ground water

Studying Ground Water Under Delmarva Coastal Bays Using Electrical Resistivity

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Abstract

Fresh ground water is widely distributed in subsurface sediments below the coastal bays of the Delmarva Peninsula (Delaware, Maryland, and Virginia). These conditions were revealed by nearly 300 km of streamer resistivity surveys, utilizing a towed multichannel cable system. Zones of high resistivity displayed by inversion modeling were confirmed by vibradrilling investigations to correspond to fresh ground water occurrences. Fresh water lenses extended from a few hundred meters up to 2 km from shore. Along the western margins of coastal bays in areas associated with fine-grained surficial sediments, high-resistivity layers were widespread and were especially pronounced near tidal creeks. Fresh ground water layers were less common along the eastern barrier-bar margins of the bays, where sediments were typically sandy. Mid-bay areas in Chincoteague Bay, Maryland, did not show evidence of fresh water. Indian River Bay, Delaware, showed complex subsurface salinity relationships, including an area with possible hypersaline brines. The new streamer resistivity system paired with vibradrilling in these investigations provides a powerful approach to recovering information required for extension of hydrologic modeling of shallow coastal aquifer systems into offshore areas.

Introduction

Fresh or brackish ground water in submarine environments has been shown to exist on a range of scales and in a variety of geologic settings. Interstitial water studies, conducted during drilling operations on the continental shelf off the Atlantic margin of Florida in 1965, revealed the presence of submarine fresh water as far as 100 km from shore (Manheim 1967). Subsequently, most of the Atlantic shelf was shown by U.S. Geological Survey (USGS) drilling operations in 1976 to be underlain by fresh and brackish water (Hathaway et al. 1979; Manheim and Paull 1981). Extensive submarine discharge networks are associated with the Floridan Aquifer in Florida, Georgia, and South Carolina (Bush and Johnston 1988; Sprinkle 1989; Swarzenski et al. 2001). Most of the offshore fresh water observations reviewed by Kohout et al. (1988) outside the Floridan Aquifer, and distant from shore, can be attributed to relict fresh water recharged into the aquifers of the exposed shelf during the late Pleistocene glacial maximum and not yet replaced by salt water penetrating from the sea floor (Meisler et al. 1984).

Active hydrodynamic phenomena (i.e., submarine ground water discharge) along the U.S. Atlantic margin are attracting increasing attention. Summaries are provided by Bratton et al. (this issue) and Burnett and Kontar (2002). Simmons Jr. (1988) suggested that direct ground water flux into Chesapeake Bay might supply nutrients equivalent to those contributed by a major tributary. The pathways for such flux and discharge, however, are poorly known. The present paper provides details of the distribution of submarine ground water occurrence and discharge associated with extensions of surficial aquifers beneath Delaware, Maryland, and Virginia coastal bays.

In 2000, the USGS, in cooperation with scientists from the University of Delaware, initiated the first of several new technological approaches to the investigation of submarine discharge in coastal bays of the Delmarva Peninsula (Figure 1). The objective of the work was to elucidate the pathways of direct discharge of ground water to the bays, since ground water was presumed to deliver a significant proportion of the high nutrient loads reported for the

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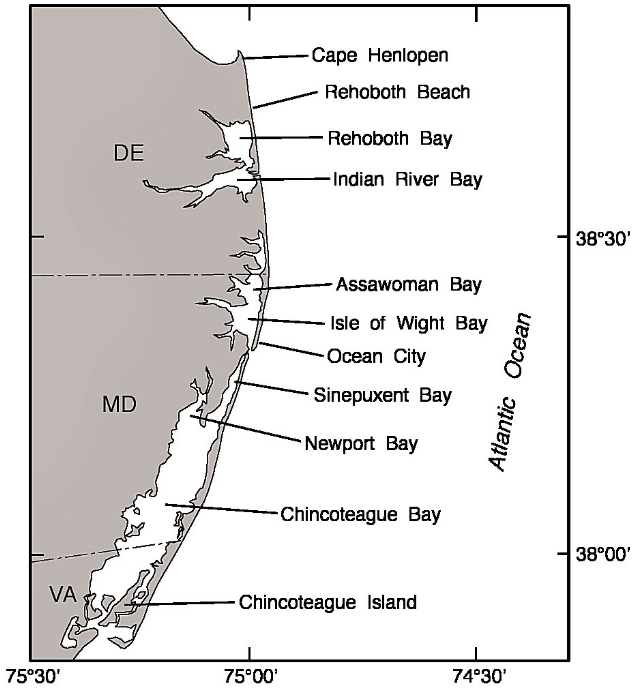

Figure 1. Location map showing the Delmarva coastal bays.

Delaware and northern Maryland coastal bays (Andres 1987, 1992; Cerco et al. 1994; Valigura et al. 2000). As a part of the larger Coastal Intensive Sites Network program (Ullman et al. 1993; Ullman et al. 2001), thermal remote sensing by aircraft confirmed submarine discharge in the form of warmer water anomalies detected in coastal areas during late winter of 2000 (McKenna et al. 2001). The extent, pathways, and composition of discharging ground water, however, remained poorly known. Further background on the current hydrogeologic studies is given by Krantz et al. (this issue). Modeling approaches applied to the Maryland bays provide regional insight on the connection between onshore aquifers and submarine discharge (Dillow and Greene 1999) and show important data gaps that the current study attempts to address.

To investigate the submarine occurrence of fresh ground water, an integrated streamer resistivity system that had been tested in the Ohio River in 1997 (Snyder and Wightman 2002) was deployed experimentally in the shallow brackish waters (20‰ to 30‰) of Rehoboth Bay and Indian River Bay in Delaware. Continuous horizontal resistivity profiles produced from this survey provided evidence of submarine discharge along much of the western margin of the bays. Based on the results of these surveys, Hoverprobe, a hydraulic vibradrilling rig mounted on a hovercraft (Phelan 2000), was deployed in Chincoteague Bay at Public Landing, Maryland, in 2001. This effort was designed to test for the presence of submarine fresh water in a similar setting to areas surveyed by the streamer resistivity technique in Delaware, and to allow fluid and sediment samples to be collected and analyzed prior to performance of a similar resistivity survey in the Maryland coastal bays.

The second resistivity survey was conducted in Assawoman, Isle of Wight, and Sinepuxent bays, Maryland, and Chincoteague Bay, Maryland and Virginia, in 2001 (Manheim et al. 2002). Results from this survey also detected fresh submarine ground water along most of the bays surveyed, with the exception of areas along the eastern shorelines of the bays adjacent to the barrier bars. Hoverprobe tests yielded direct evidence of shallow fresh water, and recovered samples for nutrient analysis in the vicinity of Public Landing, Chincoteague Bay, in 2000. Hoverprobe, designed primarily for marsh operation, was able to recover cores by anchoring in open water, but the hovercraft platform was not sufficiently stable under these conditions to permit the deeper vibradrilling, coring, and logging that its drilling system can potentially deliver. Consequently, in October 2001, a vibradrilling rig was mounted on a barge (stabilized with spuds) provided by the Delaware Department of Natural Resources and Environmental Control for follow-up drilling and logging in Indian River Bay (Krantz et al. this issue). In this investigation, four holes were drilled to a maximum depth of 30 m below the sediment surface. This paper focuses primarily on the resistivity surveys with supplementary information from selected cores and surface salinity analyses. Krantz et al. (this issue) and Bratton et al. (this issue) report detailed logs, regional hydrostratigraphic interpretations, and interstitial water chemistry. These three papers were prepared as complementary treatments of submarine hydrogeology of the Delmarva coastal bays.

Methods

Streamer Resistivity Surveys

Principles and Brief History

Electrical resistivity measurements detect variations in the conductivity of subsurface water and porous media. Not surprisingly, horizontal (dc) resistivity methods were among the earliest geophysical techniques developed for study of subsurface geologic and ground water properties on land during the first half of the 20th century (Heiland 1940). After World War II, electrical resistivity techniques were successfully used on land (Zohdy and Jackson 1969; Bisdorf and Zohdy 1979) to delineate fresh water/salt water boundaries in coastal areas. However, several factors have inhibited wider use of dc electrical resistivity methods compared with the widespread deployment of borehole resistivity and conductivity logging systems. The first is the time-consuming process of establishing electrode contact and making multipole measurements. The second issue is the complexity of inversion modeling (conversion of resistivity/dipole plots to resistivity layer/depth models). The final issue is the difficulty in separating the influences of pore-fluid conductivity from lithologic influences on bulk resistivity.

Salt water resistivity studies have been even more limited than land-based use. U.S. oil companies in the 1950s employed long streamer cables to map offshore salt domes in the north-central Gulf of Mexico, but published descriptions of this work are not widely available. Russian geophysical surveys (Kalashnikov et al. 1980) successfully mapped fresh water discharge into the Caspian Sea. Very recently, time- and frequency-based electromagnetic methods for resistivity determination have been extended to marine applications (Evans et al. 1999; Fitterman et al. 1999; Fitterman and Deszcz-Pan 2002). The current studies, however, are among the first to employ multichannel streamer resistivity techniques for systematic surveys of ground water discharge in coastal environments.

The Zonge Co. (Tucson, Arizona) first deployed the survey system used here on the Ohio River in 1997. Snyder and Wightman (2002) describe the equipment and some of the principles involved in the inversion modeling. The 120 m multichannel streamer system (Figure 2) permits instantaneous multipole measurements to be made at 2 s intervals, with maximum depth of measurements ~ 0.33 the length of the electrode array. This allows high resolution while the streamer is towed at speeds of up to 5 kn. After merging navigation data with resistivity data, processing software creates an optimized (smoothed) inversion model of the data. Finally, data are displayed in the form of colored resistivity-layer cross sections or profiles, using standard commercial plotting software, Surfer^{TM} (Golden Software Inc., Golden, Colorado). Corroborative studies were performed using vibradrilling, resistivity-probe measurements, and pore-fluid analyses. Fluid resistivity varied from < 0.25 Ω m for bay waters, to ~25 Ω m for fully fresh waters—a factor of 100. In contrast, changes in resistivity owing to lithologic changes could be shown through formation-factor analysis (discussed later) to vary generally over a range less than a factor of four. Consequently, changes in the salinity of pore fluids at the interface between fresh and salty ground water exerted much greater influence on resistivity signals than did variation in sediment properties. In the majority of cases, this permitted direct interpretations of the subsurface hydrology within the limits of vertical resolution for the profiles, as shown in the ensuing figures.

Gassy sediments are well documented in Chesapeake Bay (Hill et al. 1992), and were widely observed in the Delmarva coastal bays during seismic surveys (Krantz et al. this issue). The ebullition of methane gas creates a high acoustic contrast that inhibits penetration of acoustic waves to deeper horizons. Gas bubbles are expected to have a minor effect on resistivity distribution, particularly since methane concentrations (and bubble density) are generally highest in the upper few meters of sediment based on work in Chesapeake Bay (Pohlman et al. 2000).

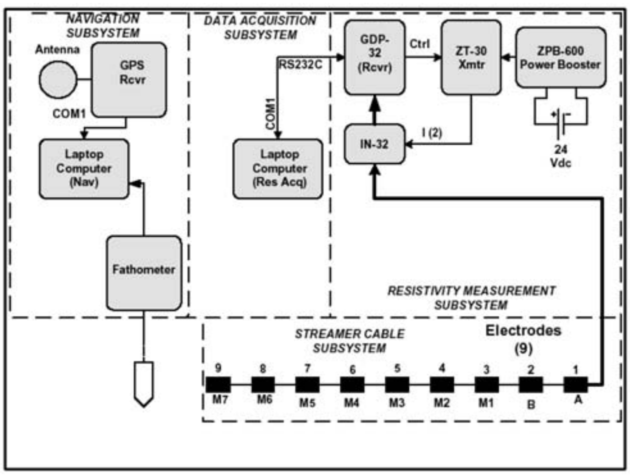

Figure 2. Schematic diagram for streamer resistivity configuration (adapted from Snyder and Wightman 2002).

Data Collection and Processing Methods

Resistivity surveys were carried out with vessels and pilots provided by the USGS Office of Ground Water, Branch of Geophysics, Storrs, Connecticut, for the Delaware surveys, and the National Park Service, Assateague Island National Seashore, for the Maryland-Virginia surveys (May 2000 and May 2001, respectively) (Manheim et al. 2002). Streaming resistivity data were acquired using a standard Zonge GDP–32 multifunction resistivity/induced polarization receiver together with a small battery-operated transmitter. Figure 2 shows a block diagram of the onboard instrument system and streamer (Snyder and Wightman 2002). Additional details are provided by Manheim et al. (2002) and Snyder and Wightman (2002). Two 12 V marine batteries provided power. For the Delaware surveys, ~4 A of power was utilized with cruising speeds of 2.5 to 3 kn. For the Maryland-Virginia surveys, 8 A was employed at cruising speeds up to 5 kn. The receiver operated at a frequency of 4 Hz.

After the cruises, the separate navigation files, captured by HyPack^{TM} software (Middletown, Connecticut), were merged with the resistivity file by the Zonge TS2DIP software. The output was delivered by the Zonge Co. in the form of three profile plots (Figure 3) using Surfer software. The bottom two plots are (from bottom to top) observed apparent-resistivity and calculated apparent-resistivity pseudosections derived from a moving average of observed resistivities and plotted against dipole number. The top profile shows the inversion (smoothed model) resistivity contours plotted against depth below sea surface. Because of the navigation input, the lines were processed in feet for the Delaware survey and meters for Maryland-Virginia. Dropouts (loss of data) in deeper raw resistivity data cause most of the artifacts in the profiles.

The TS2DIP computer program used for the inversion modeling of the Delaware data and part of the Maryland data does not discriminate between water column and sediment, which limits interpretation of shallow layers. At the request of the USGS, the Zonge Co. prepared a modified software package that permitted correction for water-column effects. This software improves the rendering of resistivity distribution in shallower sediments. Merging shallow water and sediment resistivities for the Delaware and some Maryland survey lines mainly affects data within the first dipole (10 m) and should not significantly affect deeper values.

Interpretive Postprocessing

The initial, high-resolution profiles of the resistivity data were processed in ~1100 m sections. Although the high-resolution profiles are useful for detailed local analysis (Krantz et al. this issue), methods were developed to condense and smooth the voluminous high-resolution sections for regional analysis, removing artifacts while doing so. For the Delaware survey, the raw modeled (T2SDIP) data were concatenated, filtered, and smoothed using database and spreadsheet software, and then plotted with

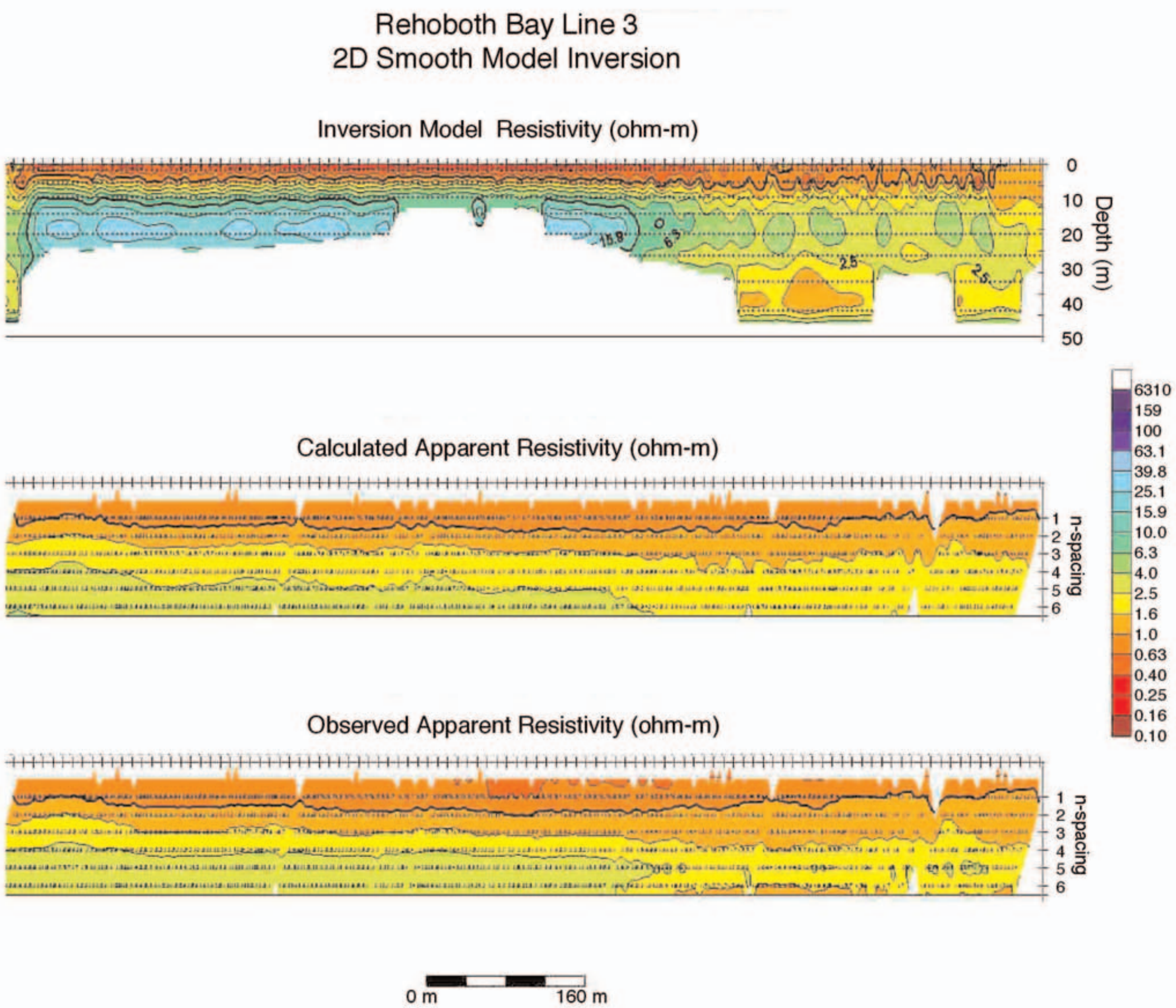

Figure 3. Triad of resistivity profiles for Delaware Line 3, produced by Zonge Co. using TS2DIP and Surfer software. The lower two pseudosections (dipole/resistivity plots) represent observed resistivity (bottom) and calculated observed resistivity (middle). The top section depicts the interpreted resistivity contours plotted against depth (inversion model).

Surfer. The Maryland data, configured in meters, were smoothed directly in Surfer. Anomalous data, defined as modeled resistivity values > 1.8 Ω m for the zero-depth dipole, were removed for Delaware data, but some artifacts remain in the Maryland data. As with the original Zonge inversion profiles, resistivity contours for the condensed regional profiles are masked at depths corresponding to limits for reliable interpretation, based on a threshold index of measured signal-to-noise ratio. Finally, track lines were plotted on topographic base maps, created using National Geographic Topo^{TM} (National Geographic Maps, Evergreen, Colorado) software (based on USGS topographic map data).

Core and Interstitial Water Methods

The first open-water Hoverprobe deployment (Figure 4) obtained four cores up to 4 m long in Chincoteague Bay at Public Landing, Maryland. Sediment squeezing and in situ pumping with a screened drive point obtained water samples, as described in detail by Bratton et al. (this issue). A special sediment resistivity probe (Manheim and Waterman 1974) yielded high-resolution resistivity data on split cores, which, in conjunction with resistivity of the pore fluid, provided formation factors to aid in interpretation of the streamer resistivity profiles. Formation factor is defined

Figure 4. Hoverprobe deployment at Public Landing (Chincoteague Bay), Maryland. Hoverprobe is self-launching via boat ramps.

as R_s/R_w, where R_s is the true sediment resistivity, and R_w is the resistivity of interstitial water at 20°C. Formation factors calculated from probe resistivities, as well as electrical logs from adjacent land boreholes, can be applied to modeled resistivity data to convert them to interpreted pore water salinity (Manheim et al. 2002). Cores other than those from the vibradrill sites at Public Landing (with PL designations) referred to in this paper were taken using a 1 m PVC piston corer that was manually pushed or hammered into beach sand or shallow sediments.

Resistivity Survey Results

Delaware Coastal Bays

Indian River Bay

Resistivity track lines from Indian River Bay (Figure 5) showed continuous high-resistivity zones beneath Assawoman Canal and Indian River, a shallow tidal stream that is the largest fresh water source for Indian River Bay. In Indian River, high-resistivity zones were largely continuous from bank to bank (Figure 6), as interpreted from angled crossings, and extend beyond the river mouth. The Line 7A profile (Figure 6) shows shallowing of the high-resistivity fresh water layers upstream in Indian River. The width of the stream mouth at the confluence with Indian River Bay is ~1 km. This and other tidal streams in the Indian River/Rehoboth Bay estuary system are characterized by brackish waters from < 10‰ to 20‰, where salinity fluctuates with tidal cycle and season (Ullman et al. 1993; Ullman et al. 2001; Cerco 1994). High-resistivity zones were also observed beneath Pepper Creek and Herring Creek. As previously noted, the inversion software used for Delaware Bay did not discriminate between the shallow water column and surficial sediments in the interpretive model. This means that for the upper dipole (10 m), the effects of water column and sediment resistivities are mingled, which lowers apparent resistivities for the uppermost sediment layers. It also tends to shift the apparent vertical position of fresh water anomaly features downward from their true level. The decreasing upstream salinity (changing from 19‰ to 4‰) reduces this effect, and partly accounts for the apparent upstream decrease in depth to the high-resistivity layers.

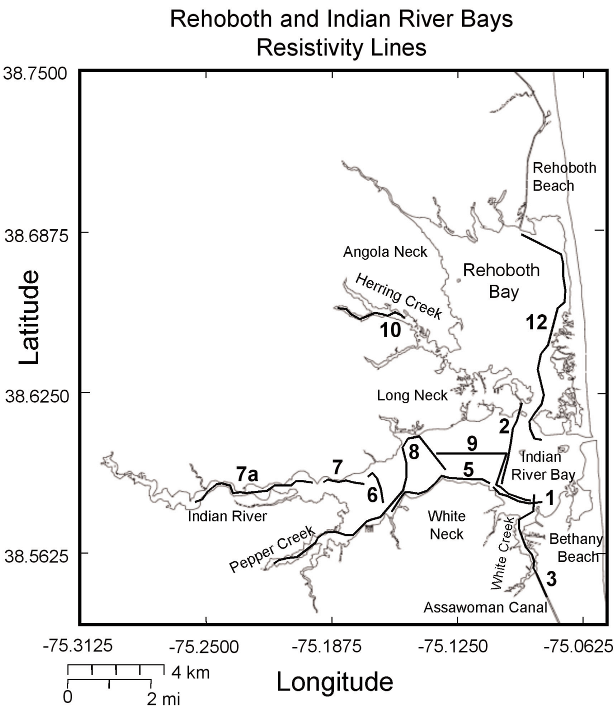

Figure 5. Resistivity survey tracklines in Rehoboth Bay and Indian River Bay, Delaware.

Indian River Bay Cross Section

The upper layer of saline ground water thickens in central Indian River Bay. Below this layer, submarine fresh water layers from the surficial aquifer protrude under the bay from both the northern and southern shores to depths of ~20 m (Figure 7). The shallow fresh lenses are underlain by brackish water, with dispersive mixing taking place both from above and below. Laterally, abrupt changes from brackish to fresh subbay ground water correspond with locations of buried offshore paleovalleys, or changes from lower-lying marshy coastlines to areas of higher onshore topography (Krantz et al. this issue). In some areas, such as east of Ellis Point, high offshore resistivity anomalies are also present adjacent to fringing salt marshes. In such cases, ground water is inferred to flow beneath the marshes, supported by recharge from adjacent uplands (Howes et al. 1996).

The high-resistivity wedges observed in Figure 7 were documented to correspond to intervals of low electromagnetic conductivity in borehole logs taken during vibradrilling operations in the White Neck area during October 2001 (Krantz et al. this issue). The fresh water layers were confirmed to represent moving ground water by age dating and the presence of oxygen in the deeper layers (Bratton et al. this issue). These oxygenated waters occurred beneath black anoxic sediments located in a vertical transition zone from bay salinities at the top to fresh water below.

Central Section of Indian River Bay, Including a Hypersaline Anomaly

In the western part of the cross-bay traverse (Figure 8, Line 9A and western end of Line 9B), close to the area probed by vibradrilling, a lens of moderately brackish to fresh ground water was observed at depths of 10 to 20 m. This feature is consistent with the vibradrilling results and other cross-bay transect lines. However, an unexpected feature was observed in the easternmost part of Line 9B. The interpreted true resistivity data values (unsmoothed by the contouring process) reach values corresponding to hypersaline brines; i.e., greater than open-ocean salinity of 34‰ and possibly in excess of 50‰. Consultation with the Zonge Co. and examination of the detailed records found no known source of artifacts. The observations are elaborated on in the "Discussion" section.

Maryland–Virginia Coastal Bays

Resistivity profiles in the western parts of the northern Maryland coastal bays—Assawoman Bay and Isle of Wight Bay (Figure 1)—showed ground water features similar to

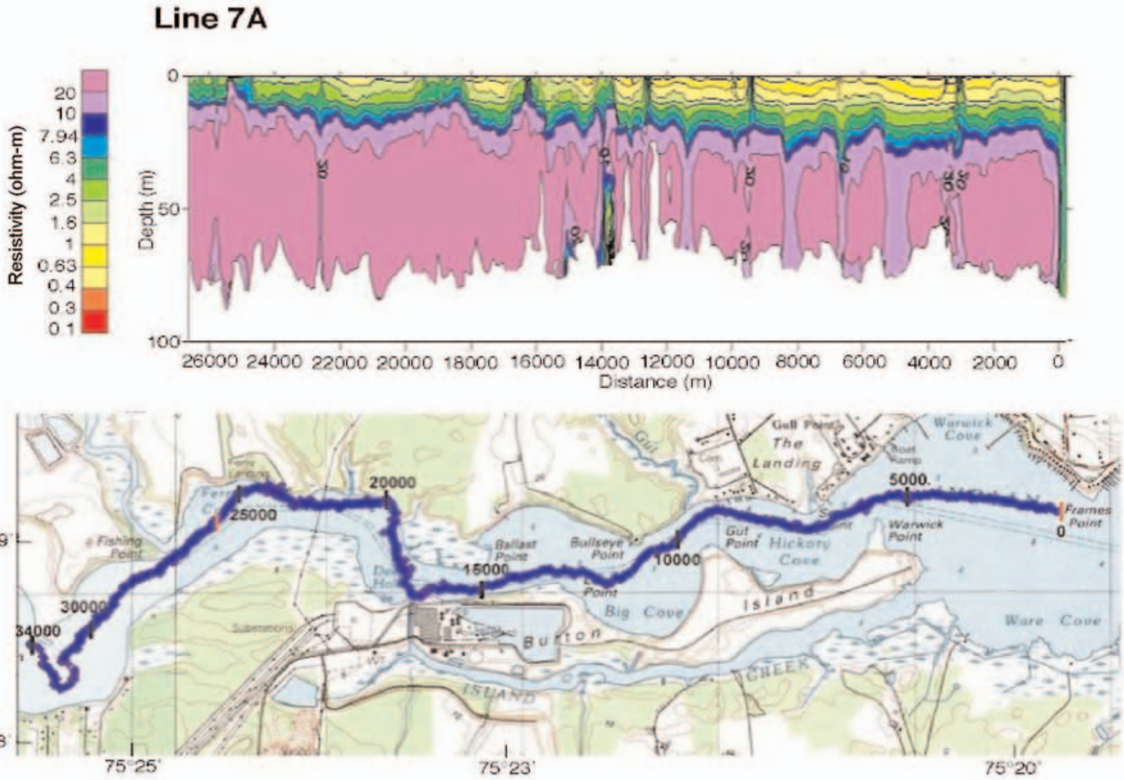

Figure 6. Resistivity Line 7A from Indian River, Delaware. This and other lines (where not otherwise referenced) were processed with Surfer software. Lower limits of contours have been masked at the limit for reliable interpretation, based on an index of measured signal-to-noise ratio. Note that in the 2000 surveys navigation data were collected in Delaware State Plane coordinates in feet. The 2001 surveys in Maryland and Virginia obtained navigation in Universal Transverse Mercator coordinates in meters.

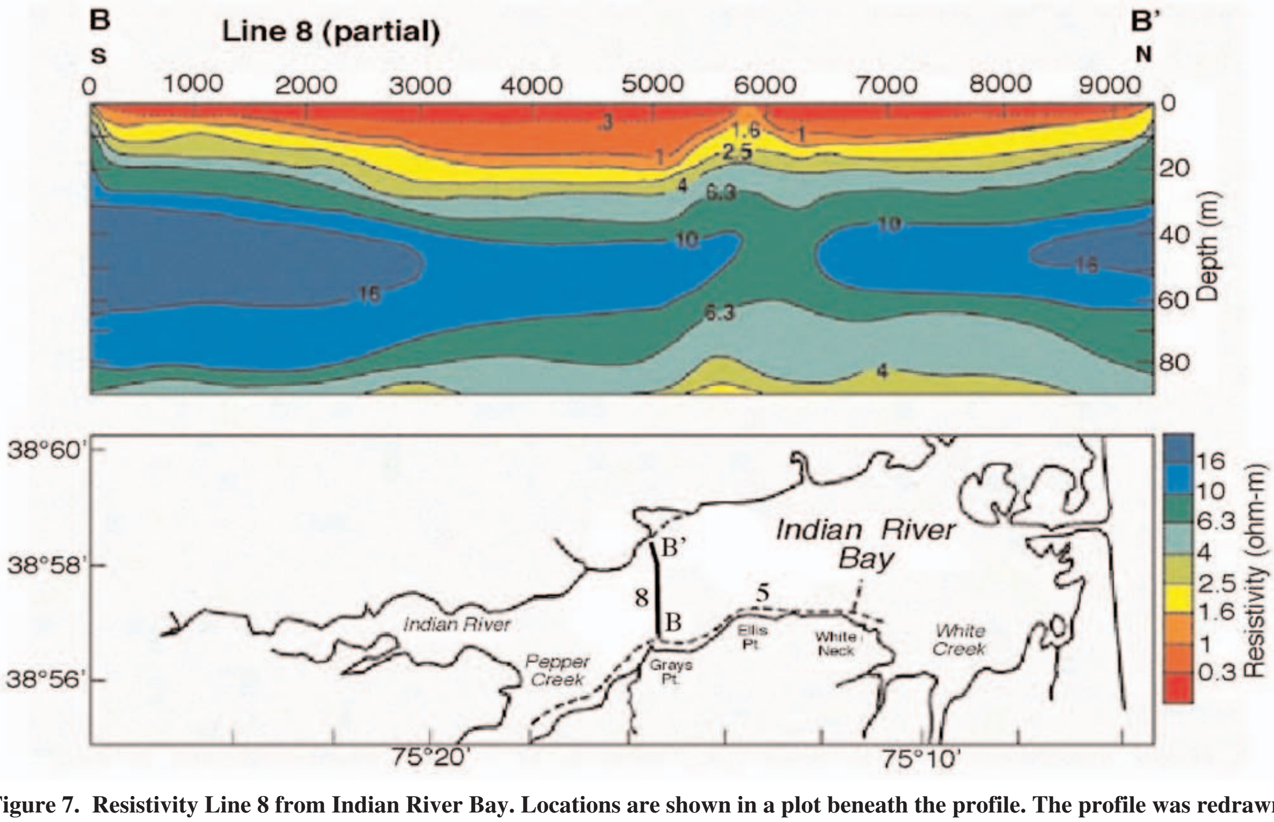

Figure 7. Resistivity Line 8 from Indian River Bay. Locations are shown in a plot beneath the profile. The profile was redrawn manually from initial high-resolution Surfer output.

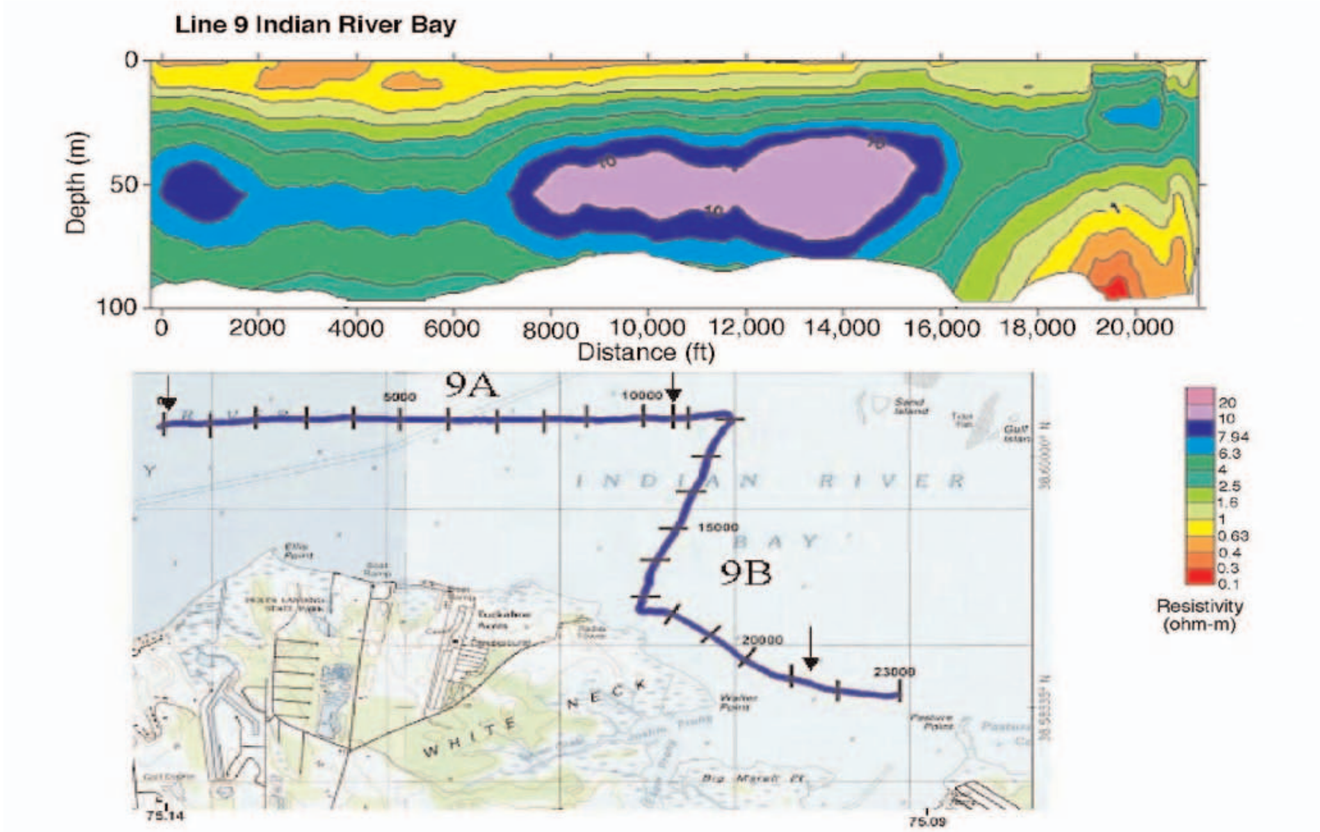

Figure 8. Resistivity Line 9 from Indian River Bay. Note that the line takes a right angle turn at ~10,800 ft from origin.

those seen in the western sections of the Delaware coastal bays. In Chincoteague Bay (Maryland and Virginia), on the western end of Line 6 in Figures 9 and 10, around Public Landing, Maryland, Hoverprobe vibradrilling had already documented the presence of shallow fresh ground water. Questions to be answered by the cross-bay resistivity transects, Lines 4 and 6, included the following. (1) How far does fresh water in the surficial aquifer extend beneath the bay? (2) Are there mid-bay anomalies that would suggest discharge from deeper aquifers (Dillow and Greene 1999)? (3) What is the nature of ground water occurrence on the barrier-bar side of the bay?

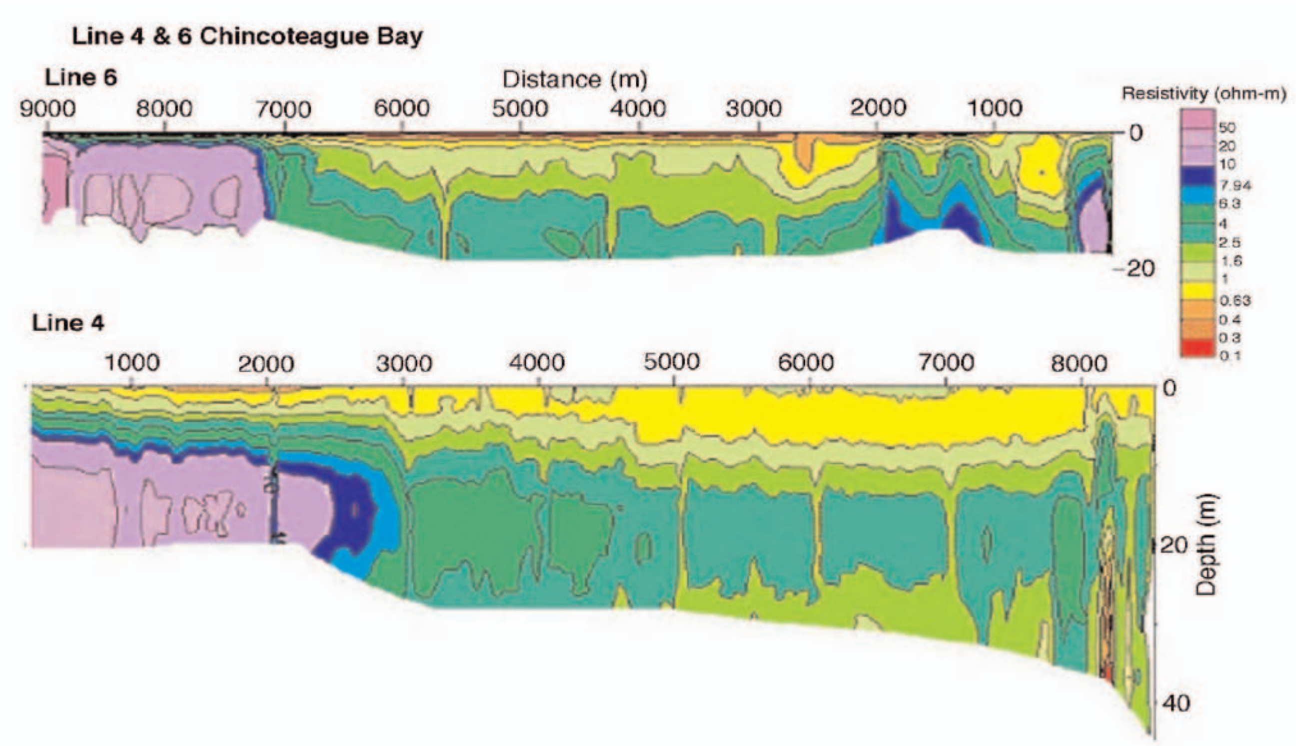

Figure 9. Resistivity Lines 4 and 6 from Chincoteague Bay. Note that depths for the Maryland-Virginia lines are in meters (see Figure 6 caption for explanation). Location shown in Figure 10.

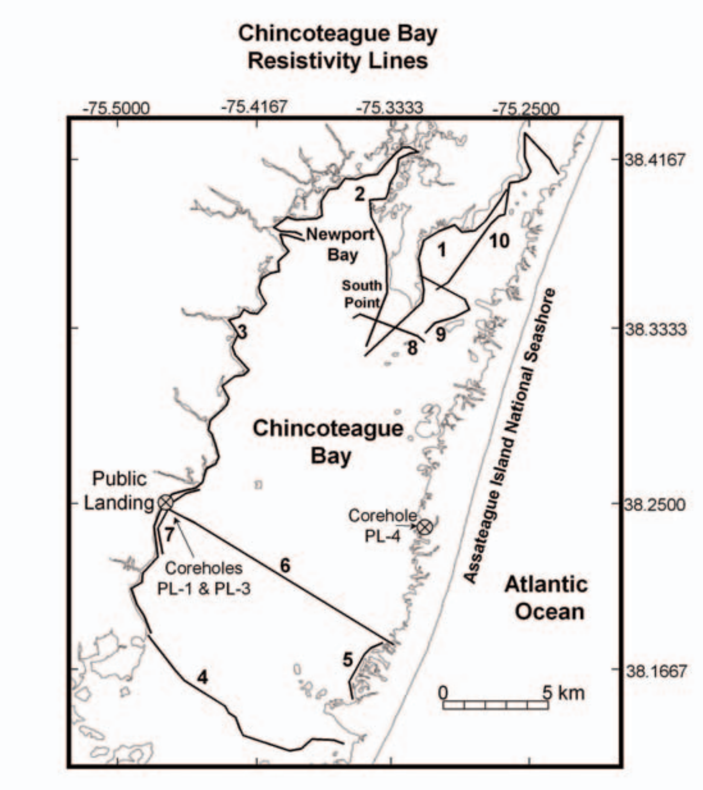

Figure 10. Map of resistivity tracklines and coreholes in Newport, Sinepuxent, and Chincoteague bays, Maryland.

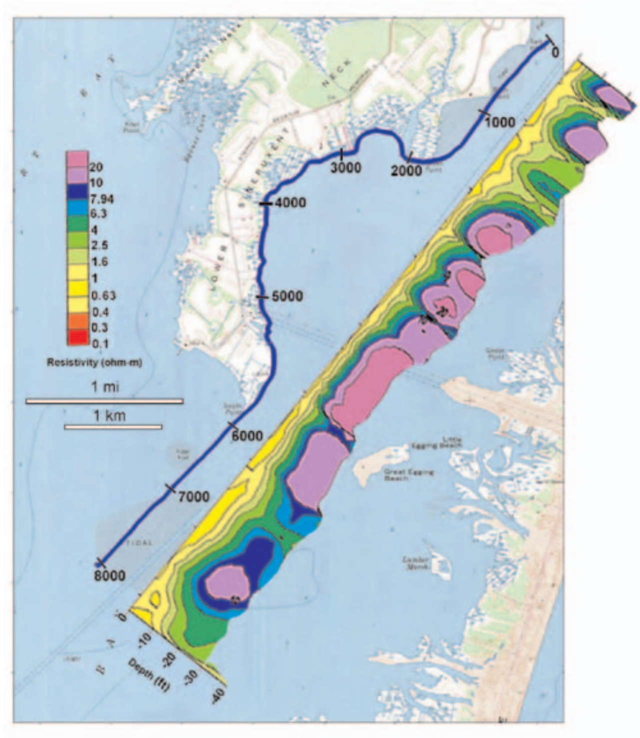

Figure 11. Resistivity Line 1 from Chincoteague Bay, paralleling the Sinepuxent Neck paleoshoreline trend (no water-column correction). The presence of saltier water in the subsurface is partially correlated with the distance of the track from shore.

Both Lines 4 and 6 in Figure 10 show evidence of bayward extension of fresh ground water from the western shore. Neither line shows any evidence of fresh water beneath the central part of the bay within the depth limit of information (~30 m). The survey lines do, however, display hydrological differences. Resistivity Lines 1 through 4 in Maryland were processed with the earlier TS2DIP software system used for the Delaware survey; Line 6 was processed with advanced software that corrects for water-column conductivity. Thus, Line 6 provides a more accurate picture of shallow conditions within the sediments. In both lines, vertical breaks in continuity are relicts of data processing that have not been removed. Notwithstanding the difference in processing software, it is clear that fresh water layers extending into Chincoteague Bay are deeper in Line 4 than in Line 6 and have more gradual dispersion into the central bay sector. A layer of brackish water, but no fresh water, extends west from Assateague Island on the eastern end of Line 4. In contrast, submarine fresh water in Line 6 is much shallower than in Line 4 on the west end, and terminates sharply in the seaward direction. This may correlate with occurrence of buried submarine peat with high fluid transmissivity, which terminates bayward in fine-grained sediments. Such peats were encountered at a variety of coring and vibradrilling sites. A small fresh water layer associated with the landward side of Assateague Island appears on the east end of Line 6.

Resistivity Line 1, extending along Sinepuxent Neck into Chincoteague Bay, demonstrated sharp lateral variations in resistivity along the part of the track paralleling the land area (Figure 11). Part of this variability is related to the distance of the track from land where the track cut across embayments and indentations. As in Indian River Bay, however, modern topography and buried paleotopography play an important role. The most hydrogeologically significant structure is a linear, northeast/southwest-trending feature (Sinepuxent Neck) that roughly parallels the offshore track (southwest part) of Line 1, and is associated with reemergence of a high-resistivity anomaly more than 1 km offshore. A transect of this feature in a direction perpendicular to Line 1 is captured by Line 8 (Figure 12). Lines 10 and 11 (not shown) demonstrated continuous presence of fresh water beneath 10 to 15 m of sediment in Sinepuxent Bay. Lines 4, 5, 6, and 9, as well as more southerly lines not illustrated, in Figure 10 provided resistivity data for the Assateague Island side of the bay. These lines revealed only the intermittent presence of high-resistivity submarine layers.

Surface Salinity Observations and Selected Core Salinity Data

Surface Salinity Measurements

The salinity/conductivity of bay surface water was monitored regularly during the resistivity surveys. Measured salinities in Indian River Bay varied from ~30‰ near

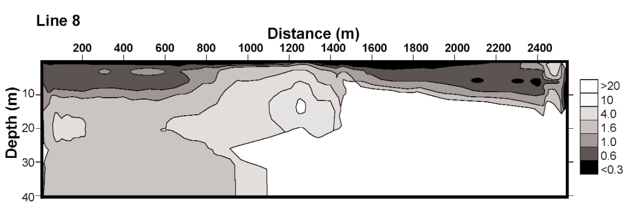

Figure 12. Resistivity Line 8 from Chincoteague Bay transverse to Line 1. The interpreted profile is corrected for water-column effect (see Figure 6 caption).

inlets to the Atlantic Ocean (e.g., Indian River Inlet and Ocean City Inlet) to values as low as 4‰ in the farthest upstream reaches of the Indian River tributary. Average values in the central parts of the Delaware coastal bays and in western Chincoteague Bay were ~26‰ to 27‰ during the periods of study. Salinities varied smoothly in the central parts of bays, but sharper fluctuations were observed near land. A zigzag cruise track taken during the Line 4 transect along the southern coast of Indian River Bay near White Neck is shown in Figure 13. The figure shows decreases in salinity of 1‰ to 2‰ as the vessel approached the shore. The Line 5 track (not shown) showed similar reductions in salinity during close approaches to shore, whereas temperature showed only minor variations. Similar changes were ubiquitous along the western margins of the Maryland and Virginia coastal bays; i.e., salinities decreased from ~30‰ to ~27‰ or below on approaching shore. The reduction was greatest where large subsurface (high) resistivity anomalies were registered. Nearshore freshening of surface water increased near tidal creeks, but was continuously present at some level regardless of the proximity to creeks or other surficial sources of fresh water. Changes in surface salinity were noted up to ~500 m from shore. Such changes were observed only intermittently upon approaching the eastern barrier-bar shorelines of the bays. In short, the surface salinity data point to nearly continuous (both spatially and temporally) discharge of fresher water at the shoreline and up to 500 m offshore actively influencing the composition of coastal surface waters.

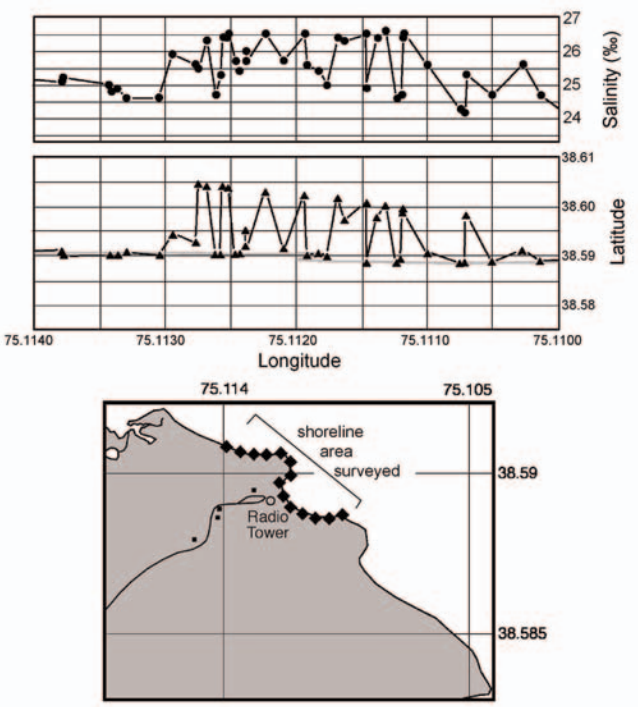

Figure 13. Surface salinity variability along the southern shore of Indian River Bay (Line 4). The lowermost plot shows latitude changes along the track. The upper plot shows changes in salinity along the track. Note the correspondence between latitudinal and salinity changes along the track, i.e., salinities increase bayward.

Core Data

In Indian River, opposite the power plant (Figure 6), a shallow piston core showed 10‰ pore water salinity in gray-black surficial sandy sediments. Salinities declined sharply upon encountering peat units and were ~0‰ below 22 cm (Table 1). Similar high-to-low salinity transition zones were observed in shallow cores in Assawoman Canal, which was characterized by surface water salinities ~20‰, and in Hoverprobe cores from Herring Creek, Rehoboth Bay, Delaware, which recovered orange iron-stained sediments similar to those described by Charette and Sholkovitz (2002).

Near Ellis Point, in Indian River Bay, a beach core taken in front of a marsh and a few meters from the bay showed modest diminution of salt from 25.5‰ to 23.8‰ in the upper 60 cm, and then a sharper drop to 18.8‰ below 80 cm. A black, sulfidic layer occurred above 80 cm. Below the black layer, after a short transition, the sediment became gray to yellowish, with a decrease in salinity. From these observations, implying oxygenated conditions, we infer that oxygenated, fresher ground water was moving seaward not far below 90 cm depth. Another core taken directly offshore in 1 m water depth (during intermediate tidal conditions) showed salinity and sediment lithology similar to the beach core down to the depth retrieved, although the bottom section was lost during core recovery.

At vibradrilling site WN–1, off Holts Landing State Park on White Neck (Figure 5) (Krantz et al. this issue; Bratton et al. this issue), a sharp salinity reduction to ~7‰ occurred at a depth roughly corresponding to the bottom of the aforementioned beach core. The combined observations suggest that shallow ground water flow from the land to the adjacent coastal bay commonly occurs, whether the conditions for deeper fresh water plumes exist or not.

The first Hoverprobe core in the Public Landing area, Chincoteague Bay (Figures 10 and 14, PL–1), was obtained

Table 1
Lithologic Description of a Shallow Piston Core from Indian River Collected Near Indian River Power Plant

Depth (cm)	Lithology	Salinity (‰)
0–3	Gray to black clayey, sulfide-smelling mud	10.4
3–7	Black-dark clayey mud, with white and stained orange shell matter, possibly from upper layer; fine-grained sticky, with wood and other plant fragments, 3–10 mm long; H_2S	
7–12	Same	6.2
12–17	Sand, greenish-gray to black, with shell fragments	
17–22	Gray to black mud, sulfidic	1.2
27–33	Peat, dark reddish-brown, with minor silty-sand	< 0.3
38–43	Peat, dark reddish-brown woody fibers	< 0.3
48–53	Peat, dark reddish-brown woody fibers	< 0.3
60–68	Peat, dark reddish-brown woody fibers	< 0.3

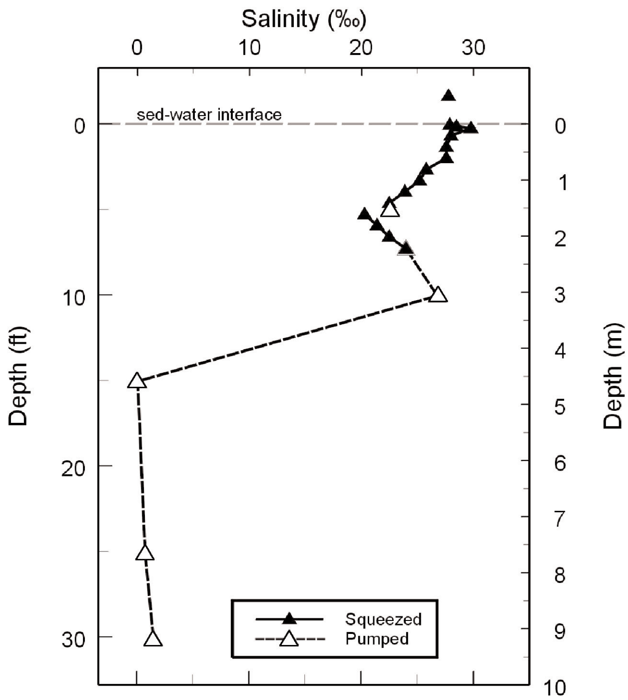

Figure 14. Vertical salinity profile for core PL–1. Note correspondence between independently sampled and analyzed pore water and pumped water salinities.

~120 m offshore, with the vessel tethered to a long public pier. Interstitial waters revealed saline, anoxic strata to depths of ~2 m, followed by a sudden drop below 3.1 m to fresh waters (Table 2). Good agreement between squeezed and pumped samples was obtained, as shown in Figure 14. The sudden drop in salinity was probably sharper than the sampling resolution. Even at site PL–3, ~700 m from shore, freshened waters were encountered at shallow depths (Table 3). Because of the improved processing software utilized for Line 6 (Figure 9), the shallow fresh water demonstrated in the core data also shows clearly in the interpreted resistivity profile. The subsurface transition from brackish to fresher water at the PL–3 site correlated with the depth of a buried peat layer, as observed at other sites in Delaware.

A vibradrill core was taken from the salt water marsh on the bay side of Assateague Island (Figure 10, PL–4; Table 4). This core had live marsh vegetation (*Spartina* sp.) at the top, beneath which peat with a sulfidic odor extended down to a clay-rich layer. Contrary to the expectation that pore fluid in the entire core would be saline or even hypersaline (due to evaporation and infiltration of salt water ponded on the marsh during spring tides and storms), salinities decreased with depth, dropping from 32‰ at the top to 20‰ at the bottom of the core (3.5 m).

Resistivity-probe measurements made on the PL cores (Tables 2–4) correspond well with lithologic boundaries. From the resistivity measurements made on the cores themselves (R_s) and conversion of the interstitial fluid salinities to pore fluid resistivity values (R_w), the formation factor can be calculated ($F = R_s/R_w$). F values may then be used to interpret formation salinity from horizontal resistivity profiles; i.e., $R_w = R_s/F$, where F is the formation factor for sediment at a given temperature. The probe values were

Table 2
Core Parameters for Core PL-1, Taken at the End of the Pier at Public Landing, Chincoteague Bay, Maryland

Depth (cm)	Lithology	C_w (mS/cm)	Salinity (‰)	R_s (Ω m)	R_w (Ω m)	F
0–3	Black sand and clay, thinly interbedded, H_2S	39.44	27.9		0.254	
3–6	Same	40.19	28.5		0.249	
6–9	Same	41.8	29.8		0.239	
9–22	Same	39.56	28	1.48	0.253	5.86
22–44	Same	39.06	27.6	1.24	0.256	4.84
44–62	Black clay, H_2S	39.06	27.6	0.63	0.256	2.46
62–82	Black clay, H_2S	36.01	25.2	0.73	0.278	2.63
82–102	Silty clay	36.01	25/2	0.98	0.278	3.53
102–120	Silty clay	33.67	23.4	1.15	0.297	3.87
120–142	Gray clay	32.49	22.6	0.9	0.308	2.92
142–156	Gray sand	32	22.6	1.45	0.313	4.64
156–162	Same	29.57	20.3	1.42	0.338	4.2
162–182	Same	31.03	21.4	1.36	0.322	4.22
182–202	Same	32.49	22.5	1.28	0.308	4.16
202–226	Same	34.45	24	1.23	0.290	4.24
306	NA	37.7	26.8		0.265	
459	NA	0.282	< 0.2		35.4	
765	NA	1.3	0.7	7.69		
918	NA	2.49	1.35		4.02	

Resistivity values in Ω m; calculated conductivity values (C_w) of water in mS/cm. Resistivity values are interpolated from raw data to correspond with depth of salinity measurements. F refers to formation factor = R_s/R_w.

Table 3
Core Parameters for Core PL-3, Collected ~500 m Toward Center of Chincoteague Bay from PL-1

Depth (cm)	Lithology	Salinity (‰)	R_s (Ω m)	R_w (Ω m)	F
0–3	Silt and clay, gray, worms	30.2		0.24	
3–6	Silt and clay, gray, worms	29.8		0.24	
6–9	Silt and clay, worm tubes	30.5		0.23	
9–28	Silt and clay, gray tubes	28.3	0.77	0.25	3.08
28–53	Silt and clay, gray homogeneous	23.5	0.73	0.3	2.47
53–78	Same		0.9		
78–108	Yellow-brown peat, sharp interface, mixed with silt and clay	15.9	0.97	0.42	2.32
108–128	Peat and silty clay		1.01		
128–153	Dense orange-brown peat	8.5	1.3	0.77	1.68
153–163	Peat and silty clay		1.43		
163–175	Muddy burrows (gray)	7.7	1.82	0.85	2.14
175–190	Change to silt and clay, dark gray with 10% peat				
190–203	Sparse organics		2.27		
203–230	Silt and clay, gray-brown	4.5	3.62	1.44	2.52
230–244	Silt and clay, gray-brown		3.65		

Notes as in Table 2

supplemented with information from electrical logs from nearby land drill holes used to create interpreted salinity profiles across Chincoteague Bay Line 4 (Manheim et al. 2002).

Discussion

Formation Factor

The formation factor ($F = R_s/R_w$) relating sediment resistivity to interstitial water resistivity is important because it permits interpretation of true formation salinity from resistivity profiles. In Tables 2–4, probe-resistivity measurements on the cores (R_s) are listed along with salinities and interstitial resistivities interpreted from them. Interstitial water resistivity (R_w) is converted to salinity using equations derived from data in standard chemical handbooks for conductivity and salinity of sea water at 20°C (i.e., $S = 7.042 \times R_w^{-1.0233}$, where S is salinity in ‰ and R_w is water resistivity in Ω m). Resistivity in Ω m can be converted from conductivity in mS/cm by the relationship $R_w = 10/C_w$, where C_w is conductivity in mS/cm.

The formation factor shows systematic differences among sites PL–1, PL–3, and PL–4 (Tables 2–4). Site PL–4, on Assateague Island's inner marsh, is underlain by coarse sand and shows formation factors of ~5. PL–3, which is situated within the inner-bay mud zone, has formation factors of 2 to 3. PL–1 is in a transition zone from sandy nearshore beach sediments to bay muds, and shows formation factors intermediate between PL–4 and PL–3 with an especially low F value in a dense peat horizon. The presence of permeable peat associated with poorly permeable clays helps account for highly anisotropic hydraulic transport characteristics, as well as the abrupt termination of the fresh water layers seaward (Figure 8). Peat and clay both have high porosity and low formation factors in the

Table 4
Core Parameters for Core PL-1, in Coastal Salt Water Marsh, Bay Side of Assateague Island, Maryland

Depth (cm)	Lithology	Salinity (‰)	C_s (mS/cm)	C_w (mS/cm)	R_s (Ω m)	R_w (Ω m)	F
0–11	Living marsh, sandy, no H_2S	32.8	9.6	45.7	1.04	0.219	4.8
11–12	Break/transition (sandy) strong H_2S	33		46.2		0.216	
12–27	Dead marsh, peaty-sandy, becoming increasingly clayey, brown to gray-black with depth, strong H_2S	33	10.4	46.2	0.96	0.216	4.4
27–29	Sand mixed with peat, H_2S	33.2	9.9	46.5	1.01	0.215	4.7
29–36	Sand, med to coarse, with few pebbles, H_2S	32.5	8.2	45.3	1.22	0.221	5.5
36–72	Sand, coarse, gray, H_2S	27.9	6.7	39.4	1.49	0.254	5.9
72–137	Sand, coarse, gray, no H_2S	27	6.9	38.2	1.45	0.262	5.5
137–156	Clay, black, H_2S	25	10.5	35.6	0.95	0.281	3.4
156–179	Silty sand, clayey in part, reddish to gray-black; with shell, no H_2S	24	7.5	34.3	1.34	0.292	4.6
179–245	Sand, shells 200-248 cm	23	5.1	33	1.97	0.303	6.5
245–351	Sand gray, coarse	20	4.7	29.1	2.12	0.344	6.2

C_s and C_w refer to calculated conductivity of sediment and water, respectively. Notes as in Table 2.

areas surveyed (Wells et al. 1994a; Wells et al. 1994b; Wells et al. 1996; Wells et al. 1997; Wells et al.1998), but peat in this environment is an excellent horizontal conductor of fluid, whereas clays having similar porosity offer resistance to flow. This underscores the importance of properly establishing sediment lithology for effective hydrologic interpretation. Formation factors increase with depth as sediments consolidate and become cemented. They increased markedly in Pleistocene sediments recovered in the White Neck cores from Indian River Bay.

Submarine Discharge Associations

We observed two types of submarine occurrence of freshened ground water beneath Delmarva coastal bays. The first type is thin, shallow fresh water layers present from < 1 m up to ~3 m beneath the beach-bay interface. This type seems to extend from a few tens of meters to < 500 m offshore. The second type may be tens of meters thick, and may extend as far as 2 km beneath the bays. In the Delaware and northern Maryland coastal bays, this type of ground water forms sharply bounded tongues of fresh to slightly brackish water that are interlayered with brackish and saline waters. Evidence that submarine discharge is active includes (1) nearshore freshening of surface waters in areas of significant subsurface resistivity anomalies, (2) shallow (< 30 cm) vertical salinity transitions to fully fresh water in cores from tidal streams and canals (Indian River and Assawoman Canal) in Indian River Bay, Herring Creek, Rehoboth Bay, and fresh water within 3 m of the sea floor in open-bay areas of Indian River Bay and Chincoteague Bay, (3) sharp lateral boundaries of high-resistivity zones that could not be sustained without continuous fluid advection along permeable pathways, and (4) presence of oxygen and nitrate in fresh ground water distant from shore and beneath sulfidic, salt water-permeated sediments (Bratton et al. this issue).

The shallower type of freshened layer seems to be present along the entire western shore of the Delmarva coastal bays, judging by detailed along-coast salinity measurements, and nearshore and beach core data. The deeper, thicker fresh ground water features are discussed in more detail by Krantz et al. (this issue). Unconsolidated bayfloor muds appear to serve as confining layers for buried permeable peat or coarse sand horizons. Fresh water plumes bayward of tidal creeks are associated with incised valleys typically filled by fluvial or littoral sand and peat overlain by muds. The upper part of the incised valley fill acts as a semiconfining layer to restrict downward flow of salt water and to allow fresh ground water to flow offshore in the permeable basal fill units and older underlying sediments. Other factors that appear to favor submarine flow and discharge include elevated land areas near bay shores, and permeable subsurface features such as paleobeach ridges.

Although pure clays are rare in the bays, lagoonal muds in the central and western parts of Rehoboth Bay and Indian River Bay typically have 20% to 30% clay, and 50% to 60% silt (Kraft and John 1976; Chrzastowski 1986). Similar, fine-grained sediments characterize the western margins of Assawoman, Isle of Wight, and Chincoteague bays (Bartberger 1973; Kerhin et al. 1988; Wells et al. 1994a; Wells et al. 1994b; Wells et al. 1996; Wells et al. 1997; Wells et al. 1998). On the eastern margins of the coastal bays, near the barrier bars, the sediments tend to be sandy, due to accumulated overwash deposits and flood-tidal deltas from inlets. Such sediments evidently do not provide consistent confining beds needed to permit extensive transport of fresh ground water beneath the bay. Another contributing factor may be the limited recharge area of the narrow barrier islands.

No evidence was found in this survey for advective discharge in the central parts of Chincoteague Bay. Gradual freshening of pore water with depth in corehole WN–4, near the center of Indian River Bay, has been suggested to reflect movement of ground water from the north shore of the bay (Krantz et al. this issue). This phenomenon might alternatively be attributable to relict fresh water recharged during either (1) the Pleistocene glacial lowstand or (2) periods in the past where barrier inlets closed and the bay became significantly freshened. Age dating of ground water samples collected closer to shore (Bratton et al. this issue) did not detect ground water beneath the bay older than ~50 yr, but no samples were collected from the WN–4 site.

Zone of Tidal Pumping and Mixing

Although hotspot areas of focused ground water discharge were detected by aerial infrared imaging in Delaware coastal bays (McKenna et al. 2001), the interface between salty bottom water and ground waters appears to generally be mediated by a brackish transition zone of tidal pumping and mixing, with a minimum thickness of ~20 cm. During the investigation in Indian River Bay, in May 2000, temperatures of discharging ground water must have been similar to those of bay waters, because temperature varied in the affected surface waters much less than salinity.

Hypersaline Brines

Hypersaline brines in deep ground waters (usually > 300 m) underlie much of the coastal Atlantic (Manheim and Horn 1968). The presence of shallow brines, however, of the type suggested by low-resistivity anomalies in Line 9 in Indian River Bay at depths > 20 m was not anticipated. Hypersaline ground water was previously encountered at 9 to 30 m depth in boreholes on Assateague Island, Maryland (Dillow et al. 2002). Similar occurrences have also been described on North Carolina's Outer Banks (Bratton et al. 2002). Deeper brines with 22,000 mg/L chlorine are known from 90 m depth around Cape May, New Jersey (Lacombe and Carleton 2002). These features all occur in coastal bay areas of the barrier-bar type. Upconing from deeper hypersaline brines is unlikely because the shallow brines are vertically separated from deep brines by hundreds of meters. The shallowest known formation with hypersaline brine in the Maryland-Delaware area is the Waste Gate Formation of Lower Cretaceous age encountered at a minimum depth of 1050 m (Hansen 1982). Modern brine-forming environments all require either arid climates or at least seasonal conditions that promote rapid evaporation of salt water from enclosed shallow basins. A possible explanation of relict brines is that they formed in evaporative lagoons along the Atlantic Coast in arid conditions during the middle to late Pleistocene. Sea level curves and paleogeographic reconstructions (Belknap and Kraft 1977; Chrzastowski 1986) place the ocean too far away from the current brine sites to supply salt water for evaporation between 25,000 and 7000 yr ago, so this mechanism would only work for older (> 25,000 yr) or relatively young (< 7000 yr) formations. The potential for Pleistocene hypersaline lagoons in the Delmarva region, like the modern Laguna Madre in Texas, remains speculative, but is supported by the presence of ooidal sediments on the continental shelf south of Cape Hatteras, dated at between 22,000 and 29,000 yr before present. These ooids showed $\delta^{18}O$ values that suggested formation in either a low temperature environment at contemporary salinity, or in a hypersaline environment at present-day temperatures (Milliman 1972).

Control of Submarine Flow by Inherited Drainage Geometry

The resistivity profiles and core data show that the pathways for fluid movement beneath the bays are strongly influenced by inherited subsurface geometry of the barrier-bar estuaries examined in this study (Krantz et al. this issue). The low sea level stand drainages (paleodrainages) of the Chincoteague Bay and Indian River/Rehoboth Bay regions are best described as trellised and dendritic, respectively, as shown schematically in Figure 15. Specific illustrations and discussion of dendritic paleodrainage features in Indian River Bay are documented in more detail by Krantz et al. (this issue).

The flooded trellised system beneath Chincoteague Bay contains shallow, shore-parallel paleovalleys separated by beach ridges. As sea level submerges them, buried beach ridges, such as the offshore extension of Sinepuxent Neck, can act as submarine conduits for tongues of fresh water flowing into the bay (Figure 11). Where the shoreline of the estuary occupies an interridge trough, submarine ground water appears to flow offshore in sheets (rather than tongues) perpendicular to the axes of modern barriers or onshore beach ridges. Peats that were deposited in marshes between the ridges are favored channels for horizontal fluid movement. Submarine ground water flow in the submerged dendritic system, however, may be dominated by oblique tongues of fresh water flowing below incised valleys filled with peats and fine-grained sediments.

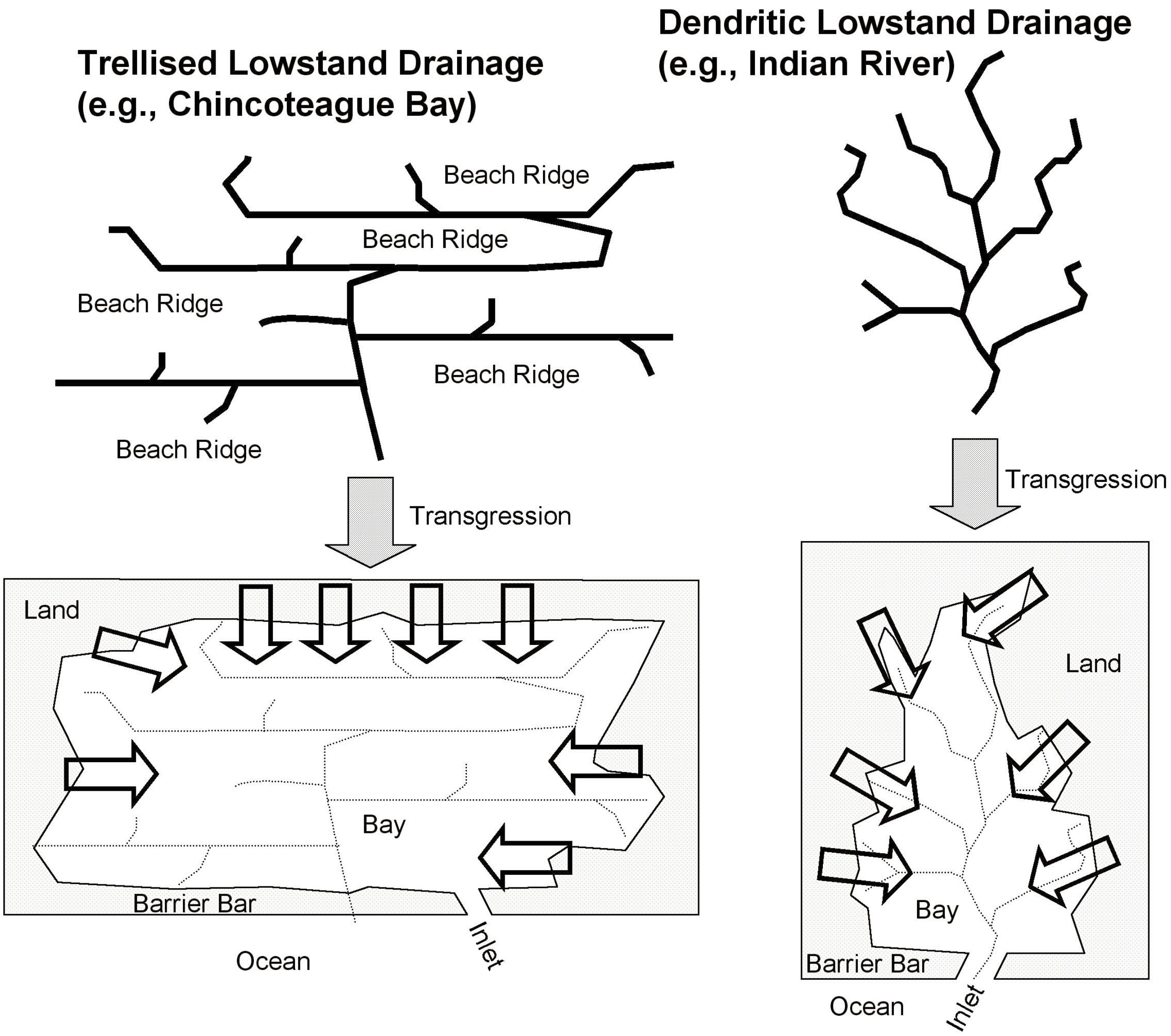

Figure 15. Schematic diagram for paleodrainage patterns in the Delmarva coastal bays and subsequent development of submarine ground water flow (arrows after inundation by rising sea level).

Application of Delmarva Observations to Other Areas

Numerous observations of submarine discharge have been reported along the Atlantic continental margin (Bratton et al. this issue). However, documentation of spatial and vertical distribution of hydrogeological parameters sufficient to track submarine ground water flow pathways, especially on the seaward side of the systems, has been limited. This is also true of the less common phenomenon of saline permeation into surficial aquifers beneath tidal streams and canals. Hagemeyer and Stewart (1991) conducted detailed electromagnetic studies of the Cross-Florida Barge Canal. From horizontal resistivity patterns and borehole logs, they showed a saline wedge extending ~20 m into the subsurface beneath the canal and adjacent areas. Related studies of salt water/fresh water interaction in south Florida have been greatly expanded as a part of the Everglades restoration effort there (Langevin 2000).

A brief review of literature on coring or other studies along the Atlantic Coast was performed, to see where relationships observed for the Delmarva coastal bays might predict submarine ground water occurrence. This review was focused on small, shallow features such as those observed in the Delmarva coastal bays, rather than the large (> 50 m deep) Pleistocene paleochannels described from Chesapeake Bay and the southern part of the Delmarva Peninsula (Colman et al. 1990). The hydrology of the southern Delmarva Peninsula has also been influenced by a giant impact crater of Eocene age, which has been actively investigated during the past several years (Powars et al. 2001; Nowroozi et al. 2004), and which exerts a major influence on deep ground water flow and salinity in the region.

Detailed interstitial water analyses from suites of cores collected in Chesapeake Bay tidal tributaries, such as the James, Patapsco, Ware, and Elizabeth rivers, showed reduced salinities with depth (Hill et al. 1985). These patterns are consistent with submarine discharge conditions analogous to those in the tidal streams and coastal bays of the Delmarva Peninsula, particularly the upstream portion of Indian River.

Fresh water anomalies have been measured in open-water cores from both Chesapeake Bay (Bricker et al. 1977; Hill et al. 1985) and the Potomac River estuary (Goodwin et al. 1984). In many of these cores, interstitial waters just below the sediment-water interface were fresher than ambient bottom waters, but became saltier with depth, trending toward constant values below depths of 16 to 50 cm. This contrasts with the relatively constant salinities (19‰ to 20‰) observed in cores collected in open-water areas between the Patuxent River and Annapolis, Maryland, during 2000 (Bratton unpublished). These results may be linked with extensive fresh water flooding and deposition of turbid muds caused by Hurricane Agnes in June 1972 (Nie et al. 2001). Sampling by Bricker et al. (1977) bracketed the period before and after the hurricane, and posthurricane cores showed the dramatic changes. Hill et al. (1985) and Goodwin et al. (1984) collected pore water samples in the late 1970s and early 1980s. It appears that freshened pore fluids in muds deposited during the hurricane were not fully replaced by salty bay waters ~8 yr later, whereas 28 yr later, most evidence of the hurricane's influence had dissipated. The influence of three hurricanes that severely impacted North Carolina's estuaries in 1999 may also complicate investigations of submarine ground water discharge in those environments (Paerl et al. 2000).

Submarine fresh water influence a few hundred meters offshore has been observed in Biscayne Bay, Florida (Langevin 2000), and in Hillsborough Bay (Tampa Bay), Florida, by resistivity surveys using the same techniques reported here (Swarzenski and Meunier 2003; Swarzenski et al. 2004), and confirmed by core studies. Freshened subsurface pore waters attributed to submarine discharge were also found in vibracores from Bogue Sound, North Carolina, up to 1 km from shore (Latterman 1997), and in Buzzards Bay, Massachusetts (McCobb and LeBlanc 2002). In summary, shallow submarine flow with significant lateral extent and large discharge volume may be a widespread phenomenon in estuarine and coastal sediments.

Conclusions

Submarine discharge is widespread on the landward margins of the Delmarva coastal bays. Fine-grained sediments on the sea floor serve as confining beds for movement of fresh ground waters from the unconfined aquifers on land through peat or permeable sands. Submarine fresh water may extend more than 1 km into the bays. Fresh ground water is intermittent or absent adjacent to the eastern (seaward) margins of the bays where sandy overwash sediments dominate.

Tidal streams in the Delmarva coastal bays do not normally serve as sources of vertical infiltration of salt water that permeates downward into the surficial aquifer. Rather, small streams may be the shoreward continuation of larger submerged paleodrainage systems that act as conduits for focused flow and discharge of submarine fresh water into the bays.

The data reported here show the importance of the hydrogeology of the seaward side of estuaries and coastal regions in controlling submarine discharge phenomena, including the high degree of anisotropy in hydraulic conductance that may be present in unconsolidated sediments.

Streamer resistivity systems are effective tools for defining fresh water distribution in coastal environments. They deliver continuous profiles in very shallow water (< 1 m) conditions, collect data rapidly (i.e., up to 30 times faster than the rate for comparable on-land resistivity studies), and lend themselves to systematic interpretive output in standard formats. Coring studies using a variety of ancillary hydrologic measurements are important adjuncts to resistivity surveys of ground water conditions beneath bays and coastal waters.

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