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Quantifying groundwater discharge through fringing wetlands to estuaries: Seasonal variability, methods comparison, and implications for wetland–estuary exchange

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Abstract

Because groundwater discharge along coastal shorelines is often concentrated in zones inhabited by fringing wetlands, accurately estimating discharge is essential for understanding its effect on the function and maintenance of these ecosystems. Most previous estimates of groundwater discharge to coastal wetlands have been temporally limited and have used only a single approach to estimate discharge. Furthermore, groundwater input has not been considered as a major mechanism controlling pore-water flushing. We estimated seasonally varying groundwater discharge into a fringing estuarine wetland using three independent methods (Darcy's Law, salt balance, and Br^- tracer). Seasonal patterns of discharge predicted by both Darcy's Law and the salt balance yielded similar seasonal patterns with discharge maxima and minima in spring and early fall, respectively. They differed, however, in the estimated magnitude of discharge by two- to fourfold in spring and by 10-fold in fall. Darcy estimates of mean discharge ranged between -8.0 and 80 L m^{-2} d^{-1}, whereas the salt balance predicted groundwater discharge of 0.6 to 22 L m^{-2} d^{-1}. Results from the Br^- tracer experiment estimated discharge at 16 L m^{-2} d^{-1}, or nearly equal to the salt balance estimate at that time. Based upon the tracer test, pore-water conductivity profiles, and error estimates for the Darcy and salt balance approaches, we concluded that the salt balance provided a more certain estimate of groundwater discharge at high flow (spring). In contrast, the Darcy method provided a more reliable estimate during low flow (fall). Groundwater flushing of pore water in the spring exported solutes to the estuary at rates similar to tidally driven surface exchange seen in previous studies. Based on pore-water turnover times, the groundwater-driven flux of dissolved organic carbon (DOC), dissolved organic nitrogen (DON), and NH_4^+ to the estuary was 11.9, 1.6, and 1.3 g C or g N m^{-2} wetland for the 90 d encompassing peak spring discharge. Groundwater-induced flushing of the wetland subsurface therefore represents an important mechanism by which narrow fringing marshes may seasonally relieve salt stress and export material to adjacent water masses.

Identifying the hydrological factors affecting chemical fluxes within and through wetlands is critical to understanding wetland function and maintenance within the landscape. Intertidal wetlands have been suggested as transformers and potential regulators of nutrient fluxes with nearby coastal systems (Jordan et al. 1983; Childers 1993). Furthermore, the rates of solute fluxes and solute flushing within wetlands is essential in maintaining the high levels of primary productivity characteristic of the ecosystem (Bradley and Morris 1991; Osgood and Zieman 1998). The intertidal and nearshore subtidal zones tend to be zones of maximal groundwater discharge along coastal margins (Reilly and Goodman 1985), and fringing wetlands commonly inhabit much of the temperate intertidal shoreline. Consequently, the importance of groundwater discharge through fringing wetlands in mediating biogeochemical exchange with estuaries has been argued (Harvey and Odum 1990). Although chemical fluxes between the groundwater aquifer, wetlands, and the adjacent estuary are hydrologically mediated, all components of the wetland water balance have not typically been well characterized (Yelverton and Hackney 1986; Whiting and Childers 1989). Therefore, uncertainties in wetland chemical flux budgets, particularly those influenced by groundwater discharge, arise in a large part because of uncertainties in the water balance (La Baugh 1986). These uncertainties may be compounded by biases inherent in the different methods used to estimate hydrologic fluxes.

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Where groundwater inputs to wetlands are large, accurately quantifying the discharge may be critical to understanding internal nutrient cycling (Tobias 1999), to assessing the importance of wetland interception of watershed nutrients (Valiela et al. 1978; Bohlke and Denver 1995; Howes et al. 1996; Portnoy et al. 1998), and to nutrient fluxes out of the wetland. With few exceptions (e.g., Morris 1995), most previous estimates of groundwater discharge to coastal wetlands have been limited to the macrophyte growing season and have used a single approach to estimate discharge. Similarly, most wetland solute flux studies (Jordan et al. 1983; Childers 1994) have not considered groundwater input as a potential mechanism controlling solute release from wetland sediments.

Whole wetland-scale estimates of discharge using tidal salt balances have been made in systems that possess a single well-defined tidal conduit (Valiela et al. 1978; Howes et al. 1996). Calculation of discharge through fringing wetlands that lack a well-defined tidal conduit (e.g., tidal creek), however, is more difficult. The two approaches most often used to estimate groundwater discharge into these types of wetlands are based on measurements of hydraulic gradient and hydraulic conductivity in sediments (referred to here as the Darcy calculation approach) or on mass balances of water or solutes in sediments. Calculating a water flux using Darcy's Law often assumes homogeneity in sediment properties and steady flow conditions. Errors include those that arise from sediment heterogeneity (e.g., macropore flow) and from the measurement of hydraulic gradients caused by the tidally fluctuating heads encountered in most intertidal wetlands (Winter 1981; Hemond and Fifield 1982; Nuttle and Harvey 1995). Models that solve time-varying equations of groundwater flow are useful where tidal fluctuations cause rapid changes in the magnitude and direction of groundwater fluxes (Harvey et al. 1987). However, the resulting groundwater discharge estimates may still contain large errors due to problems such as accurately estimating the average sediment hydraulic conductivity (Nuttle and Harvey 1995).

Mass balance methods are thought to have the advantage that measurement errors are typically smaller than errors in estimating hydraulic conductivity. Water and solute mass balances sometimes yield groundwater discharge estimates whose total error is less than that of Darcy calculations (Gehrels and Mulamoottil 1990; Nuttle and Harvey 1995; Hunt et al. 1996). A greater number of measurements, however, are required. Uncertainties in individual terms of the water balance propagate through the calculation and increase the uncertainty of the groundwater flux estimate when it is calculated by difference (LaBaugh 1985). The combined mass and water balance approach shows the greatest utility when the sediment pore water is well-mixed, solute concentrations differ substantially between groundwater and surface water, the number of inputs and outflows to the sediment is limited, and steady state assumptions can be made (Harvey et al. 1995; Morris 1995, Hunt et al. 1996). Still, in some wetlands, one term of the water balance often requires a Darcy flux calculation or a calculation by difference in a mass balance (Harvey and Odum 1990). Although the Darcy and solute balance approaches have different sources of error and Darcy calculations cannot account for groundwater/estuarine water mixing prior to discharge, both approaches should yield similar discharge estimates when a fresh groundwater end member is assumed. To our knowledge, the methods have not been compared on an annual basis. A coupled water and solute (salt) mass balance approach (heretofore referred to as the "salt balance" approach) has been used to estimate hydrologic fluxes in lakes (LaBaugh et al. 1997) but has not previously been used to estimate transient and seasonal groundwater fluxes through fringing intertidal wetlands.

The purpose and scope of this study was to (1) describe seasonal patterns of groundwater discharge to an intertidal fringing wetland (2) determine differences in discharge estimates resulting from different methods used to estimate the groundwater flux, and (3) define the magnitude of pore-water solute transport out of the wetland caused by groundwater discharge.

We used three independent methods to estimate the groundwater flux into a mesohaline marsh near the upland/marsh border. Darcy calculations and a salt balance approach were used to provide estimates of groundwater discharge over an annual cycle, and a conservative tracer injection was used to empirically estimate the discharge rate during a period of high flow. The estimates of groundwater discharge were combined with pore-water chemistry data from a previous study (Tobias 1999) to estimate the rate of solute export caused by groundwater input.

Methods

Site description—The Ringfield study site is located in the Colonial National Historical Park (37°16′42″N, 76°35′16″W), at the confluence of King Creek and the York River in southeastern Virginia (Fig. 1). The steep (1 : 1) forested upland slope transitions into a 25-m-wide wetland composed of a mixed community of *Spartina cynosuroides* and *Spartina alterniflora* (short form). Upland geology near the site is discussed in Libelo et al. (1991). The small-scale marsh stratigraphy consists of the upper 30–80 cm of sandy marsh peat underlain by a semicontinuous layer (10–20 cm thick) of lower permeability glauconitic silty sand. Below 150–200 cm the glauconitic deposits grade into cleaner oxidized iron-rich sands and shell hash of pre-Holocene origin. The study area borders the mesohaline portion of the York River (salinity range 12–21 ppt) and experiences a 1-m tidal range. Dissolved inorganic nitrogen concentrations in the shallow aquifer are typically <1–2 μM. Groundwater inputs of nitrogen to the wetland were therefore considered negligible.

Research instrumentation consisted of one upland water table well and four parallel transects of multilevel piezometers extending perpendicularly from the upland marsh border 10 m out into the marsh (Fig. 1). The multilevel piezometers were arranged into clusters of four to five, with the depths at the base of the screens ranging from 50–250 cm in 50-cm intervals (Fig. 1). Piezometer construction and installation is described in Tobias (1999). Briefly, the water table well and wetland piezometers were installed by drilling holes with a hand auger and placing 2.54-cm diameter PVC pipe fitted with a 0.025-cm PVC slot screen at the end into the borehole. The annular space around the screen was filled

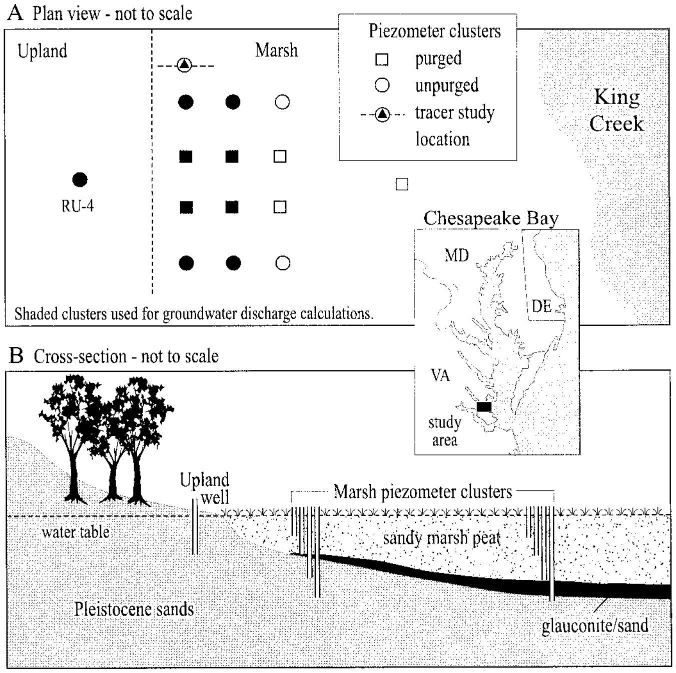


Fig. 1. Site location and schematic showing the multilevel piezometer grid. Each piezometer cluster consisted of four to five sampling depths, as illustrated schematically in panel B. Piezometer clusters used in the groundwater discharge calculations appear as shaded symbols. Clusters denoted by circles were sampled for water levels and salinity and the shaded circles were used in the Darcy calculations. Clusters denoted by squares were sampled for salinity only and, with the shaded circles, used for the salt balance calculations. The icon for the tracer study location in panel A denotes a piezometer network illustrated in detail in Fig. 6A.

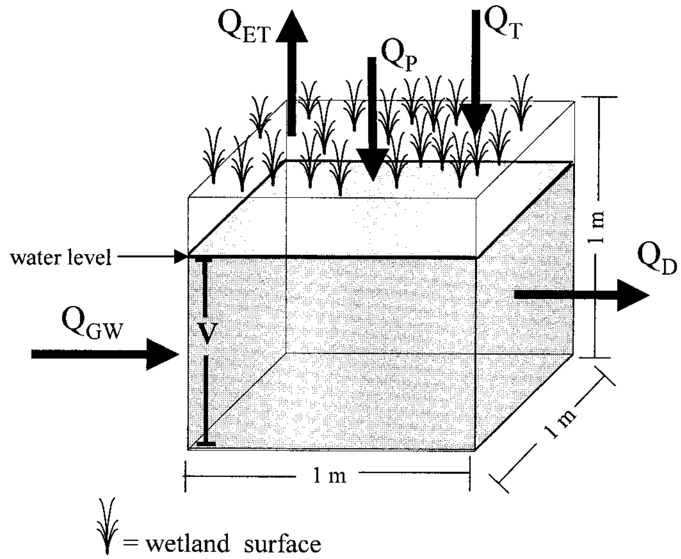


Fig. 2. Control volume and water balance terms used in the salt balance method. Both horizontal and vertical fluxes of fresh groundwater are included in the Q_{GW} term.

with clean sand and above the screen with bentonite and auger cuttings.

Methods of estimating groundwater discharge—All methods estimated groundwater discharge into the upper 1 m of marsh sediment (the zone of fastest biogeochemical cycling; Tobias 1999) within 2 m of the upland forest border. A sediment control volume of 1 m^3 was chosen for cross-method comparison. The control volume is not the volume of the entire wetland but rather the volume of a representative 1 m^2 area of wetland surface to a depth of 1 m located within 2 m of the upland border (Fig. 2). The Darcy and salt balance estimates of groundwater flux were derived from spatially averaged measurements within a month from the eight piezometer clusters (two per transect) nearest the upland border. Discharge estimates represent average discharge found within an ~10-m^2 marsh area.

Darcy calculations—Average horizontal and vertical groundwater discharge into the marsh in each month were calculated from hydraulic heads and sediment hydraulic conductivities according to

$$q = -K_{\mathrm{h,v}}\frac{dh}{dl}, \tag{1}$$

where q is the specific discharge of groundwater, $K_{\mathrm{h,v}}$ is the average hydraulic conductivity (vertical [K_{v}] or horizontal [K_{h}]) of the saturated sediment between piezometers, h is the average hydraulic head measured in the piezometers, and l is the linear distance between the midpoints of the piezometer screens. Hydraulic conductivity of the marsh sediment, the basal marsh deposits, and the shallow aquifer was determined in 1996 from slug tests (Hvorslev 1951) performed in all piezometers. Head measurements used to calculate monthly discharge were measured using a water level meter at slack high and slack low tides on the peak spring and neap tides (eight measurements per month). Water levels were referenced to a common datum (the top of the well casing for well RU-4). Although head measurements at the site prior to study indicated variability in water level coincident with tides and precipitation, the temporal sampling scheme (spring/neap, high tide/low tide) was designed to encompass the largest fluctuations in water levels within a given month. Horizontal discharge was calculated from average head gradients and K_{h} values between the upland well (RU-4) and each of the eight piezometers (50 and 100 cm deep) located within 2 m of the upland border. Individual calculations for the 50- and 100-cm horizons were averaged. Vertical discharge was calculated using average vertical gradients and K_{v} values from piezometers screened at 2 and 1 m deep. Because $K_{\mathrm{h}}:K_{\mathrm{v}}$ ratios of 1–30 have been reported for coastal aquifers with a similar average K_{h} as seen at the Ringfield site, K_{v} was calculated from the K_{h} measurements at $0.1 \times K_{\mathrm{h}}$ (Barlow 1994; Heidari and Moench 1997). The sum of the horizontal and vertical discharge flux described above defined total groundwater flow into the control volume.

A range of error in the monthly Darcy-derived groundwater flux was estimated as the difference between maximum and minimum discharge estimates prior to spatial averaging. This range may not represent biases in measuring $K_{\mathrm{h,v}}$ that could be inherent in using the slug test method and could therefore be an underestimate in total error. The seasonal pattern of discharge was developed by repeating monthly flux calculations for 21 months from January 1997 through August 1998. Connecting an average sequence of steady state calculations was valid because there was essentially no change in water storage within the wetland on a monthly timescale.

Subsurface salt balance—A mass balance model for water and salt (salt balance) was used to estimate the flux of fresh groundwater into the same control volume as defined in the Darcy estimates. Conservation of water in the wetland sediment is expressed as

$$\frac{dV}{dt} = Q_{GW} + Q_P + Q_T - Q_{ET} - Q_D, \quad (2)$$

where dV/dt is the change in water storage volume over a month, Q_{GW}, Q_P, *and* Q_T are the monthly mean inputs of groundwater, precipitation, and tidal infiltration flux, respectively. Q_{ET} and Q_D are the mean monthly export fluxes of evapotranspiration and drainage, respectively (Fig. 2). Tidally driven subsurface fluxes of water were considered negligible components of the water balance because of the location of the control volume far from the creek bank (Harvey et al. 1987). Because Q_D, in addition to Q_{GW}, is not known in Eq. 2, an equation describing conservation of salt in the wetland was combined with Eq. 2 in order to solve for Q_{GW}. Conservation of salt in the wetland sediment is expressed as

$$\frac{dS}{dt} = Q_{GW}C_{GW} + Q_PC_P + Q_TC_T - Q_{ET}C_{ET} - Q_DC_W, \quad (3)$$

where dS/dt is the rate of change in salt mass in the control volume ($S = C_W \cdot V$) within a month and C_{GW}, C_P, C_T, C_{ET}, and C_W are the monthly mean salt concentrations (g salt L^{-1}) associated with the fluxes Q_{GW}, Q_P, Q_T, Q_{ET}, and Q_D described above. Salt excretion by plants was not considered in the salt balance because Morris (1995) demonstrated that it accounted for a small percentage of daily salt removal. Rearranging Eq. 2 and substituting into Eq. 3 yields the combined equation for conservation of water and salt (Eq. 4).

$$\frac{dS}{dt} = Q_{GW}C_{GW} + Q_PC_P + Q_TC_T - Q_{ET}C_{ET} - \left(Q_{GW} + Q_P + Q_T - Q_{ET} - \frac{dV}{dt}\right)C_W \quad (4)$$

Solving Eq. 4 for Q_{GW} and simplifying by assuming that C_{ET}, C_P, and C_{GW} possess negligibly low salt concentrations yields the governing equation for the salt balance model (Eq. 5).

$$Q_{GW} = \frac{\left(-Q_TC_T + Q_PC_W + Q_TC_W - Q_{ET}C_W - \frac{dV}{dt}C_W + \frac{dS}{dt}\right)}{-C_W} \quad (5)$$

This equation is similar to that of Krabbenhoft et al. (1990), where it was applied to steady state water and isotope mass balances in lakes.

Definition of salt balance terms—The equation used to compute the monthly infiltration of estuarine water into sediment (Q_T) during high tide follows the water balance presented in Harvey and Odum (1990).

$$Q_T = N(Z_{SED} - h_{MIN})S_y, \quad (6)$$

where (N) is the frequency of inundation (number of times the site is flooded per month), $Z_{SED} - h_{MIN}$ is the difference between the elevation of the sediment surface and the average minimum head (water level) beneath the marsh during each tidal cycle, and S_y is the average specific yield (0.12) of marsh sediment as determined on duplicate cores according to Harvey et al. (1995). Q_T was computed for each piezometer cluster and averaged. Because the site is flooded irregularly, N was determined by summing the number of tidal events per month whose maximum tidal height exceeded the sediment elevation at each well cluster. Tidal elevations recorded at the Gloucester Point, Virginia, York River NOAA tide gauge were used to determine N after groundtruthing the site for specific flooding events.

Monthly evapotranspiration (Q_{ET}) of water out of the control volume was assumed to be equal to the average monthly potential evapotranspiration rate (P_{ET}) derived from air temperature and daylength (Hamon 1961).

$$P_{ET} = \frac{[0.021(H_t^2 \cdot e_{sat})]}{(T_t + 273)} \quad (7)$$

H_t is the average number of hours of daylight per day in the month, T_t is the monthly average air temperature (°C), and e_{sat} is the relative humidity estimated from air temperature as defined by Bosen (1960).

$$e_{sat} = 33.86[(0.00738T_t + 0.8072)^8 - 0.000019(1.8T_t + 48) + 0.001316] \quad (8)$$

Monthly Q_{ET} was normalized to a per day rate (L m^{-2} d^{-1}) for input to the model. Air temperature from Newport News, Virginia (located ~10 km from the site), was used to estimate P_{ET}.

Monthly precipitation inputs to the control volume (Q_P) were estimated from meteorological data collected at the Virginia Institute of Marine Science located ~9 km from the study site according to

$$Q_P = P - Px, \quad (9)$$

where (P) is the total monthly precipitation, and (x) is the fraction of the total rainfall that fell on the marsh when NOAA tidal records predicted that the marsh was flooded.

The average monthly salt concentration in pore water in the control volume (C_W) was estimated by averaging measurements in the eight paired clusters of piezometers. Each paired cluster consisted of a 50- and 100-cm piezometer. Four clusters were mixed with a polyethylene sampling tube and sampled with a syringe 4–8 times per month but were not purged prior to sampling. Later testing determined that unpurged piezometers provided questionable estimates of salinity because water and salt were stored in the body of the piezometer above the screen. The remaining four clusters were purged with a peristaltic pump prior to sampling and were sampled once every 2–3 months. Following filtration of pore water, salinity was determined using a refractometer or salinity meter for the purged and unpurged samples, respectively. For the purged piezometers, salinity estimates for missing months were determined by linear interpolation. We attempted to minimize the effect of well storage artifacts on measured salinity in the nonpurged wells by averaging the salinity estimates from the purged piezometers with salinity values recorded at the nonpurged wells for those months.

The rate of change of water storage in the control volume (dV/dt) within each month was calculated from the equation

$$\left(\frac{dV}{dt}\right)_i = \left(\frac{\Delta V}{\Delta t}\right)_i = \frac{\Delta h \cdot A \cdot S_y}{\Delta t}, \tag{10}$$

where δh is the average difference between hydraulic heads in the piezometers measured on or near the first and last days of month i, A is the surface area of the control volume (1 m^2), S_y is the specific yield, and δt is the number of days between the head measurements. The rate of change in salt mass in the control volume (dS/dt) was calculated from the equation

$$\left(\frac{dS}{dt}\right)_i = \left(\frac{\Delta S}{\Delta t}\right)_i = \frac{(S_f - S_0) \cdot V}{\Delta t}, \tag{11}$$

where S_0 and S_f are the salt concentrations (g salt L^{-1}) measured on or near the first and last days of month i, respectively; V is the mean control volume (liters); and δt is the number of days between salt measurements. Salt concentrations were stable between tides and adjacent days but changed on the scale of months. The mean volume of water in the control volume is defined by

$$V = [1 - (Z_{\mathrm{SED}} - h_{\mathrm{MIN}}) \cdot A, \tag{12}$$

where the sediment thickness of the control volume is assumed to be 1 m deep.

The estimate of error in the salt balance calculated groundwater flux and its sensitivity to variation in input parameters were assessed independently through use of a Monte Carlo simulation and a sensitivity analysis, respectively. Monte Carlo simulation of the salt balance was achieved by simultaneously manipulating the input values of $Z_{\mathrm{SED}} - h_{\mathrm{MIN}}$, C_{W}, T_t, and P within one standard deviation assuming a normal distribution for each parameter. The assignment of individual monthly standard deviations for $Z_{\mathrm{SED}} - h_{\mathrm{MIN}}$, C_W, and T_t used in the simulation were calculated from head and salinity measurements from all piezometers within the control volume within a month, and from fluctuations in the daily mean temperature within a month, respectively. The standard deviations therefore represent both spatial variability and temporal variation within a month. Variance in P was estimated from a maximal 75% error in accounting for localized precipitation events (Winter 1981). Manipulating values of $Z_{\mathrm{SED}} - h_{\mathrm{MIN}}$, C_{W}, T_t, and P in the Monte Carlo simulation provided an estimated standard deviation in Q_{GW} while permitting covariance among different components of error that are introduced by the measured input flux terms.

The sensitivity of the salt balance estimated Q_{GW} to individual variations in Q_{T}, C_{W}, Q_{P}, Q_{ET}, and dV/dt was determined by increasing and decreasing each input term separately by two of its standard deviations within each month. The magnitudes of the individual monthly standard deviations used in the sensitivity analysis were determined according to the procedure described for the Monte Carlo simulation. Sensitivity to changes in dS/dt was not assessed specifically, but its effect on Q_{GW} was inherent when varying C_{W}.

Tracer study—To empirically determine the groundwater discharge velocity into the wetland, a small-scale tracer release was performed at a period of high groundwater discharge in March 1998. The tracer experiment protocol was 1.0–1.5 M potassium bromide (KBr) solution injected as a single slug into two adjacent injection wells (5.08 cm PVC) located at the upland marsh border; injection wells pumped dry following insertion of a PVC liner, the well volume (now dry) filled with 1 liter of injectate, and the liner removed. This technique allowed for near total replacement of well volume with injectate in <10 min while increasing hydraulic head by <5%. The injection and target wells were screened from 10 to 50 cm below the marsh surface. The target well array and all the injection wells were sampled 10 times for a 63-d period. Br^- concentrations in samples were measured in the laboratory using an Orion 94-35 Br^- specific electrode following temperature equilibration. Bromide breakthrough curves were generated for the target piezometers. The discharge flux (Q_{GW}) was determined from the three piezometers (No. 2, 3, 4), which contained the highest maximum bromide concentrations according to

$$Q_{\mathrm{GW}} = (L/t)A \cdot n, \tag{13}$$

where L is the distance between the injection point and the target well(s) along the center of mass, t is the elapsed time from the injection start until the peak of the bromide breakthrough curve, A is the cross-sectional area (defined as 1 m^2), and n is the average sediment porosity (0.56) between 10 and 50 cm deep. The porosity was defined as the ratio of void volume to total volume determined from replicate ($n = 4$) saturated sediment cores collected in 1996 and dried to a constant weight (Freeze and Cherry 1979). Error for the March tracer-derived groundwater flux was based on averaging transport times for the three piezometers.

Results

Darcy estimates—Horizontal (K_{h}) and vertical (K_{v}) hydraulic conductivities for the upland and shallow wetland strata used in the Darcy discharge estimates ranged between 2 and 17×10^{-4} cm s^{-1} (Table 1). The fairly constrained range of K_{h} supports the averaging of hydraulic conductivity values between the upland and marsh piezometers used in the calculations. The hydraulic conductivity of the basal marsh deposits was generally lower, ranging from 0.4 to 5.0×10^{-4} cm s^{-1}, and the conductivity of the oxidized sands and shell hash composing the underlying aquifer was similar to that encountered in the upper 1 m of marsh strata (1.8–12.0×10^{-4} cm s^{-1}).

The groundwater flux estimates derived from the Darcy calculation followed a seasonal pattern with peak discharge in early spring near the end of April 1997 and April 1998 (Fig. 3A). The groundwater flow minimum was encountered

Table 1. Distribution of hydraulic conductivity in the wetland sediment control volume and nearest upland well at the Ringfield Site. K_h and K_v denote estimated horizontal and vertical conductivity, respectively. Depths are below the marsh sediment surface.

Piezometer No.	Location	Depth (cm)	Conductivity (10^{-4} cm s^{-1}) K_h	K_v
RU-4	Upland	0–50*	6.5	0.7
1-1	Marsh	0–50	14.0	1.4
1-6	Marsh	0–50	11.4	1.1
2-4	Marsh	0–50	14.2	1.4
2-8	Marsh	0–50	17.3	1.7
1-2	Marsh	50–100	5.1	0.5
1-5	Marsh	50–100	2.0	0.2
2-3	Marsh	50–100	2.7	0.3
2-7	Marsh	50–100	6.7	0.7

* Depth below water table in October 1996.

in early autumn near the end of September 1997 and was accompanied by 10–20 L m^{-2} d^{-1} flow reversal into the shallow aquifer, primarily in the landward direction (Figs. 3A, 5). The coefficient of variation (C.V.) of the individual piezometer vertical and horizontal discharge estimates were ~40–100% higher, respectively, during low flow from September to December. However, because the total fall discharge was much lower than spring discharge, the larger variance among individual piezometers yielded a smaller range of total discharge error when compared to the high flow period (Fig. 7). Horizontal fluxes dominated the groundwater discharge during high flow periods in the spring by ~5 : 1 but were similar to vertical fluxes during periods of low discharge in the fall. Head data were not normalized to freshwater prior to calculation of the discharge fluxes. Given the observed salinities in piezometers used to calculate discharge, density correction of hydraulic head would decrease the horizontal flux by a maximum of 2% and the vertical flux by <1%.

Salt balance and tracer estimates—The annual pattern of subsurface salinity mimicked the seasonal fluctuations in river salinity, except that subsurface salinities were lower relative to river water by ~30–75% (Fig. 4). Greater variations in salinity were encountered in piezometers that were purged before sampling. The 50- and 100-cm piezometers possessed a ratio of screen length to submerged length of the piezometer of 0.33 and 0.25, respectively. Piezometer dead space above the screen may have therefore dampened the response of unpurged piezometers to changing salinities surrounding the screen. In spring, the flushing of salt from the wetland subsurface extended beyond the 2 m closest to the upland (Fig. 5), and decreased specific conductivity could be detected in piezometers farthest from the upland (e.g., closest to King Creek).

Table 2 shows the different components of the water and salt balance. Groundwater and tidal infiltration are the dominant inputs in spring and autumn, respectively. Drainage exceeded water loss via evapotranspiration for all months except August.

Figure 3B shows the estimates of fresh groundwater discharge derived from the water–salt balance model. A peak discharge of 22.5 L m^{-2} d^{-1} occurred in March 1998, and a minimum discharge of 0.6 L m^{-2} d^{-1} occurred in August 1997. The annual average and (range) of the monthly coefficients of variation for each of the terms was Q_T = 59% (22–157%), dV/dt = 84% (25–151%), C_W = 13.5% (2.1–

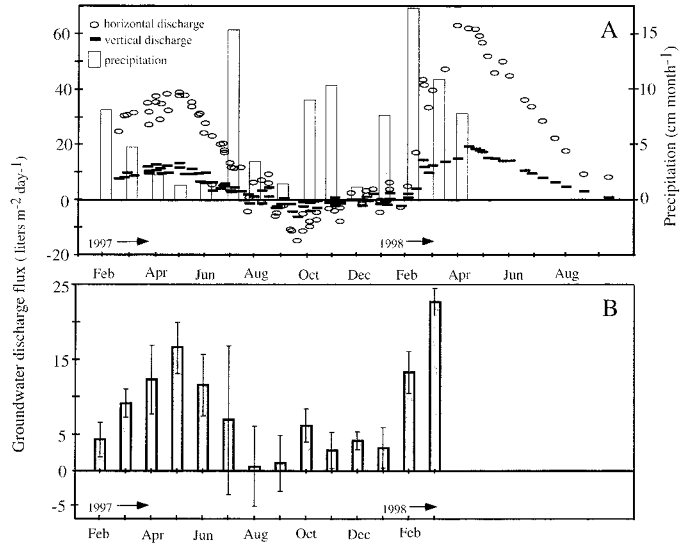


Fig. 3. (A) Horizontal and vertical groundwater discharge estimates derived from Darcy calculations and monthly total precipitation. Negative discharge values denote flow into the aquifer from the marsh. For the vertical and horizontal components, the mean standard deviation for discharge estimates during high flow was 82% (vertical) and 38% (horizontal) of the mean discharge. During low flow, the mean standard deviation for vertical discharge was 140% (vertical) and 86% (horizontal) of mean discharge. Precipitation data collected 9 km from the site at the Virginia Institute of Marine Science. (B) Groundwater flux estimate derived from the salt balance model. Error bars are standard deviations estimated from Monte Carlo simulations (n = 50).

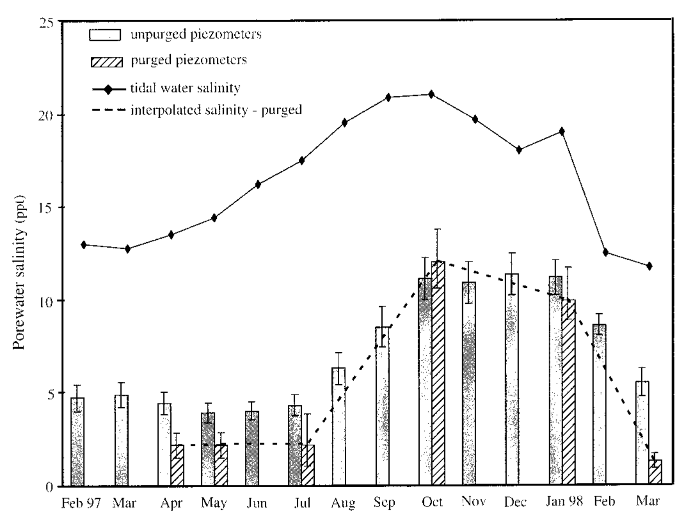


Fig. 4. Average monthly pore-water salinity in the upper 1 m of marsh sediment (purged and unpurged piezometers) and of tidal flooding water. Error bars are standard error.

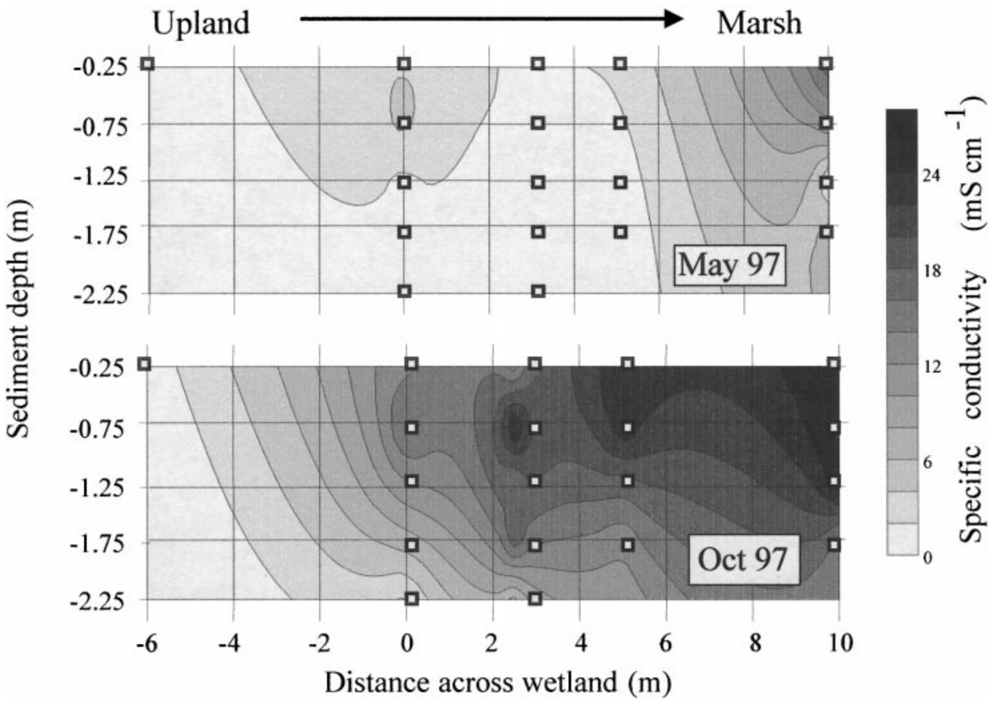


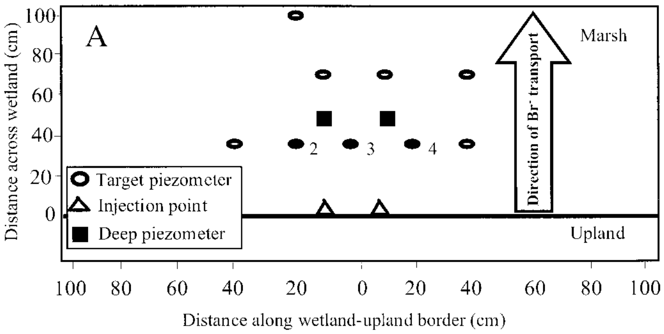
Fig. 5. Distribution of subsurface conductivity at periods of high (May 1997) and low (October 1997) groundwater flow. Positive distance is into the wetland. Distance is equal to zero at the upland border. Sampling point locations (squares) indicate the depth of the center of the 50-cm piezometer screen. The conductivity of tidal water was 23.2 and 34.6 mS cm^{-1} for May and October, respectively.

23%), and Q_{ET} = 40% (10–73%). On the basis of previous error analysis in wetlands (Winter 1981), a coefficient of variation for Q_P of 75% was used in error analysis. The error bars determined by a Monte Carlo simulation indicated the smallest coefficient of variation (C.V. = 8%) in the groundwater flux during high discharge in April 1997. During low flow in August and September the coefficient of variation was nearly 10-fold larger than mean discharge. Results from the sensitivity analysis indicated that the model was most sensitive to changes in any parameter in late summer (July–September). Although Q_P, dV/dt, Q_T, Q_{ET}, and C_W were the terms with the highest individual standard deviations relative

Table 2. Average monthly water fluxes. Groundwater was calculated from the salt balance derived from Eq. 5.

	Water flux (L m^{-2} d^{-1})					
Month	Tidal infiltration (Q_T)	Precipitation (Q_P)	Evapo-transpiration (Q_{ET})	Groundwater (Q_{GW})	Pore-water drainage (Q_D)	% groundwater of all inputs*
1997						
Feb	2.7	3.3	0.9	4.2	9.3	41
Mar	2.1	1.9	1.4	9.0	11.6	69
Apr	3.5	0.8	1.8	12.2	14.7	74
May	2.6	0.4	2.8	16.5	16.7	85
Jun	4.2	0.6	4.2	11.5	12.1	70
Jul	4.3	6.0	5.0	6.8	12.1	40
Aug	4.4	1.2	4.0	0.6	2.2	10
Sep	7.3	0.7	2.8	1.0	6.2	11
Oct	9.8	3.3	1.8	6.0	17.3	31
Nov	8.9	4.0	0.9	2.7	14.7	17
Dec	5.1	0.0	0.6	4.0	8.5	44
1998						
Jan	5.8	2.8	0.7	3.1	11.0	26
Feb	1.7	2.2	0.9	13.1	16.1	77
Mar	1.7	2.2	1.4	22.6	25.1	85

* Percent groundwater was calculated from $[Q_{GW}/\Sigma\ (Q_T + Q_P + Q_{GW})] \times 100$.



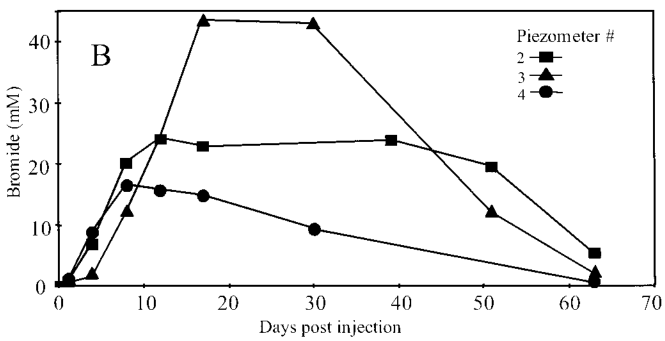


Fig. 6. (A) Injection and target piezometer array used in the March 1998 tracer test. Distance across the wetland is equal to zero at the upland border. The numbered, solid ellipses denote target piezometers with the highest observed bromide concentration. Squares denote deeper (70–100 cm below the surface) piezometers. No bromide concentrations above background were detected in the deeper piezometers for the first 60 d after the release. (B) Bromide breakthrough curves for piezometers (No. 2, 3, and 4) used to calculate groundwater velocity and discharge for the tracer test.

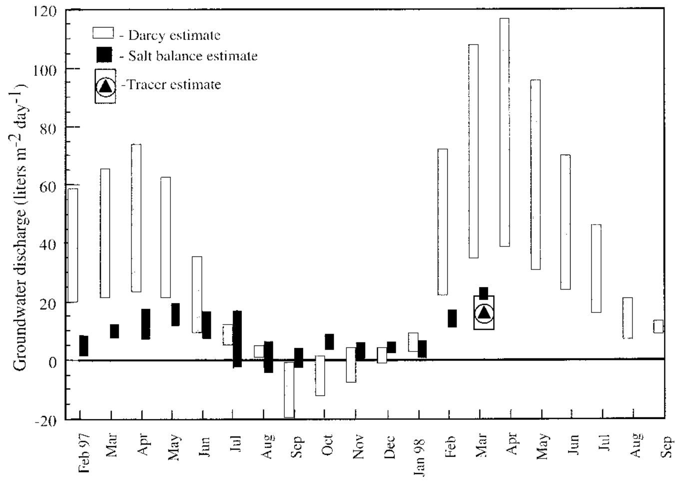


Fig. 7. Comparison of groundwater discharge estimates: Darcy, salt balance, and tracer test. The size of the vertical bars represents the range of estimates. The range presented for the Darcy estimate is the sum of minimum horizontal and vertical estimates (min) and the sum of the maximum horizontal and vertical estimates (max). Minimum or maximum estimates were determined by subtraction or addition of the standard deviations to the individual means. Determination of the range for the salt balance estimate was derived from standard deviations estimated from the Monte Carlo simulation. Coefficient of variation for the March tracer test estimate is 0.30. The height of the rectangle bordering the tracer icon denotes the standard deviation (5.1 L m^{-2} d^{-1}) of tracer discharge estimates calculated from piezometers 2, 3, and 4.

to the mean (in descending order), the model was most sensitive to changes in Q_T, followed by C_W, Q_P, Q_{ET}, and dV/dt.

The March 1998 tracer experiment yielded peak bromide breakthrough times of 12, 17, and 9 d for piezometers 2, 3, and 4, respectively (Fig. 6B). The mean transport velocity was 2.95 cm d^{-1}, which corresponded to a groundwater discharge of 16.6 $\pm$ 5.13 L m^{-2} d^{-1}. Although tracer transport to piezometer 4 was the fastest of the target piezometers, the highest observed bromide concentration was observed in piezometer 3. This nonuniform migration of the solute front suggested the dominance of transport through heterogeneously distributed macropores characteristic of marsh peats and may be the principal cause of the factor-of-two difference in linear velocities between piezometers. Because elevated Br^- was not detected in the 70–100-cm-deep piezometers until the end of the 63-d sampling period, downward transport of the plume (i.e., sinking) was considered insignificant.

Mean monthly discharge estimates derived from each method with associated error (presented as a range) illustrate similar seasonal patterns of discharge, but Darcy-derived fluxes were 200% of the salt balance estimates at high discharge and 50% at low discharge (Fig. 7). The range of discharge estimates was largest (38–118 L m^{-2} d^{-1}) at high flow using the Darcy approach in March 1997, when the salt balance predicted a discharge range of 21–24 L m^{-2} d^{-1}. The Darcy method however, yielded the smallest range in discharge estimates at low flow in August (0.5–1.2 L m^{-2} d^{-1}). The salt balance estimate in March 1998 (22.5 L m^{-2} d^{-1}) was within the standard deviation of the Br^- tracer discharge estimates (~26% lower than the mean tracer calculated discharge at that time).

Discussion

Seasonal discharge patterns—Both the Darcy and salt balance approaches identified a similar seasonal pattern of groundwater discharge, but differed in their estimates of discharge during periods of both maximum and minimum discharge. Despite differences between methods, even the most conservative estimate of peak discharge was sufficient to change the salt content of marsh pore water and mediate export of marsh-derived solutes toward the estuary (Fig. 5).

Our estimates of groundwater discharge through the wetland were relatively large in comparison to previous studies. Discharge peaked during the early spring, a time when few other studies have characterized the hydrology of tidal wetlands. During the summer growing season, however, groundwater discharge was lower and within the range of previously reported values for temperate marshes (Table 3). August groundwater fluxes were ~40% of discharge values estimated using Darcy calculations at nearby Carter Creek, Virginia, which is dominated by tidal inputs (Harvey and

Table 3. Comparison of groundwater discharge into tidal ecosystems. Additional groundwater flux estimates prior to 1990 are summarized in Harvey and Odum (1990). Negative sign (−) denotes flow from marsh into aquifer.

Reference	Location(s)	System	Groundwater flux (L m^{-2} d^{-1})	Method	Period
Chambers et al. 1992	Philips Creek, VA	Coastal marsh	0.012	Darcy	Jun, Oct
Nuttle and Harvey 1995	Philips Creek, VA	Coastal marsh	2.8	Water–salt	Aug–Sep
Price and Woo 1988	James Bay, Ontario	Coastal marsh	0.006	Darcy	Summer
Harvey and Odum 1990	Eagle Bottom, VA	Freshwater marsh	0.2	Darcy	Aug
	Carter Creek, VA	Estuarine marsh	1.0	Darcy	Aug
This study	Ringfield Marsh, VA	Estuarine marsh	−8.0–80.0	Darcy	Monthly*
This study	Ringfield Marsh, VA	Estuarine marsh	0.6–22.6	Water–salt	Monthly*
Bokuniewicz and Pavlik 1990	Long Island, NY	Subtidal	10.0–70.0	Seepage	Jun–Aug
Cable et al. 1997	St. George Sound, FL	Subtidal	2.9–9.0	Seepage	Monthly†
Giblin and Gaines 1990	Cape Cod, MA	Subtidal	24.0–72.0	Seepage	Jun, Jul
Portnoy et al. 1998	Nauset Marsh, MA	Subtidal	72.2	Seepage	Jun
Staver and Brinsfield 1996	Wye River, MD	Subtidal	2.2–37.4	Darcy	Monthly*

* Mean annual minimum and maximum.
† June–September values reported.

Odum 1990) and nearly equal to a water balance estimate of the late discharge into a Virginia coastal marsh whose groundwater input exceeded tidal infiltration (Nuttle and Harvey 1995). Both the Darcy and salt balance approaches, however, yielded summertime estimates of discharge that were nearly two orders of magnitude larger than that reported for a subarctic coastal marsh where precipitation dominated the water inputs (Price and Woo 1988). In comparison, previously reported subtidal groundwater discharge fluxes exceeded our discharge estimates by one to two orders of magnitude for similar seasonal periods (Table 3).

The relative importance of the horizontal and vertical flow components of discharge shifted through the year. In the late winter and spring, high discharge was accompanied by a greater proportion of horizontal groundwater flow when water table head exceeded the elevation of the hydraulically less conductive basal wetland deposits. Rapid horizontal flow in shallow sediments was also evidenced by the Br^- tracer transport during high discharge in March 1998 (Fig. 6).

The increased head in the wetland caused by the high groundwater input decreased the size of the unsaturated zone ($Z_{SED} - h_{MIN}$) available to receive tidal infiltration. Price and Woo (1988) similarly found tidal inputs to intertidal marshes to be insignificant to the water balance of a subarctic coastal marsh when sediments were saturated. Consequently, the large groundwater flux during this spring replaced tidal infiltration as the dominant input to the Ringfield marsh sediment water budget (Table 2). For this infrequently flooded high marsh zone, the higher percentage of groundwater to the total inputs is consistent with previous work showing the importance of groundwater to water balances in the absence of frequent regular tidal inundation (Hemond and Fifield 1982; Nuttle and Harvey 1995). The high groundwater discharge resulted in the flushing of pores with freshwater and export of the displaced pore water toward the estuary, which was evident as maximal drainage (Q_D) in the spring (Table 2).

Although precipitation was lower in spring 1997 than in summer 1997, a greater proportion recharged the shallow aquifer in the spring when evapotranspiration was still relatively low. The timing of the peak groundwater discharge appeared to be controlled both by the pattern of precipitation and the degree of evapotranspiration. As upland head decreased into the summer, groundwater discharge became decreasingly dominated by the horizontal flow component, culminating in July and August when the vertical and horizontal flux vectors were approximately equal. This shift likely resulted from a decrease in the upland head due to watershed evapotranspiration in summer, along with an increase in the effect of wetland evapotranspiration (Hemond and Fifield 1982; Dacey and Howes 1984). Because many previous studies have looked at the hydrologic balance in wetlands within the context of macrophyte ecology, they were performed during the growing season in the summer when groundwater inputs were low and vertical flow vectors were strongest. Hence, the horizontal transport of fresh groundwater through wetlands has usually been considered to be negligible (Nuttle and Hemond 1988). Although vertical transport into the wetland was the primary route of groundwater input in the summer (July and August), it comprised $<$20% of the total groundwater flux annually. Instead, most of the groundwater entering the wetland annually flowed horizontally during spring discharge.

Comparison of methods—One of the goals of this study was to compare the estimates of groundwater flux obtained using the Darcy approach with the salt balance and the Br^- tracer experiment. Although Darcy calculations can describe the flow of groundwater, which may have been mixed with saline pore water on previous tidal cycles, the salt balance is constrained to estimating the influx of freshwater only. However, the constancy of salinity (0 ppt) in the upland well (RU-4) throughout the year indicated that both methods should be estimating the flux of fresh groundwater to the Ringfield marsh.

Although the seasonal pattern of discharge is similar for the Darcy and the water/salt balance methods (Fig. 7), the comparison between methods can be divided into the following three periods: June through August when the groundwater discharge estimates were nearly equal (which we do not need to address), February through May (high flow) when Darcy estimates exceeded the salt balance predictions, and September through December (low flow) when the salt balance output exceeded Darcy estimates in magnitude and differed in flow direction.

High flow—During high-flow conditions from late winter through spring, Darcy fluxes were approximately two times higher than the salt balance estimates at peak discharge. It is possible that the Darcy-measured flux included some saline water of estuarine origin and represented an overestimate of true fresh groundwater input. Increasing the salinity of the C_{GW} term in the salt model increased the calculated groundwater flux to values close to the range of Darcy estimates at high flow. Yet because this C_{GW} increase also pushes the salt balance estimate out of the range of the tracer discharge estimate, salt water in-mixing prior to discharge it is not likely to be the sole reason for the Darcy overestimates. A simpler explanation is that the high Darcy estimates resulted from incorrect estimates of hydraulic conductivity (K) along the flow path. Results from duplicate slug tests performed on nearby piezometers (0–50 and 50–100 cm deep) replicated within a factor of two. In addition to spatial variation, K may have varied temporally in response to changes in peat structure during the growing season or between periods of saturation and desaturation (Nuttle and Hemond 1988). Furthermore, Darcy estimates of flow contained a larger uncertainty than salt balance estimates at high discharge. Estimated error in the Darcy groundwater flux increased proportionally to the mean discharge and was thus maximal (exceeding 100% of mean discharge) in the spring. The large errors reported for the Darcy calculations were estimated using variability in hydraulic heads measured at infrequent intervals against a background of tidally fluctuating water levels. That source of variability could have been reduced with more frequent head measurements. However, we think that our error estimates are conservative because they ignore the uncertainty in estimating K_v and K_h associated with sediment heterogeneity between piezometers and the inaccuracies in the slug test method. Uncertainty in K_v may have been further compounded by possible sediment anisotropy. Winter (1981) estimated ~50–100% error in measuring hydraulic conductivity in anisotropic sediments. Errors in estimating hydraulic conductivity error have previously been identified as a likely source of disparities between groundwater fluxes estimated using K and those calculated from mass balance measurements (Chambers et al. 1992; Nuttle and Harvey 1995).

During peak discharge, the salt balance model estimated the groundwater flux at 50% of the Darcy method and was corroborated by the results of the Br^- tracer release. Previously tracers have been used in wetlands to estimate drainage and pore-water turnover (Harvey et al. 1995) but not for the purpose of estimating groundwater inputs to the system. We assume that because the tracer estimate is an empirically observed quantity, it is probably the best estimate of the average groundwater flux during the first few weeks following the tracer release. Tracer-derived estimates of discharge are closer to the trend and range of average values predicted by the salt balance model in March 1998 and outside the range of error associated with the Darcy method. Furthermore, during high flow, the salt balance estimate of Q_{GW} was well in excess of the estimated error from the Monte Carlo simulation, which ranged between 8 and 40% of the mean discharge. Consequently, confidence in the salt balance groundwater flux estimate was greatest at that time. Given the discharge and error estimates from the Darcy and salt balance methods, calculation of groundwater-driven solute fluxes into or out of the wetland could be overestimated up to fivefold using the Darcy estimate at high flow.

Low flow—During low-flow conditions in the late summer and fall, the salt balance approach predicted groundwater discharge into the marsh of a magnitude three times higher than the Darcy-predicted flux. This predicted flow of groundwater into the wetland was opposite to the measured hydraulic gradient. It is unlikely that any substantial fresh groundwater flux into the marsh occurred during periods when the hydraulic gradient predicted net flow from the wetland into the hill slope aquifer. Salt intrusion into the aquifer underlying the wetland was observed during early fall (Fig. 5) and thus provided support for the slight flow reversal predicted by the Darcy method. The increased estimate of error encountered in the salt balance during low flow (exceeding mean discharge by 10-fold) is further evidence of the weakness of this approach at low flows. In contrast, the estimate of error for the Darcy-calculated groundwater flux at this time was only 70–80% of the estimated mean groundwater discharge.

Several factors may contribute to the inaccuracy of the salt balance at low flow. High evapotranspiration, a highly uncertain term in many wetland water balances (Winter 1981; Carter 1986; Hunt et al. 1996), was coincident with the highest estimates of error in mean discharge. However, the salt balance showed only moderate sensitivity to variations in Q_{ET}. In contrast, large error estimates encountered in September through November may reflect individually larger monthly variances associated with the more dominant model inputs at that time (Q_T and dS). These are the terms whose uncertainty also had the greatest effect on the salt balance estimated Q_{GW} as determined by the sensitivity analysis. Specifically, the inability to accurately estimate the rapid changes in salt storage occurring at this time (Fig. 4) may have affected the reliability of groundwater discharge calculations. The inability to adequately characterize spatially averaged solute concentrations has been shown to explain large errors in estimating groundwater flow using water and solute mass balances (Choi and Harvey 2000). We suggest that while mass balances have provided the most reliable estimates of groundwater flux in more hydrologically isolated wetlands (Hunt et al. 1996), in fringing intertidal wetlands when groundwater discharge is a small component of the overall water budget (i.e., during low flow), the mass balance approach yields a less certain estimate of Q_{GW} than the Darcy approach. Predictions of solute fluxes based on the salt balance method at low discharge would fail to identify the wetland as a source of compounds to the hill slope aquifer at that time.

Table 4. Comparison of wetland–estuary solute fluxes. Tidally driven fluxes versus the groundwater-driven solute fluxes at Ringfield during high discharge. Fluxes are reported as a range from multiple studies followed by the mean flux in parentheses and the number of studies reviewed by the reference author(s). All fluxes except this study are normalized to a 90-day interval from annual rates. The DOC and NH_4^+ flux values reported by Childers (1994) were derived from flux data regressed against tidal height (r^2 = 0.91 and r^2 = 0.87 for DOC and NH_4^+, respectively) such that the flux reported in this table was determined from the linear regression for the tidal range observed at the Ringfield study site. DOC fluxes summarized by Jordan et al. (1983) were also reviewed in Taylor and Allanson (1995).

Reference	Flux: DOC (g C m^{-2} marsh [90 d]$^{-1}$)	DON (g N m^{-2} marsh [90 d]$^{-1}$)	NH_4^+ (g N m^{-2} marsh [90 d]$^{-1}$)
Taylor and Allanson 1995	2.0–80.8, (26.0), n=11	—	—
	3.2–10.3, (5.7), n=4*	—	—
Jordan et al. 1983	See above	0.4–2.3, (1.0), n=6	−0.5–0.7, (0.2), n=8
Childers 1994	36.0, n=4	—	0.75, n=5
This study	11.9	1.6	1.3

* DOC fluxes for irregularly flooded high marshes.

Implications for pore-water flushing and wetland–estuary exchange—The flux of water governs the exchange of materials in and out of wetlands. Consequently, Lent et al. (1997) suggested that the dominance of different components of water balances defines the degree of wetland interaction within the landscape. The Ringfield marsh was thus more hydrologically isolated within the landscape in the late summer (July and August) when precipitation and evapotranspiration were a greater proportion of the sediment water budget. Conversely, wetland connection with the watershed and estuary was maximal in the spring when the water budget was controlled primarily by groundwater and drainage fluxes. Aside from a seasonal input of nutrients from anthropogenically affected watersheds, the ecological ramification of this shift is twofold.

First, the wetland subsurface is purged of pore water in the spring. The sediment was poorly flushed in the late summer and accumulated biogeochemically relevant solutes such as dissolved organic carbon (DOC) and H_2S as well as salt (Tobias 1999). Although increased flushing of the subsurface began in the fall with greater rates of tidal infiltration, the major purge of pore-water metabolites occurred in the spring when groundwater dominated inputs to the sediment. Based on the salt balance, the estimated pore-water turnover in the wetland sediment in March 1998 driven solely by the maximum groundwater discharge was ~32 d, which was a rate sufficient to seasonally alter subsurface geochemistry (Tobias 1999).

Second, the groundwater-mediated flushing of pore water may represent a sizeable pulse of marsh-derived solutes to the adjacent estuary during a period when diffusive marsh and subtidal benthic fluxes are typically smaller because of lower springtime water temperatures (Taylor and Allanson 1995; Cowan and Boynton 1996). For example, using the average spring pore-water turnover time and the average annual pore-water concentrations of DOC, dissolved organic nitrogen (DON), and NH_4^+ of 871, 100, and 80 μM (Tobias 1999), the groundwater-driven flux of these constituents from April to June 1997 (90 d) would be 12.2, 1.6, and 1.3 g C or g N m^{-2} marsh (90 d)$^{-1}$, respectively (Table 4). These estimates are similar in magnitude to results from previous studies that only considered surface exchanges and exceed the reported flux values for higher elevation irregularly flooded marshes more similar to our site (Table 4). Although we are not implying that groundwater-driven fluxes are always dominant, we do suggest that this groundwater-induced flushing may be a mechanism that seasonally exports a significant proportion of the total solute flux from fringing wetlands to adjacent water masses.

Groundwater fluxes through wetlands are typically highest nearest the upland border where we made our measurements (Harvey and Odum 1990; Nuttle and Harvey 1995). In large expansive marshes where much of the total marsh area is distant from high-discharge zones near the upland, this advective flux of pore-water solutes would likely constitute a small percentage of the total flux (Howes and Goehringer 1994). Therefore, the importance of groundwater in seasonally purging pore water and mediating solute flux to adjacent water bodies is likely to be greatest in long narrow marshes with a large frontage but small total area more typical of our site, the Chesapeake Bay, and its subestuaries. As such, this seasonal flushing by groundwater may be a mechanism by which these fringing wetlands are relieved of accumulated salt and export material to estuaries.

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