergence in the unsaturated zone.

The model domain contained 3 distinct hydrostratigraphic units with respect to porosity and permeability: (1) surficial silty, fine sand of forested upland; (2) surficial, well-sorted, medium grained sand of the beach; and (3) buried, silty mud layers (Table 1). We tested a reasonable range of permeability values for coastal sand deposits (10⁻¹⁰ to 10⁻¹² m²), as presented in *Wilson et al.* [2008]. The permeability values used for the upland and beach sands (Table 1) were ultimately chosen after a sensitivity analysis was conducted to match simulated hydraulic head with observed values. The max permeability of the mud layer was determined iteratively by comparing simulated groundwater salinities of the confined aquifer to electrical resistivity results and piezometer samples. The ratio of longitudinal to transverse dispersivity was held constant at 10 ($\alpha_L = 2.5 \alpha_T = 0.25$) (Gelhar et al., 1992; Smith, 2004).

The soil saturation in the unsaturated zone was calculated using the van Genuchten equation (van Genuchten, 1980):

$$S_w = S_{wr} + \frac{(S - S_{wr})}{[1 + (\alpha|\psi|)^n]^m} \quad (5)$$

where S_{wr} is the residual saturation of the sediment, ψ is suction pressure head, n , m , and α are fit parameters:

$$\alpha = \frac{1}{h_b} \left(2^{1/m} - 1 \right)^{1-m} \quad (\text{L}^{-1}) \quad (6)$$

where h_b is the air entry pressure. The fit parameters m and n are related by:

$$m = 1 - 1/n \quad (7)$$

The van Genuchten parameters used in the model are presented in Table 1.

Boundary conditions were selected to simulate transient groundwater flow driven by tides and salinity gradients (Fig. 4). The western boundary was assumed to be a no-flow boundary due to the presence of a flow divide in the forested upland near well cluster TT4 (Anderson, 2010). The bottom boundary was assigned a no-flow boundary representative of an underlying low permeability layer. The eastern boundary was extended offshore sufficiently that flow rates were negligible and a no-flow boundary was appropriate. The upper boundary was assigned a combined specified fluid pressure and flux boundary condition depending on the elevation of the tide and saturation. For surface nodes that were never inundated by the tide, the boundary condition was a specified fluid flux using hourly observed rainfall data. We assumed no surface runoff. For nodes that were inundated by the tide, the boundary condition was specified according to the height of the tide (Wilson and Gardner, 2006). In order to accurately match measured hydraulic head peaks at high tide, an additional 20 cm of head was added to the tidal signal boundary condition to account for wave-runup (Nielsen, 1990). These nodes were assigned a pressure representative of the weight of the overlying column of seawater. Surface nodes that were exposed above the height of the tide and were fully saturated developed a seepage face.

Initial conditions for tidal simulations were obtained by first running a simulation without tides. Initial pressure for every node was set to be equal to hydrostatic pressure. Initial salinity was equal to 35 for the entire model domain. Salinity is reported here using the Practical Salinity Scale of 1978 (UNESCO, 1981). Mean sea level was set to the average for June 2012 (0.23 m). The simulation ran for 30 years with 3-day time steps until quasi-equilibrium was reached. The results from this preliminary simulation were used as the initial conditions for the subsequent tidal simulations.

An initial tidal simulation using a simplified tide (sine wave) was then performed to calculate initial conditions for a final simulation. Tidal amplitude was 1.5 m and the period was 12.4 hours. This run required 550 days at 5 minute time steps to reach quasi-equilibrium with respect to pressure and salinity.

Next, a simulation was conducted using observed tide and precipitation data from June 2012 over two spring-neap tide cycles (28 days). When surface nodes were not inundated by the tide, a specified flux representative of either precipitation or evaporation was applied based on the rainfall record. During periods of no rainfall, a constant evaporation rate of 4 mm/day was applied to the surface nodes. Time steps of 1 minute were required to achieve numerical convergence during rapidly changing pressure and salinity conditions in the shallow subsurface.

Finally, we repeated this sequence of simulations with a second model domain where the buried, silty mud layers were removed and replaced with sand. The model was otherwise identical with respect to boundary conditions and hydrogeological parameters. These simulations were used to quantify the influence of the low permeability sediments on the configuration of the freshwater-saltwater interface and SGD in the beach.

The strength of the simulated USP that developed in the beach was determined by measuring a saline plume salinity gradient (SPSG) according to the method detailed in *Evans and Wilson* [2016]. In these theoretical models, the strength of the gradient is a function of density and tidally driven recirculation through the beach aquifer (Evans and Wilson, 2016). We caution that the SPSG is a simple linear measurement that does not fully describe the complex nonlinear interactions of the driving forces for groundwater flow and solute transport in a beach. However, the advantage of using a simple field measurement to gain potential insight into beach groundwater systems means that the SPSG merits further investigation.

Rates of SGD and saline water infiltration (Q_{SGR}) into the beach aquifer were measured in the tidal simulations for the month of June 2012. The total volumes of water infiltrating (Q_{TOT}) and discharging (SGD_{TOT}) from the surface boundary of the model were calculated for each time step by summing the groundwater flow velocity in each surface element and multiplying by porosity and element area. Infiltrating seawater was separated into density (Q_{DDR}) and tidally driven recirculation (Q_{TDR}) as described in *Evans and Wilson* [2016]. Q_{DDR} was seawater that infiltrated below the elevation of low tide and Q_{TDR} was seawater that infiltrated above. These fluxes were then normalized by dividing by the freshwater flux (Q_F) and multiplying by 100 to yield TDR and DDR as percentages to remain consistent with the literature (Robinson et al., 2007b). This technique works well for idealized simulations with a sinusoidal tide and specified freshwater flux boundary condition. Q_F and the position of low tide are harder to determine in simulations calibrated to real tide data and observed hydraulic head. We categorized TDR and DDR based on the elevation of mean low water (MLW) in this study. SGD was categorized by salinity using the USGS guidelines: fresh (SGD_F) (0 to 1); brackish (SGD_B) (1 to 10); and saline

(SGDs) (> 10). This allowed SGD rates to be reported with respect to proportions of fresh/brackish/saline water. All infiltrating seawater was 34 and rainwater was 0.

5. Results

5.1. Field Data

An electrical resistivity tomography survey to a target depth of approximately -10 m (MSL) suggested the presence of a complex salinity configuration under Cabretta Beach (Fig. 5). The root-mean-square error (RMSE) of the inversion between the measured and modeled resistivity was 11.5%. A small brackish-to-freshwater lens (10- 50 Ohm-m) near TT3b was modeled in the inversion under the forested upland of Cabretta Island. A brackish body of groundwater (3 -10 Ohm-m) was modeled in the confined aquifer beneath the mud layer in the beach. The unconfined aquifer in the beach was modeled to be saline (< 1.5 Ohm-m) except near the low-tide mark on the beach (TT7) where the water was slightly brackish about a meter below land surface (~ 3 Ohm-m) (Fig. 5). Using the resistivity survey and stratigraphic data, we interpreted the groundwater system to be strongly influenced by the presence of a buried mud/silt layer under the forested upland and beach. Fresh groundwater from the upland is diverted below the surficial beach aquifer, into the confined aquifer, where it discharges near the elevation of low tide on the beach. In order for fresh groundwater from the forested upland to flow into the confined beach aquifer, the confining mud layer must be incised somewhere beneath the freshwater lens. We estimated that this break in the confining unit was located approximately beneath the upper beach, near TT5a (Fig. 3).

Groundwater salinity was periodically sampled from the wells along the transect from September 2011 through the summer of 2012 (Table 2.) Wells in the forested upland and dune field were fresh to brackish, wells in the mid beach were saline, and wells near the average

elevation of low tide were still saline, but slightly fresher (~24). No fresh groundwater was measured below the beach during our study. Groundwater in the confined beach aquifer below the mud layer in the beach was sampled in the winter of 2012, between wells TT5 and TT6. We measured the salinity to be approximately 25, but the well was abandoned the same day after it filled with silt.

Hydraulic head records for wells TT4-2, TT5a-3, TT7b and TT7a for the month of June 2012 showed that variability in groundwater levels was mostly driven by daily tidal pumping and the spring/neap cycle (Fig. 6). TT4-2 was outside the propagation distance of the tidal energy. Thus hydraulic head was controlled by freshwater flux, drainage and evapotranspiration only. Water levels in the beach wells were approximately ~0.2 m higher than the elevation of the tide at high tide (Fig. 6). As previously indicated, this surplus head was most likely driven by wave-runup on the sloping beach face of the intertidal zone (Nielsen, 1990). The magnitude of the hydraulic gradient between the wells was greatest during falling and low-tide, when the majority of groundwater flow occurred in the beach. However, large spring tides and dry conditions during the first week of the month nearly reversed the hydraulic gradient over the transect at high tide. Hydraulic head at well TT7a was slightly higher than the head at well TT4-2, and was 30-50 cm greater than the head at TT5a-3 (Fig. 6).

5.2. Numerical Model Calibration

We calibrated the groundwater model to salinity measured in the field by adjusting permeability values for the hydrostratigraphic units (Table 1, 2). In general, our model successfully reproduced observed groundwater salinities from the piezometers at Cabretta Beach. The values for permeability reported in Table 1 were the only permeabilities that allowed accurate reproduction of the observed salinity distribution. Large permeabilities in the beach

sand and buried mud layer caused the entire beach subsurface to become saline. No fresh groundwater was present in the confined beach aquifer using sand permeabilities greater than $\sim 6 \times 10^{-11} \text{ m}^2$. Lower values for permeability in the beach sand led to erroneously fresh groundwater in the upper beach due to reduced seawater infiltration during rising and high tide. The largest difference between simulated and observed groundwater salinity was at well TT4-4 (~ 3.8). Deviations between simulated and observed groundwater salinities were all within a reasonable range of variability for a beach system. The distribution of groundwater salinity in the beach model was mostly controlled by the permeability of the buried, silty mud layers, for which a value of $5 \times 10^{-15} \text{ m}^2$ best reproduced the observed salinities. This permeability was also required to accurately simulate groundwater head in the forested upland region of the model.

We also attempted to calibrate the model to observed hydraulic head data. Due to the dynamic nature of the beach at Cabretta Island, we estimate that our measurements of elevation were accurate to within $\pm 25 \text{ cm}$ for the study period. Highly precise calibration between simulated and measured hydraulic head was not possible. The average head at well TT4-2 over the study period was 1.85 m, with two prominent peaks above 2 m during periods of sustained rainfall (Fig. 7a). This rapid increase in water table height could have been supported by high residual soil saturation in the island upland. A permeability of $1 \times 10^{-14} \text{ m}^2$, representative of silty sand, was required to sustain this head at the western boundary of the model. The location of well TT4 was a boundary node in our model. Minor edge effects caused the model to be less sensitive to precipitation at TT4 than the measured data suggested. There may have also been a discrepancy between the actual rainfall at the field site and the rain gauge, which was located approximately 2 miles away. A permeability of $5.8 \times 10^{-11} \text{ m}^2$ best reproduced hydraulic head at well TT5a. Simulated and measured hydraulic head were in close agreement except for a period

of approximately 7 days, where the model results under-predicted head during low tide (Fig. 7b). The modeled results were slightly higher during the first few high tides of the month compared to the field data (Fig. 7b). These differences were probably caused by a small discrepancy between the actual elevation of land surface and that used in the model, which affected the location of well TT5a in the tidal frame.

Groundwater head for beach wells TT7a and TT7b was calibrated using a permeability of $5.8 \times 10^{-11} \text{ m}^2$ (Fig. 7c-d). We were unable to reproduce the troughs in hydraulic head during periods of no rainfall in the measured data (Fig. 7c-d). Van Genuchten fit parameters did not impact the model significantly enough to reproduce the observed head at low tide. Results were much more sensitive to permeability. In order to force the model to reproduce the measured hydraulic head at low tide, a permeability of $9 \times 10^{-10} \text{ m}^2$ was required. This value is unrealistic for a beach with a median grain size of approximately 0.3 mm. Test simulations that reached quasi-equilibrium using this permeability in the beach resulted in uniform groundwater salinities of 34 throughout the entire domain in the tidal range. We propose that the discrepancy between the measured and simulated hydraulic head at low tide was caused by uncertainty in the total thickness of overlying sediment above the screened interval of the well (erosion), uncertainty in elevation datum, and evaporation in the beach during periods of drought. Although the model was unable to exactly reproduce the observed hydraulic head data for the lower beach, the general impact of spring/neap tides and precipitation on groundwater flow was captured.

5.3. Model Results

Our simulation results showed that a moderate strength USP developed under the upper beach around mean high water (MHW), adjacent to a zone of slightly fresher, but still saline groundwater discharge at the elevation of mean low water (MLW). A standard FSI developed

seaward of this zone of discharge (Fig. 8a). The salinity of the groundwater within the upper saline plume was approximately 31.5, and the salinity of the groundwater that discharged near MLW was 24. The average SPSG (Evans and Wilson, 2016) for June 2012 was 0.075 (Table 3). No fresh groundwater was present under the beach, existing only in the freshwater lens beneath the upland.

Groundwater flow paths were greatly impacted by subsurface heterogeneity. In general, groundwater flowed downward under the forested upland and into the confined aquifer where it mixed with saline groundwater and eventually discharged at the terminus of the silty mud layer (Fig. 8b). Some groundwater discharged slightly above the elevation of the highest spring tides, near well TT5a ($x = 25$ m). Above the confining unit in the beach, a flow divide developed between wells TT5 and TT6, at the approximate elevation of MHW (Fig. 8b). West of this flow divide, infiltrating seawater flowed slightly landward, where it eventually flowed downward and through the break in the confining unit and mixed with fresh groundwater from the upland. East of the divide, groundwater flowed rapidly seaward and discharged at changes in beach slope and at the elevation of MLW. Density-driven convection drove seawater to infiltrate the subsurface seaward of the FSI (Fig. 8b). Groundwater flow ranged from 1 to 2 cm/day in the forested upland, 5 to 100 cm/day in the permeable beach sands, 1 to 2 cm/day in the silty mud layers, and less than 1 cm/day offshore where the sediment surface was always inundated by the sea (Fig. 8b).

We calculated total fluxes of SGD and SGR from Cabretta Beach. The total volume of SGD over a single tide was 5.7 m^3 per meter of shoreline (Table 3). The total volume of seawater to infiltrate the beach aquifer over one tide (Q_{SGR}) was 2.6 m^3 per meter of shoreline (Table 3). Recirculated seawater accounted for approximately 46% of the total SGD from Cabretta Beach

in one tide. The remainder originated as fresh water that became saline through mixing below the beach.

Simulation results for the month of June 2012 were further examined with respect to the spring/neap lunar cycle. A large spring tide occurred during the first week of June, reaching MHHW of approximately 1.5 m and MLLW of -1.3 m. Groundwater salinities and flow velocities were averaged over the spring tide period (Table 3; Fig. 9a). Impacted by the higher high tides associated with the spring lunar cycle, the center of the upper saline plume was transported up the beach approximately 10 m and was shortened about 35% compared to the June 2012 average. The salinity of the center of the upper saline plume increased from about 31.5 to 32.7, and the salinity at the seepage face near MLW decreased slightly to 24. The width of this seepage face increased 200% (10 to 30 m) for the spring tide cycle compared to the month average (Fig. 7a; Fig. 9a). The location of the FSI did not change. The magnitude of the SPSG increased from the month average of 0.075 to an average of 0.084 for the spring tide cycle.

Groundwater flow directions for the spring tidal cycle were mostly unchanged when compared to the monthly average results. The only significant change was a landward shift of the flow divide in the upper beach approximately 10 m, from MHW to MHHW (Fig. 9a).

Groundwater flow velocities increased significantly during spring tides in the upper beach and confined beach aquifer. Shallow groundwater flow above the buried mud layer in the beach increased from an average of about 30 – 60 cm/day to 60-100 cm/day. Velocities in the confined beach aquifer increased from an average of about 40 cm/day to approximately 65 cm/day (Fig. 9a).

Following the spring tide cycle, a neap tide cycle lasted through the second week of June 2012. During this period, MHW decreased to about 0.9 m and MLW increased to about -0.75 m.

Groundwater salinities and flow velocities were averaged over the neap tide cycle and reported (Table 3; Fig. 9b). Driven by the shift in the tidal range, the center of the USP migrated down the beach approximately 15 m compared to the spring tide location (~5 m below the month average). The total width of the USP increased 40% (~10 m) from the spring tide and salinity remained constant (Fig. 9b). The salinity in the upper 1 m of the seepage face adjacent to the USP increased by about 2 as compared to the spring tide. The location of the FSI did not change. The magnitude of the SPSG decreased from 0.084 for the spring tide cycle to 0.072 for the neap cycle (Table 3).

Groundwater flow paths for the neap tide cycle were unchanged compared to the spring tide results except that the flow divide in the upper beach migrated seaward approximately 15 m (Fig. 9a-b). Groundwater flow velocities decreased significantly compared to the results for the spring tide cycle in the upper beach and confined beach aquifer. Above the confining mud layer, groundwater velocities decreased from approximately 60-100 cm/day for the spring tide cycle to about 20-50 cm/day for the neap cycle (Fig. 9b)

Total fluxes of SGD and SGR were calculated through the spring/neap tide cycle (Table 3). The SGD flux during a spring tide cycle at Cabretta Beach was 5.8 m³ per meter of shoreline per tide. Over 99% of this SGD was saline groundwater. The total volume of Q_{SGR} over a single tide was 2.5 m³ per meter of shoreline. Recirculated seawater accounted for approximately 43% of the total SGD during this period. The SGD flux during the neap tide cycle at Cabretta Beach was 5.5 m³ per meter of shoreline per tide. Virtually the entire volume of SGD was saline. Brackish and fresh groundwater discharge made up less than 1% of the total SGD. The total volume of Q_{SGR} was 2.7 m³ per meter of shoreline per tide. Recirculated seawater accounted for approximately 49% of the total SGD over the neap cycle.

The configuration of the freshwater-saltwater interface was significantly different when the buried silty mud layers were removed (Fig. 10a). Most notably, the depth to which the USP extended beneath the beach surface increased approximately 4 m. The width of the USP beneath the beach remained the same, approximately 90 m. Beneath the forested upland, the freshwater lens extended the entire depth of the model domain. Groundwater salinities were overall much higher for the simulation without confining units. The average salinity of groundwater in the center of the USP was approximately 33 and the salinity in the adjacent seepage face near MLW was 28. The location of the FSI remained around $x = 150$ m. The average SPSG over the simulation was 0.063 (Table 3).

Groundwater flow paths were also significantly different for the simulation without buried mud beneath the beach. Average groundwater flow paths for the month showed a large convection cell across the freshwater-saltwater interface that developed approximately below the dune ridge ($x = 20$ m) (Fig. 10b). This convection reversed the direction of groundwater flow beneath the forested upland from downward to upward, where it eventually discharged on the high beach ($x = 25$ m). Slightly seaward of this seepage face, a large tidally-driven circulation cell developed where groundwater generally flowed downward from the upper beach and discharged near the elevation of MLW (Fig. 10b). The flow divide that was present in the previous simulations did not develop here, suggesting this phenomenon was controlled by the presence of a buried mud layer beneath the beach. Groundwater flow velocities were relatively similar in the two simulations for the forested upland, shallow beach subsurface and offshore. Beneath the beach, velocities were higher (75 – 100 cm/d). Total volumes of SGD and SGR were very similar whether the mud layer was included or not (Table 3).

The flux of seawater that infiltrated the beach aquifer over the month of June 2012 was separated into tidally driven and density driven recirculation (TDR and DDR) (Table 3). We only report average values for the month of June 2012 for the simulations with and without a buried mud layer. For the simulation containing the buried mud layer, Q_{TDR} into the beach over a single tide was 1.8 m^3 per meter of shoreline. Q_{DDR} was 0.8 m^3 per meter of shoreline per tide. Normalized using Q_F , TDR and DDR were 60% and 25% respectively. Q_{TDR} and Q_{DDR} were 1.9 and 0.6 m^3 per meter of shoreline per tide for the simulation without the mud layer. Normalized using Q_F , TDR and DDR were 64 % and 21% respectively. TDR increased slightly in the simulation without the buried mud, which drove a decrease in the magnitude of the SPSG. DDR also decreased in this simulation due to a lower salinity gradient in the beach.

6. Discussion

Our results illustrate several differences between theoretical models and the real beach at Cabretta Island. Cabretta Beach supported a wide, dispersed, moderate-strength USP (SPSG = 0.075) with an average beach slope of approximately 0.03. Based on estimates from *Evans and Wilson* [2016], a theoretical beach similar to Cabretta Beach should contain a weaker USP. A beach with a SPSG of 0.075 required a beach slope of approximately 0.05. The strength of the plume was enhanced because the relict marsh mud layer under the beach decreased mixing between tidally-driven infiltrating seawater and the land-derived groundwater below. This confined aquifer system under the beach also permitted fresher groundwater to discharge near the elevation of MLW. Therefore, the inclusion of geologic heterogeneity in the shallow subsurface of the beach increased the strength of the USP actually supported under Cabretta Beach.

The inclusion of a realistic topographic profile in numerical simulations also led to deviations from theoretical predictions for groundwater flow associated with tidal amplitude and spring/neap tidal cycles. Spring tides at Cabretta Beach during June 2012 had tidal amplitudes of approximately 3 m, and neap tides about 2 m. Using theoretical predictions from the literature as guidelines, spring tides should have higher rates of SGR (sum of Q_{TDR} and Q_{DDR}) into the beach than neap tides, and spring tides should have a lower SPSG (weaker upper saline plume). However, our results indicated the contrary: spring tides had a slightly lower flux of SGR into the beach, and a higher SPSG. This deviation was caused by variation in slopes across the beach profile at Cabretta. TDR and DDR increase with increasing beach slope when beach slopes are less than or equal to 0.1 (Evans and Wilson, 2016). The total flux of SGR increased at Cabretta Beach during neap tides because the average slope of the intertidal zone for neap tides was slightly higher than it was during spring tides. Furthermore, the magnitude of the SPSG was lower for neap tides even though the tidal amplitude decreased.

The location of the freshwater discharge tube (FWT) is controlled by the elevation of MLW, which is a product of the average elevations of low tide through the semi-diurnal and spring/neap tidal cycles. The average elevation of low tide during the neap cycle was only slightly further down the beach than the position of the FWT. During spring tides, low tide was nearly 50 m further down the beach. Therefore, the location of the FWT on the beach was inundated by the sea for a longer portion of the tidal cycle during neap tides than during spring tides. This more frequent inundation period and associated SGR increased the salinity in the FWT, which decreased the magnitude of the SPSG. If the tidal amplitude were to remain equal to the neap tide amplitude (2 m), the system would re-equilibrate and the position of the FWT would shift landward, and the resultant SPSG would likely be higher. The neap tidal period of one week is

not long enough to induce this change. The position of the FWT and USP in the beach are the product of the combination of dynamic tidal cycles and beach topographic profile. Clearly a difference exists between theoretical beach models with a static tidal amplitude and uniform beach slope, and beach models that include these heterogeneities.

Despite the differences discussed above, the SPSG measured at Cabretta Beach was similar to what would be predicted using a theoretical model only.. The SPSG in Cabretta Beach was approximately 0.075 for the month of June 2012. According to *Evans and Wilson* [2016], a SPSG of 0.075 would lead to a value of approximately 3.5 for the ratio of the width of the intertidal zone to tidal propagation distance (ϵ) in a beach, given a tidal amplitude of 1.25 m and a hydraulic conductivity of 10 m/d. Furthermore, their study suggested that a value of 3.5 for ϵ would result in a TDR of ~55% and a DDR of ~9%. Our model results indicate that for the month of June 2012, TDR and DDR were equal to 60% and 25% respectively. Although not perfect, these results for TDR and DDR in Cabretta Beach are similar to what *Evans and Wilson* [2016] predicted using theoretical, numerically convenient beaches. The magnitude of ϵ decreases with increasing values of hydraulic conductivity, which could explain why the predicted theoretical values for TDR and DDR are lower than those calculated in this study. *Evans and Wilson* [2016] used a hydraulic conductivity of 10 m/d. We found a hydraulic conductivity of ~50 m/d ($k = 5.8 \times 10^{-11} \text{ m}^2$) to best represent Cabretta Beach. These comparisons suggest that first-order estimates for TDR and DDR could be obtained using a simple field measurement in the beach and the numerical relationships of *Evans and Wilson* [2016] and *Robinson et al.* [2007b].

Our results indicate that the average SGD flux for Cabretta Beach is equal to approximately 5.7 m^3 per meter of shoreline per tide. Expanded to the total coastline length of

Sapelo Island (22 km), the SGD flux is approximately $2.6 \times 10^5 \text{ m}^3/\text{d}$. Georgia has a total coastline length of approximately 160 km, which suggests a total SGD flux of approximately $1.9 \times 10^6 \text{ m}^3/\text{d}$. To put this volume of water into context, the Altamaha River, which discharges to the Atlantic Ocean just south of Sapelo Island, has an average discharge of $1.8 \times 10^7 \text{ m}^3/\text{d}$ (USGS 84 year mean statistic). This river is the third largest contributor of freshwater to the Atlantic from eastern North America (Frangiamore and Gibbons, 2016). When compared directly, SGD from the beaches of Georgia contribute approximately 10% as much water to the Atlantic as the Altamaha River. The average width of the beaches on Sapelo Island is about 200 m, and the length of shoreline is about 22,800 m. This gives a total beach surface area of about 4.6 km^2 . The total area of the Altamaha River drainage basin is about $36,000 \text{ km}^2$. When accounting for total discharge per unit area, the beaches of Sapelo Island discharge approximately 120 times more water to the Atlantic Ocean than the Altamaha River ($\sim 6 \times 10^{-2} \text{ m}^3/\text{m}^2/\text{d}$; $\sim 5 \times 10^{-4} \text{ m}^3/\text{m}^2/\text{d}$). Beaches clearly have the potential to discharge large volumes of water to the coastal ocean that could have significant impacts on nearshore chemistry and biogeochemical cycles.

Groundwater dynamics at Cabretta Beach moderated significant biogeochemical processes (Schutte et al., 2015, 2016). Our results support the published hypothesis that a nitrous oxide hotspot on the upper beach at Cabretta was supported by infrequent inundation by high spring tides (Schutte et al., 2015). This zone became unsaturated during neap tidal cycles, when the concentration of N_2O decreased. During high spring tides, this zone was recharged with a mixture of infiltrating seawater and fresh groundwater from the upland that could potentially supply enough NO_3^- to maintain rates of N_2O production (Schutte et al., 2015). Moreover, Schutte et al. [2016] showed that methane export from a zone of methanogenesis in the

freshwater lens under Cabretta Island was largely controlled by methanotrophy in the buried coastal sediments. They showed that in the unconfined, surficial aquifer of the beach, rates of methanotrophy were high enough to act as a sink for CH₄, preventing significant export to the ocean. They indicated that a more plausible transport pathway for CH₄ to the coastal ocean was down through the confined aquifer buried under the beach. *Schutte et al.* [2016] used a groundwater flow velocity estimate from a previous version of the Cabretta groundwater flow model that predicted an average flow rate (~0.025 m/d) in the confined beach aquifer. This model did not include a break in the confining unit, and as a result, groundwater flowed much more slowly beneath the beach. They used this velocity to calculate an average residence time of greater than 10 years. Our new results suggest a much higher representative average flow velocity in the beach confined aquifer of ~0.4 m/day, and an average residence time of closer to 1 year. Higher velocities and therefore a shorter residence time could allow the confined aquifer under the beach to export higher concentrations of CH₄ to the coastal ocean than previously considered.

Although we attempted to calibrate our numerical model to data collected in the field as closely as possible, truly verifying and validating a numerical model of natural Earth systems is not possible (Oreskes et al., 1994). As a result, numerical models at their best can only provide a rough conceptual framework for groundwater dynamics and SGD fluxes at the coast. Real beach systems can exhibit extreme local heterogeneities with respect to the subsurface sedimentology, topographic profile, sediment transport, and dynamic forces such as waves, tides and storms, which could potentially drive extreme variability in local groundwater flow systems. As a result, numerical models are useful mostly for providing best estimates of how beach groundwater systems behave despite the inherent assumptions and non-uniqueness that models themselves

contain. However, SGD is extremely difficult to measure in the field, especially at large spatial scales, and numerical models currently provide one of best methods for quantifying these fluxes.

7. Conclusion

We found that a calibrated numerical model of beach groundwater flow that included observed heterogeneities, such as layered stratigraphy and variable beach topography, differed in some ways from conceptual models developed using theoretical, numerically convenient beaches only. Real beaches are unlikely to be comprised solely of isotropic, well-sorted permeable sands to depths of 10-30 m, or to have uniform slopes. At Cabretta Beach, a buried, silty mud layer allowed the beach to support a stronger USP than a similar beach that consisted of only permeable sands. Variations in slope across the beach profile allowed slightly larger fluxes of SGR to occur during neap tides than during spring tides. The elevation of mean sea-level did not fall symmetrically between MHW and MLW on the surface of the beach, which caused significant variability in the size and extent of the upper saline plume between spring and neap tidal cycles. This asymmetry is probably common in real beaches.

Our results show that Cabretta Beach had an average flux of 5.7 m^3 per meter of shoreline per tide for the month of June 2012. Virtually all of this SGD to the coastal ocean was saline. The average flux of SGR was 2.6 m^3 per meter of shoreline per tide. Approximately 69% of this flux was due to TDR, and 31% was due to DDR. After replacing the buried mud layer with permeable sand, the total SGD and SGR fluxes were constant, but TDR increased 4% and DDR decreased 4%. This shift in the proportions of TDR and DDR that made up the total SGR was consistent with a decrease in the magnitude of the SPSG.

Additionally, we note that our results support the hypothesis in *Evans and Wilson* [2016] that a simple field measurement (SPSG) could be used to develop preliminary estimates for SGD and

SGR in a beach. We found the calibrated Cabretta Beach groundwater model results to be very similar to the predicted values for the SPSG, TDR/DDR, and ε in *Evans and Wilson* [2016] for theoretical beaches with similar hydrologic properties. Although our results were in close agreement with studies using theoretical beaches only, more extensive investigation into the reliability of using SPSG measurements to calculate rough estimates for TDR/DDR fluxes is required.

Finally, our results confirm that flow systems hypothesized to support biogeochemical cycling are hydrologically feasible. Our results support a conceptual model in which flow through a dynamic unsaturated-saturated zone under the upper beach, which was controlled by the spring-neap tidal cycle, could allow rapid nitrogen cycling and subsequent nitrous oxide production to occur (Schutte et al., 2015). This N₂O hotspot was located at the interface between land-derived meteoric groundwater and seawater that infiltrated the shallow subsurface during high spring tides. Additionally, the confined aquifer beneath the beach potentially facilitated transport of CH₄ from a zone of high methanogenesis under the island upland, out to the coastal ocean (Schutte et al., 2016). We calculated average flow rates through the confined aquifer of approximately 0.4 m/d, which resulted in a revised residence time calculation of approximately 1 year. This lower residence time could result in higher fluxes of CH₄ from the beach to the coastal ocean. Our results from Cabretta Beach suggest that important biogeochemical cycles in coastal sediments are largely influenced by the complex groundwater flow dynamics under the beach.

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Figure

Figure 1

Figure

Figure 2

Figure

Figure 3

Figure 4

August 8, 2012

R (Ohm-m)

Figure 5

June 2012

Figure 6

Figure

Figure 7

Figure 8

Figure 9

a)

b)

Figure 10