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Spatial patterns in soil water fluxes along a forest-marsh transect in the southeastern United States

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Abstract. Time series of water levels in piezometer nests along a forest-marsh transect near North Inlet, SC, show fluctuations that are attributable to recharge by precipitation and tidal flooding and to removal by evapotranspiration (ET) and seepage out of the soil. Volumes of water associated with these water level changes were estimated by correlating rain-induced water level increases with measured rainfalls. In the forest the ratio of water table rise to rainfall is about 10. This ratio increases with decreasing elevation to about 40 in the mid marsh where the antecedent soil moisture is generally higher.

The relative influence of removal by ET and seepage and recharge by rain and tides varies systematically along the transect. In the mainland forest, loss of water by ET is somewhat less than infiltration, leading to a net recharge of fresh water which eventually discharges in the adjacent mid marsh. With decreasing elevation, the relative importance of recharge by rain decreases as recharge by tidal flooding increases. In the low marsh, however, these mechanisms of loss and recharge can not be discerned in the water level time series because the water table rarely, if ever, drops below the marsh surface.

Key words. Forest; marsh; recharge; evapotranspiration; soil water.

Introduction

The forest-salt marsh interface is a major, but largely unstudied, ecological boundary (ecotone) as it marks the transition between terrestrial and marine ecosystems along low gradient coasts such as the southeastern United States. The boundary is of interest not only because of its great linear extent but particularly because it is characterized by a sequence of botanical zones that parallel the main forest-marsh boundary. In the southeastern United States the transition typically features the following sequence of botanical zones; *Quercus* + *Pinus* → *Juniperus* → *Iva* → *Juncus* → *Borrichia* → *Salicornia* → short *Spartina* → medium *Spartina* → tall *Spartina*. In general, this sequence is controlled by elevation through its effect on the frequency of tidal inundation and thus soil salinity. However, Thibodeau et al. (1998) showed that soil salinity, and thus botanical zonation, are also influenced by groundwater flow patterns. Where fresh ground water upwells beneath marsh areas adjacent to extensive uplands, soil salinity is depressed and salt tolerant species, such as *Salicornia*, are absent from the ecotone at elevations where they would otherwise occur. Upwelling ground water from adjacent uplands may also affect plant productivity by supplying nutrients and flushing toxic sulfides (Bradley and Morris, 1991; Chalmers, 1982; Osgood and Zieman, 1998; Valiela et al., 1978). Furthermore, removal of soil water by drainage and/or evapotranspiration induces entry of air into the soil

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(Dacey and Howes, 1984; Hemond and Chen, 1990; Howes et al., 1986), promoting oxidizing conditions which may impact nitrogen cycling and sulfide oxidation (Bowden, 1986; Brosemer et al., 1987; Chalmers and Odum, 1990; Harvey and Odum, 1990). As patterns of soil salinity, nutrient cycling and sediment oxidation depend on spatial variations in the local soil water balance, a comprehensive understanding of ecosystem structure and function across the ecotone will require quantification of water balance fluxes across the landscape.

Most previous studies of soil water fluxes in temperate-region salt marshes have focused on relatively few sampling sites and for periods rarely exceeding a month or so (Harvey et al., 1987; Harvey and Odum, 1990; Hemond and Fifield, 1982; Hughes et al., 1998; Nuttle and Harvey, 1995; Osgood, 2000; Tobias et al, 2001). Generally these workers assumed that evapotranspiration (ET) was equal to potential evapotranspiration (PET) which was calculated using various methods (Hamon, 1961; Penman, 1948; Priestly and Taylor, 1972). Similarly, infiltration, if considered at all, is usually assumed to be equal to the resulting rise of the water table multiplied by the specific yield (Tobias et al., 2001). The only attempt to directly measure infiltration (and exfiltration) was that by Hemond et al. (1984) who found that this component of the water balance in a New England marsh amounted to only a few millimeters per tidal inundation. Given the elaborate instrumentation required for such measurements (Hemond and Burke, 1981), it is not surprising that subsequent studies have resorted to theory and calculation in estimating this flux.

We applied a hydrologically intuitive method for identifying and quantifying the infiltration of rain water across the soil surface from time series of watertable elevations along a forest-marsh transect. The method does not require measurements of specific yield because it is based on the correlation of watertable rises to measured rainfalls. For the infiltration of tidal water it assumes that the correlation between watertable rise and rainfall applies to tidal recharge as well because there is no reason to believe that, on average, soil moisture conditions at a particular site prior to rain events differ from those before tidal inundations. We also show that, at stations where drainage due to downward hydraulic gradients is absent, the method yields reasonable estimates of ET. Applying this method, we report annual estimates of these water fluxes across the forest-marsh ecotone at North Inlet, South Carolina.

Methods

Study site

This study was conducted in the forest-marsh-estuary system near the Belle W. Baruch Coastal Research Field Laboratory at North Inlet, Georgetown County, South Carolina (Fig. 1). This area consists of a modern Holocene beach and inlet facing the Atlantic Ocean to the east. Landward, to the west, lies an extensive area (32 km^2) of Holocene marsh which on the west and north is bordered by a low-lying upland forest rooted in Pleistocene beach ridge sands. To the south the marsh is bounded by Winyah Bay, the estuary of the Pee Dee River. The marsh has evolved under a regime of slowly rising sea level and accordingly has transgressed over formerly forested beach ridge sands (Gardner and Bohn, 1980; Gardner et al., 1992; Gardner and Porter, 2001). Fingers of marsh along the western margin of the basin formed recently in swales between beach ridges and thus are in early stages of botanical succession. The marsh-estuary system is subject to a semidiurnal tide with a mean range of 1.5 meters Kjerfve (1986). The climate of the area is subtropical with a mean annual rainfall of about 140 cm.

Figure 1 shows the locations of three transects in the Crabhaul Creek basin along which topographic, botanical and salinity data and time series of watertable elevations have been collected (Thibodeau et al., 1998). Ground elevations along the transects were measured with a surveyors theodolite using the U.S. Geodetic Survey bench mark at the intersection of Marsh and Clambank Roads (33.335N, 79.210W) as a datum (2.60 meters above mean sea level). This study focuses on transect D which, being the most seaward of the three, is in the most advanced stage of botanical succession and marsh development. It varies in elevation from about 175 cm above mean sea level at its northwestern end in the mainland forest to about –25 cm in the channel of Crabhaul Creek. Soils and sediments along the transect vary from permeable, medium to coarse, sand in the forest to low-permeability mud in the low marsh (Thibodeau, 1997).

Groundwater dynamics along the transect are described in Thibodeau (1997), Thibodeau et al. (1998) and Gardner et al. (in press). Briefly, flow in the western forest is predominantly horizontal towards the marsh as there are no measurable vertical components of the hydraulic gradient in the forest (Fig. 2). Moderate to strong upward gradients are present at piezometer stations in the western high and mid marsh (stations D34 to D58, Fig. 2) indicating upward seepage of water towards the marsh surface. In contrast the eastern high and mid marsh are characterized by downward gradients and thus flow (stations D187 to D261).

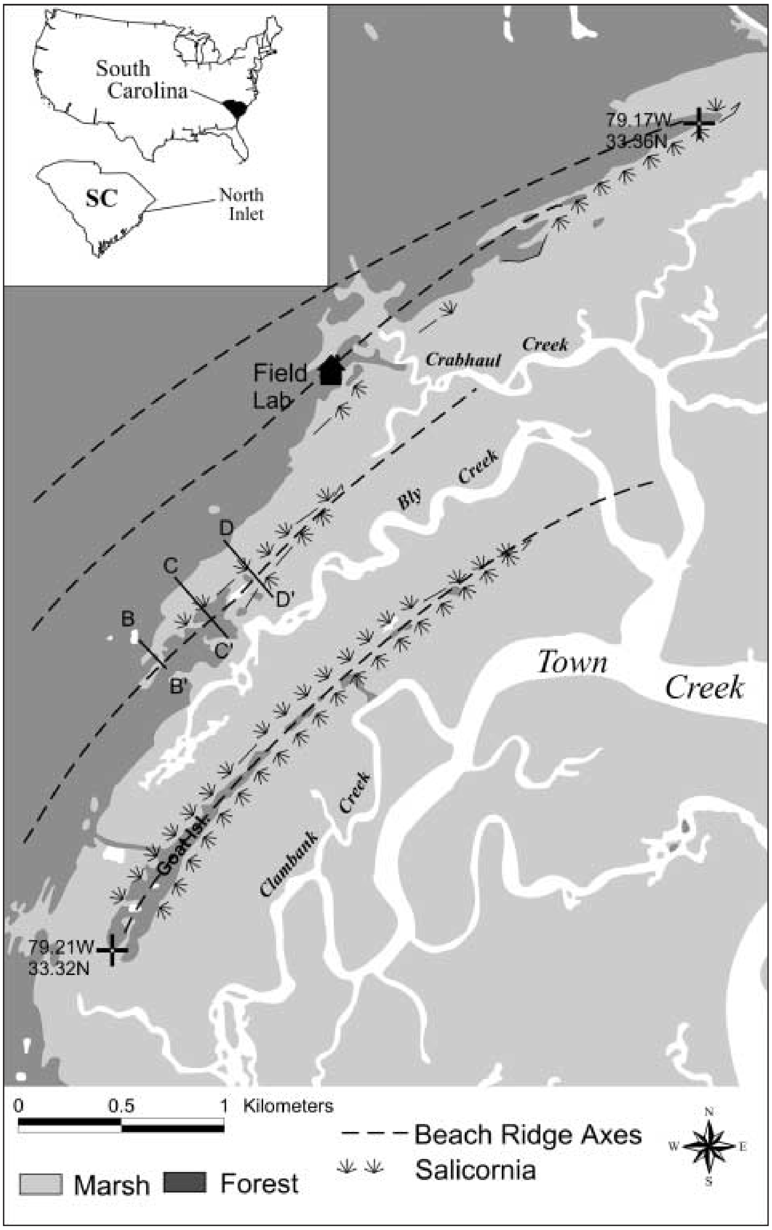

Figure 1. Study site location map. Note absence of *Salicornia* along western margin of marsh basin adjacent to the extensive upland forest.

Instrumentation

Piezometer nests were installed at 29 stations along this transect at depths of 0.6, 1.2, and 2.4 meters and, in addition, at 3.6 and 4.8 meters in the forest and high marsh stations. Piezometer stations are labeled according to their distance in meters from the first station in the western forest (D00). The piezometers consist of 3.2 cm inner diameter PVC pipe which was slotted and screened over a six cm interval at the installation depth. Each piezometer was instrumented with a data logger and pressure transducer in order to record time series of water elevations (hydraulic heads) in the pipes. The pressure transducers were housed in 2.54 cm outer diameter PVC pipes so that the annular space between the transducer and piezometer pipes was minimized to allow the system to respond rapidly to changes in ambient head. Details pertaining to the instrumentation can be found in Keenan et al. (1996). The data loggers recorded piezometer water levels at 15 minute intervals over deployments spanning 10 to 15 days beginning on March 3, 1994 and ending on August 21, 1995. Not all of the piezometers could be monitored simultaneously or continuously. Thus station coverage ranged from a low of about one month to at most one year (Fig. 3 and Table 1).

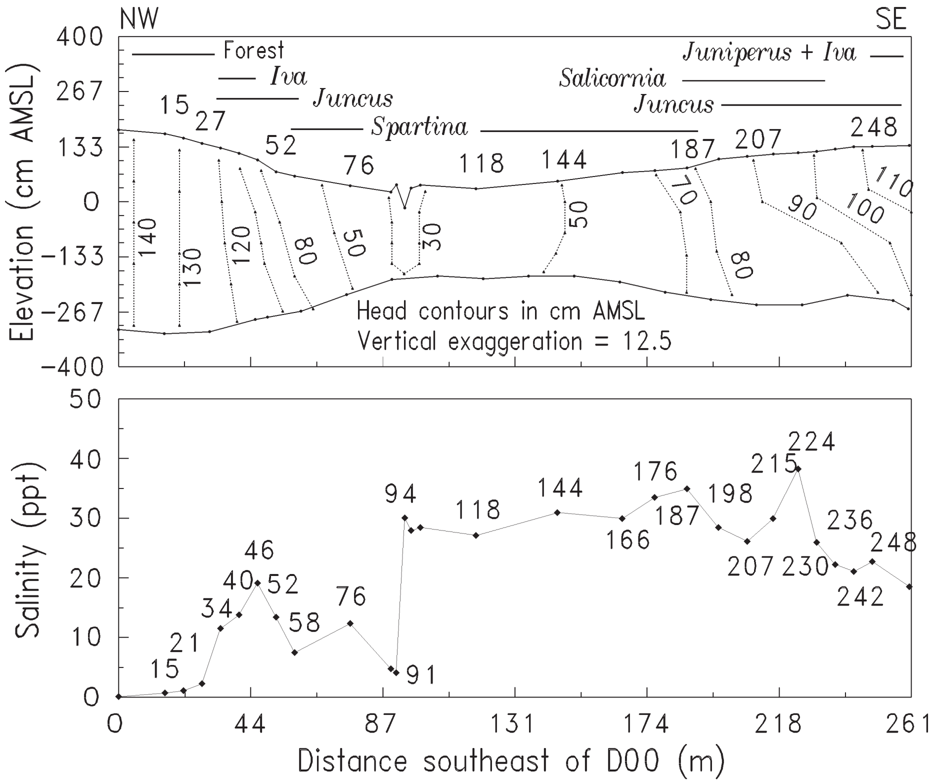

Figure 2. Cross section along transect D showing botanical zonation, salinity in the 0.6 m piezometers and contours of time-averaged hydraulic head. Base of the aquifer is assumed to be the top of the impermeable basal mud described by Thibodeau (1997). Numbers along the topographic profile and salinity data show locations of stations mentioned in the text. Note that sloping head contours between stations D34 and D76 imply upward components of groundwater flow in the western marsh whereas downward flow is implied by contours in the eastern marsh.

Salinity

Groundwater salinities were measured on a roughly bi-monthly schedule from 1994 to 1996. Piezometers were bailed three times before water samples were taken. Salinities were measured with a refractometer.

PET calculations

In order to constrain estimates of annual evapotranspiration, time series of potential evapotranspiration (PET) were computed using the Priestly-Taylor method (Priestly and Taylor, 1972; Dingman, 1994; Gardner and Reeves, submitted) and meteorological data obtained from the NOAA supported National Estuarine Environmental Research Site (NERRS) weather station about a kilometer to the northeast of transect D. There, air and water temperature, tide, salinity, solar radiation and precipitation are measured at hourly intervals. Relative humidity is not measured at the NERRS station so we used data collected at the Myrtle Beach NOAA station, approximately 50 kilometers northeast of the study site. Although the hourly relative humidity measurements at the Myrtle Beach and Charleston, S.C. (100 km southwest of North Inlet), weather stations are only moderately correlated ($R^2 = 0.5$), the average annual humidities are nearly equal (65 % versus 68 %) indicating that the extrapolation of the Myrtle Beach data to North Inlet probably does not introduce significant error into estimates of annual PET. An albedo of 0.2 and an emissivity of 0.95 were assumed for the marsh and forest based on data given by Dingman (1994). As plants cannot transpire during submergence (Salisbury and Ross, 1969), these calculations further assumed that loss of soil water via evapotranspiration does not occur whenever a site is flooded by the tide. Thus at each station PET has been prorated for the effect of variable tidal submergence along the transect.

Infiltration and ET estimation

Estimates of recharge by rainfall were made by regressing watertable rises, determined from water-level time series described in the Results section, against rainfall amounts measured at the NERRS weather station. Linear regressions were obtained for 13 of the 29 transect stations that had adequate data. For each station only those rain events that failed to drive the water table to the ground surface were used in the regressions. Rain events of greater than a critical size produce truncated responses

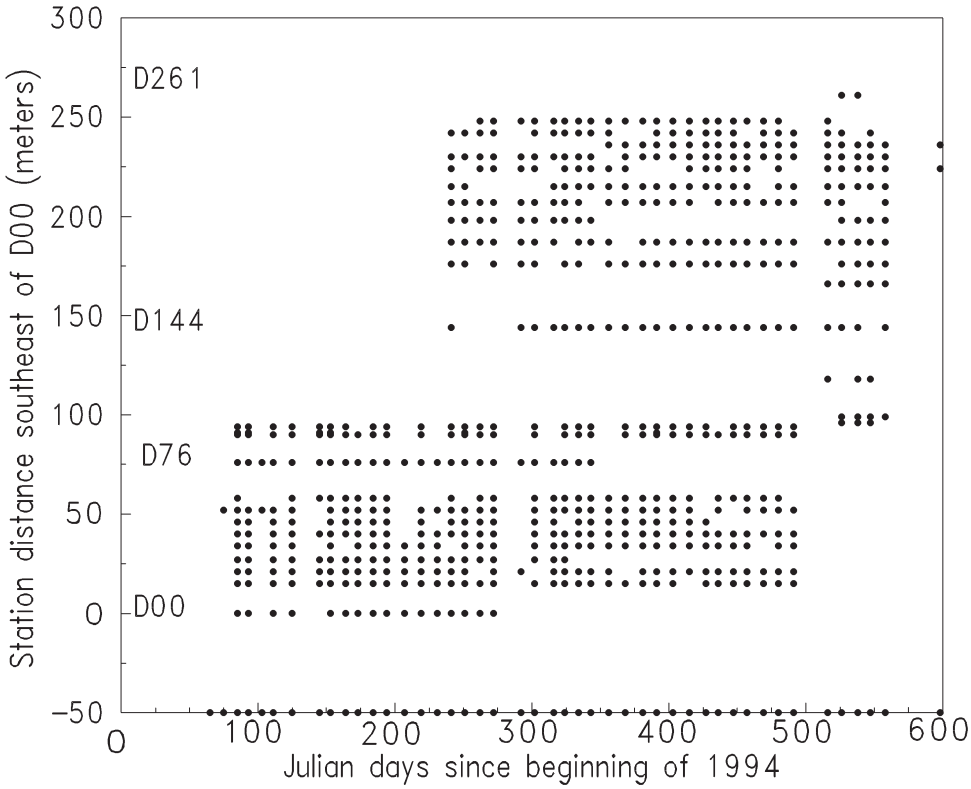

Figure 3. Spatial and temporal patterns of data logger deployments and water level data recovery along transect D. Solid circles along abscissa of the graph denote dates on which data were downloaded from loggers. Solid circles above the abscissa show stations that actually yielded data on those dates. Deployments typically lasted 10 to 14 days.

because the excess rain is lost by surface runoff. At each station the cumulative watertable rise for the entire length of record was determined and then divided by the slope of the regression equation (which represents the average response of the water table at the station to one centimeter of rain). To obtain annual estimates of infiltration, this result was multiplied by the ratio of the number of hours in a year (8760) to the number of hours of water level measurements.

Estimates of recharge due to tidal inundation also were based on the slopes of the regression equations for watertable rise versus rainfall amount. Here we assumed that, on average, soil moisture conditions preceding tidal inundations were similar to those preceding rain events, as there is no obvious reason to believe otherwise. Accordingly, we determined the cumulative watertable rise due to tides, divided by the slope of the regression equation for the station of interest, and then prorated to an annual basis as above. In determining tidal recharge, only the water-level rise between the pre-tide level and the ground surface was used because piezometer water levels higher than the ground elevation are due only to the loading of tidal water on the marsh surface and do not represent infiltration into the marsh soil.

In those areas where measurable vertical components in the hydraulic gradient are absent, such as between stations D00 and D34 (Fig. 2), drawdowns of the water table preceding recharge events are due largely, if not entirely, to ET. As described in the Results section, this is indicated by the step-like drops in the water table that occur only during daylight, with no changes in water level at night. Thus estimates of ET in such areas can be made by summing up the daily drawdowns and dividing, as before, by the slope of the regression equation. Where upward gradients in the hydraulic head prevail, as between D34 and D76 (Fig. 2), modest and slow rebounds of the water table commonly occur at night as ground water from below is driven into the near surface soil. Again the drawdowns during daylight can be summed to give estimates of ET using the regression slope. In areas where downward hydraulic gradients prevail, such as between D144 and D261 (Fig. 2), the water table continues to drop at night, albeit more slowly, due to downward drainage of soil water. In such areas it is not clear how to accurately separate the drawdown due to ET from that due to drainage and thus it is not possible to estimate ET on the basis of watertable time series alone.

Table 1. Total water table responses to rain, tide and ET. Number of events are given in parentheses.

Station	Elevation cm	LOR[a] hours	Rain cm	Tide cm	ET cm
D00 (F)[c]	175	3665	557 (39)	3 (1)	383 (114)
D15 (F)	164	8325	928 (82)	98 (9)	562 (189)
D21 (F)	154	8512	1067 (91)	166 (14)	536 (188)
D27 (F)	142	5013	637 (44)	164 (11)	562 (147)
D34 (H)	131	7840	600 (61)	198 (16)	662 (122)
D40 (H)	117	8375	448 (33)	592 (44)	1054 (175)
D46 (H)	101	6525	330 (26)	404 (37)	675 (113)
D52 (M)	73	7372	124 (15)	1394 (106)	1706 (141)
D58 (M)	62	7070	27 (5)	445 (88)	524 (92)
D76 (L)	38	5040	0 (0)	0 (0)	0 (0)
D90 (L)	23	7755	0 (0)	0 (0)	0 (0)
D91 (L)	41	980	0 (0)	0 (0)	0 (0)
D94 (C)	– 15	7755	0 (0)	0 (0)	0 (0)
D96 (L)	33	735	0 (0)	0 (0)	0 (0)
D99 (L)	41	1000	0 (0)	0 (0)	0 (0)
D118 (L)	31	5295	0 (0)	0 (0)	0 (0)
D144 (L)	50	6272	13 (2)	98 (24)	ind[b]
D166 (M)	71	1245	92 (8)	310 (29)	ind
D176 (M)	76	6796	116 (13)	1243 (135)	ind
D187 (M)	82	6565	158 (18)	1769 (213)	ind
D198 (H)	104	3590	216 (19)	1065 (89)	ind
D207 (H)	110	6410	790 (65)	1857 (128)	ind
D215 (H)	116	5675	990 (66)	1157 (65)	ind
D224 (H)	119	4775	904 (65)	831 (55)	ind
D230 (H)	122	7357	1177 (84)	821 (62)	ind
D236 (H)	127	5140	844 (55)	417 (27)	ind
D242 (H)	133	5280	1027 (72)	174 (9)	ind
D248 (H)	134	5320	928 (62)	213 (16)	ind
D260 (H)	137	530	108 (5)	39 (3)	ind

[a] LOR = Length of record.
[b] ind = indeterminate.
[c] F = forest, H = high marsh, M = mid marsh, L = low marsh, C = creek.

Finally, in the low marsh, between D76 and D144 the water table rarely, if ever, drops below the marsh surface making it impossible to estimate any of these fluxes using the methods described above.

Results

Watertable dynamics

We first describe how recharge by rain and tide and loss by evapotranspiration are manifested and quantified in water level time series along the transect. Figure 4 shows water levels in the 0.6 m piezometers at sites along the transect between July 3, 1994 and July 13, 1994. These include sites in the western forest at station D00, at the forest-marsh boundary (D34), in the mid marsh (D52) and in the low marsh (D76). Three distinct rain events occurred during this deployment on Julian days 183, 184 and 185. The first two caused the water table to rapidly rise several cm at both D00 and D34 while the third caused a rise of about 25 cm at D00 and 14 cm at D34. The rains had no observable recharge effects at D52 or D76 because the water table was already at the ground surface or because these sites were about to be flooded by the tide. The third rain event was large enough to cause the water table to rise to the ground surface at both D00 and D34. Thus at D00 and D34 the rain may have exceeded that required to completely saturate the soil whereas the earlier rains were clearly inadequate to produce complete saturation of the forest soil.

In addition to rain, these records show the effects of ET and tidal recharge. At D00 ET is manifested by a sequence of step-like, day-time drawdowns during the eight days following the Julian day 185 rain event, each step lowering the water table by about 5 cm. D34 shows a similar pattern except that the water table appears to rebound a centimeter or two each night. This is due to the fact that the hydraulic head in the deeper piezometers at this site is several cm greater than that in the 0.6 m piezometer (Fig. 2) so that water seeps upward into the soil from below. The upward gradient in head is even greater at D52 (Fig. 2), which also occasionally shows gradual night-time rebounds whenever the timing of the tides permit, such as on Julian days 188 and 192. Such

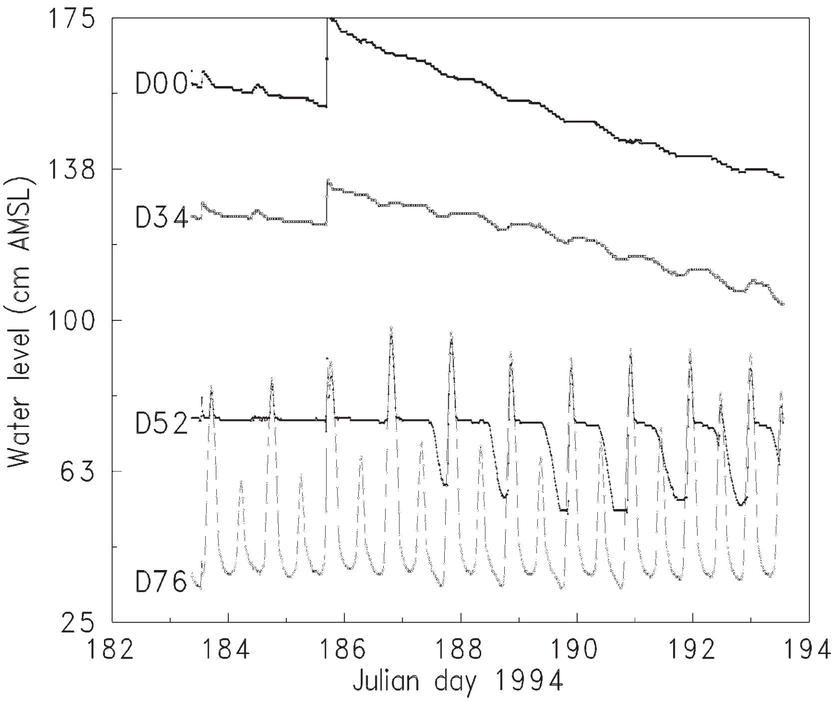

Figure 4. Representative water level time series in 0.6 m piezometers along transect D showing ET drawdowns and recharge by rain and tides. Julian day tick marks are located at the beginning of respective days. Station ground elevations are: D00 = 175 cm, D34 = 131 cm, D52 = 73 cm, D76 = 38 cm.

rebounds are absent at D00, as are upward gradients in head. When tides permit, the daytime drawdowns at D52 typically exceed 20 cm but are usually terminated by tidal flooding which quickly returns the soil to a state of complete saturation. In contrast, the low marsh station D76 rarely experiences drawdowns below the ground surface so that rebounds due to rain, tidal recharge or upward seepage are never observed (Fig. 4). Water-level time series at low marsh stations (D76 to D144) merely record the periodic loading of the marsh surface by the tide.

Between Julian days 183 and 194, nine ET events, three rain recharge events and no tidal recharge events were observed at D00 and D34. The ET events caused an aggregate drawdown of about 40 cm at both D00 and D34 while the large rain event on Julian day 185 caused the water table to rise 25 cm at D00 and 15 cm at D34 thereby reaching the ground surface at both stations (at elevations of 174 cm and 131 cm respectively). The small rain events on Julian days 183 and 184 caused rises of only several centimeters and thus failed to drive the water table to the surface. In addition, about 10 cm of night-time seepage recovery occurred at D34. At D52 observable ET drawdowns occurred on seven days resulting in a total drawdown of about 135 cm. Note that the drawdown on the last day of this record was interrupted by tidal recharge at about midday. Accordingly, eight tidal recharge events are evident in this record with an aggregate rise of about 135 cm. The rain event on Julian day 185 did not produce an observable recharge response at D52 because the water table was already at the marsh surface or because the surface was in the process of being flooded by the tide. No responses to ET, rain or tidal recharge are evident at D76 because ET and drainage apparently are not capable of lowering the water table below the marsh surface.

Flux event frequencies

This interpretation scheme was applied to all of the stations for which water-level time series are available. Table 1 shows the total response of the water table to rain, tide and ET events over the period of record for each station, along with the number of each event type (in parentheses). Because of various logistical problems, the length of record (Fig. 3 and Table 1) ranges from a low of 530 hours (22 days) to a maximum of 8512 hours (355 days). ET estimates could not be made for stations in the eastern marsh (D144 to D260, Fig. 2) because ET drawdowns are confounded with drainage driven by the downward hydraulic gradients that prevail on the east side of the basin.

Despite their limitations, some meaningful trends can be derived from the data in Table 1. By dividing the number of each type of event by the length of record one can obtain the event frequency (number/day) for each station. These are plotted as a function of station location

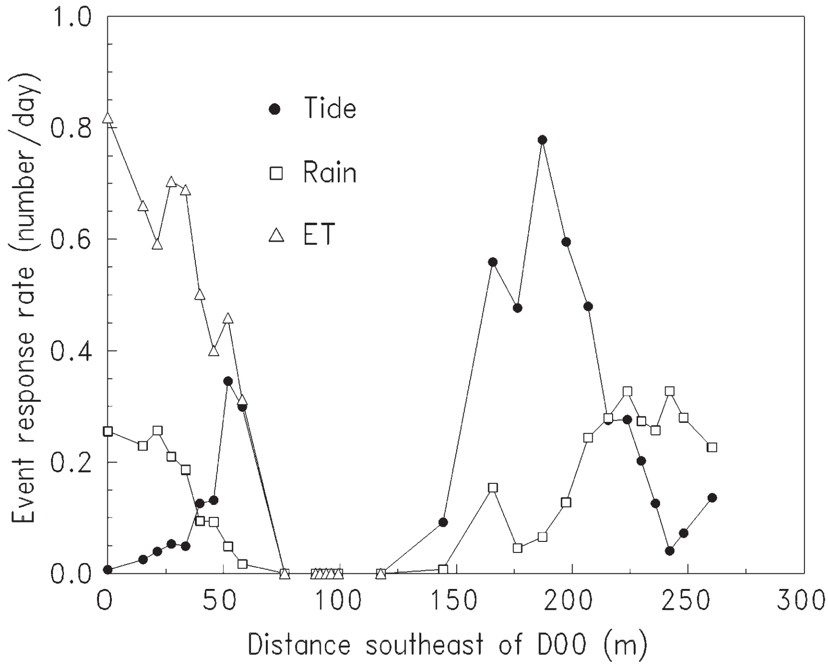

Figure 5. Event frequencies for recharge by rain and tide and loss by ET as a function of distance along transect D.

on Figure 5. As might be expected, the frequency of tidal recharge events increases with decreasing station elevation but then drops abruptly to zero in the low marsh because of the inability of ET and drainage to exhaust depression storage and thereby lower the water table below the marsh surface. More interesting is the fact that, for any given elevation, the frequency of tidal recharge in the eastern marsh is generally about two times higher than that in the western marsh. This asymmetry in frequency is due to the opposite directions of seepage in the eastern versus western marsh (Fig. 2). Upward seepage in the western marsh tends to keep the water table closer to the marsh surface than in the eastern marsh, thus reducing the probability that tidal flooding will result in a recharge event. The peak tidal recharge event rate, 0.78 day^{-1} at D187, is about 40 percent of the theoretical maximum of 1.93 day^{-1} so that at this station tidal recharge occurs on average once every two to three tides.

Rain recharge events in the forest and high marsh occur at a frequency close to that based on the complete rain record for 1994–1995 (0.23 day^{-1}) but the event rate drops off gradually towards the low marsh. In the western forest, the ET event rate averages 0.63 day^{-1}, somewhat below the theoretical maximum of 1.0 day^{-1}. This result seems reasonable given the fact that PET is low between November and March and during rain events. Overall these event frequencies are not unrealistic, indicating that the sampling provided by our data are adequate to define the general spatial trends in these processes.

Regressions

A summary of the linear regressions of water table rise versus rain amount used to estimate fluxes is given in Table 2. Examples of these correlations are shown by the scatter diagrams on Figure 6. For most stations the coefficient of determination (R^2) exceeds 0.80 and the intercept is close to zero, as would be expected for zero rainfall. The scatter in the data most likely is due to the fact that rainfall was measured at the Oyster Landing meteorological station about a kilometer north of transect D, so, for many events, the actual rainfall at transect D may have differed from that measured at Oyster Landing. Indeed, some small measured rain events failed to produce any water-level responses on transect D. On the other hand, some small water-level rises occurred at times when no rainfall was recorded. Variations in antecedent moisture conditions also may have contributed to the scatter.

Systematic variations in average moisture conditions along the transect may also be the cause of the general increase in the regression slope with decreasing elevation shown in Table 2 (with the exception of station D260 where only five measurements were available for the regression calculations). At lower elevations the interval between recharge events (rain plus tides) is relatively short and thus, both the depth to the water table and the pre-event saturation deficit are not as large as at higher elevations. Therefore, on average, less rain is needed to cause a unit rise of the water table at lower lying sites than at higher sites.

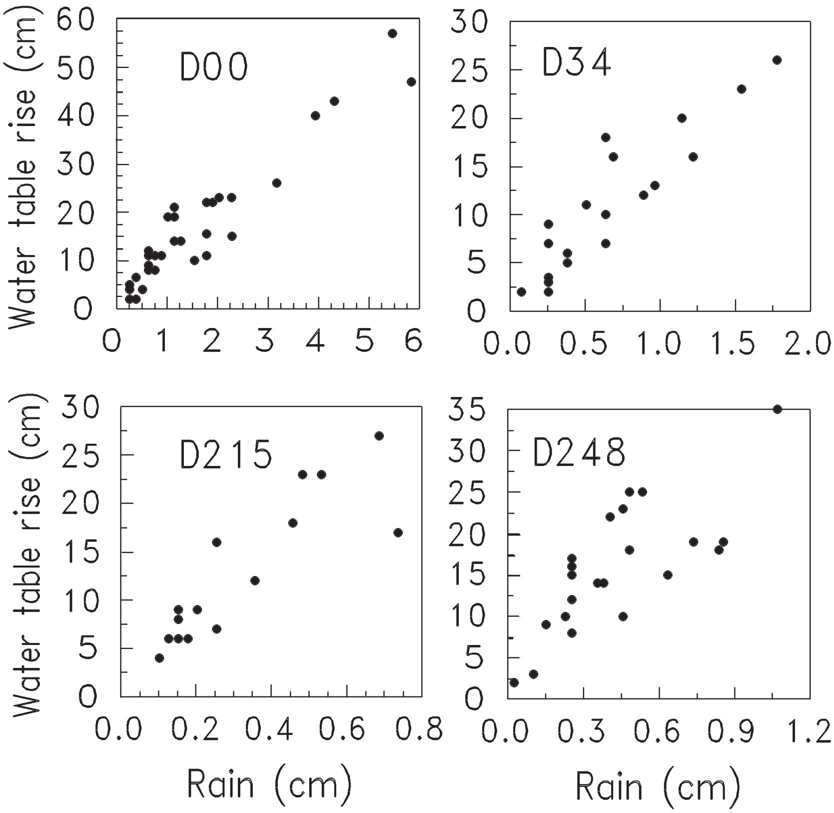

Figure 6. Scatter diagrams showing watertable responses to rainfall.

Table 2. Regressions of water table rise (in cm) to rainfall (in cm). Quantities in parentheses are parameter standard deviations.

Station	Elevation cm	Intercept cm	Slope	R^2	N
D00	175	3.24 (0.07)	8.60 (0.01)	0.90	34
D15	164	1.92 (0.04)	9.42 (0.02)	0.89	60
D21	154	1.22 (0.04)	10.98 (0.03)	0.85	58
D27	142	0.37 (0.09)	12.53 (0.04)	0.95	28
D34	131	1.61 (0.15)	14.14 (0.23)	0.84	21
D40	117	3.60 (0.82)	30.27 (2.21)	0.77	6
D215	116	2.80 (0.24)	30.86 (1.64)	0.76	15
D224	119	3.90 (0.57)	34.53 (13.6)	0.64	10
D230	122	3.20 (0.13)	31.04 (0.71)	0.85	20
D236	127	2.37 (0.14)	25.76 (0.63)	0.76	22
D242	133	4.94 (0.11)	24.14 (0.66)	0.67	28
D248	134	6.50 (0.17)	21.75 (0.68)	0.57	22
D260	137	– 1.26 (0.42)	39.75 (3.43)	0.99	5

Flux estimates

Recharge estimates along the transect based on the data in Tables 1 and 2 are shown on Figure 7. Also presented as solid diamonds on Figure 7 are estimates of ET on the western side of the basin based on the same method of calculation as for recharge but using the observed ET drawdowns in Table 1 and the regression slopes in Table 2. For comparison, PET along the transect based on meteorological and tide data are also shown. As will be discussed below, the ET estimates are probably minima. Note also that recharge estimates could not be calculated for some mid to low marsh stations (e.g., between D125 and D200) where recharge events were recorded but where there were not enough qualifying rain recharge events to establish regression slopes.

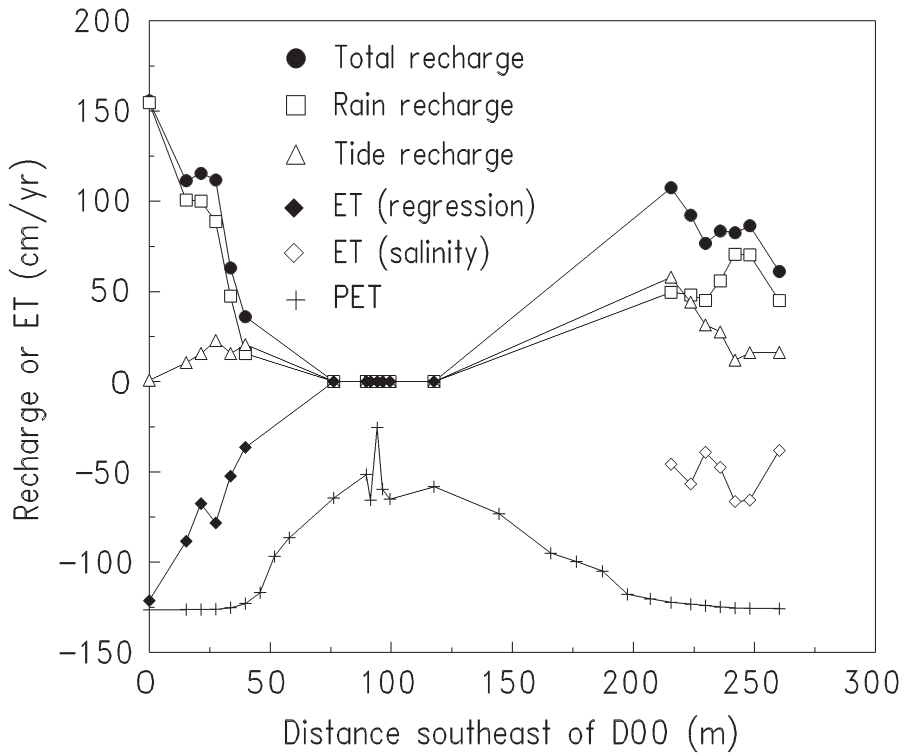

Figure 7. Estimated annual ET and recharge along transect D. See Discussion for the meaning and significance of ET (salinity) shown in open diamonds.

Tidal recharge in the eastern marsh appears to be roughly twice as large as that in the western marsh (Fig. 7), consistent with the generally higher salinity of the soil water in the eastern marsh (Fig. 2). It is also consistent with the downward hydraulic gradient in the eastern marsh which promotes drainage and thus provides more unsaturated pore space to accommodate infiltration. In the western marsh, peak tidal recharge appears to occur somewhere between stations D27 and D40, while in the eastern marsh it occurs at station D207. However, on Figure 5 it should be noted that peak frequencies of tidal recharge occur at stations D52 and D187. Thus, if regression equations were available for these stations, the peaks in tidal recharge might shift to lower elevations.

Based on analysis of measurements of rainfall and tide elevations made hourly during the course of this study, we found that virtually all of the rain (164 cm per year) occurred at times when the tide was lower than 130 cm. At station D00, 154 cm of rain (or 94 percent of the total rain) entered the soil as recharge. Between D00 and D34 the percentage of the total rainfall that infiltrated drops to 27 even though all of these stations are at elevations above 130 cm. Again this is due to the fact as the mean elevation of the water table approaches the ground surface with decreasing ground elevation, the pore space available to receive recharge diminishes. Also, as a result of the difference in the direction of the vertical hydraulic gradient beneath the western and eastern high marshes, rain recharge in the former (stations D34 to D46) appears to be somewhat lower than in the latter (D207 to D260).

Except for station D00, estimated ET in the western forest and high marsh is about 50 percent of the calculated PET. This is somewhat anomalous because one would expect that ET would be less limited by water availability in the moister soils at lower elevations. This problem is discussed below.

Discussion

Despite some data gaps, we feel that Figure 7 outlines the major features of recharge and ET along Transect D. These include the asymmetries in recharge by rain and tides on opposite sides of the basin. Overall recharge by rain is substantially greater in the western forest than in the eastern high marsh while peak tidal recharge in the eastern marsh appears to be at least twice as large as that in the western marsh. These asymmetries in recharge are probably largely responsible for the generally higher groundwater salinities observed in the eastern marsh and the corresponding asymmetry in botanical zonation (Fig. 2; Thibodeau et al., 1998).

ET estimates

As indicated above, the ET estimates for the west side of the basin are probably minimal as they are, for the most part, substantially lower than the tide-adjusted PET. The close agreement of the estimated ET with PET at D00 could be due to the fact that the data for this station were collected between mid March and October when ET is high. Had this station been monitored for an entire year, the ET estimate might have been lower. On the other hand, given the fact that the water table along most of the transect is rarely more than 20 cm below the surface, one might expect the actual ET to be close to PET, perhaps even at D00 where the maximum recorded depth to the water table was only 80 cm and averaged about 30 cm. The hypothesis that the actual ET in the western marsh is equal to PET is further supported by the high measured *in situ* rates of transpiration by *Juncus* and *Spartina* (Giurgevich and Dunn, 1982). However, if the actual ET is indeed equal to PET, then, with the exception of D00, ET would exceed recharge in the western marsh of the basin and one might expect to observe hypersaline soil water in the western high marsh as well as in the east.

The fact that this is not observed, as well as the fact that the estimated ET is only 50 to 70 percent of PET, can be explained as follows. West of D15, rain recharge appears to exceed PET so that there is left-over fresh ground water available to seep down gradient towards the marsh. The annual volume of this seepage can be estimated from the following data. The mean drop in elevation of the water table between D00 and D15 is 7.3 cm (Fig. 2) so that the mean horizontal hydraulic gradient between these two stations is 7.3 cm divided by 1500 cm or 0.0048. Based on slug tests in eight piezometers in this section of the transect, the mean hydraulic conductivity of the aquifer is 0.004 cm/sec (Thibodeau, 1997). The saturated thickness of the aquifer between these two stations is about 500 cm. Thus the annual volume of seepage arriving at D15, assuming essentially horizontal flow through a cross section 500 cm thick and 1.0 cm wide, is (0.004 cm/sec) × (0.0048) × (500 cm^2) × (31,536,000 sec) or about 303 liters. For comparison, we computed the difference between the annual volumes of estimated total recharge and PET for the transect between D15 and D58. (Downslope from D58 we assume that ET is entirely balanced by infiltration of water left in depression storage and seepage from up gradient during tidal exposure, as the water table in the low marsh never drops below the ground surface). In the low marsh, PET averages about 55 cm/yr which could be supplied by the infiltration of only 0.15 cm/day from depression storage.) From Figure 7, the average difference between PET and estimated total recharge is 53 cm/yr which when multiplied by the distance between D15 and D58 and the assumed width of the flow cross section, 1.0 cm, yields an annual deficit of about 225 liters. Thus, over this section of the transect, seepage from the western forest exceeds the deficit imposed by PET by about 78 liters/yr. Therefore, the hypothesis that actual ET is equal to PET is not precluded by the availability of water. Indeed there is enough left over to allow seepage of ground water to the land surface in an amount equal to about 78 liters/yr. Most of this probably occurs between stations D46 and D58 where upward hydraulic gradients are greatest. The failure of our estimates of actual ET (Fig. 7) to coincide with PET, therefore, may be due to the fact that the observed ET drawdowns are less than what would have been observed had there been no seepage from the upland forest to partially offset the loss of water by ET.

Although the hypothesis that actual ET equals PET is tenable for the western side of the basin, it encounters problems when applied to the eastern marsh. As noted earlier, it is not obvious how to estimate the effect of ET on the water table because on this side of the basin vertical hydraulic gradients are for the most part downward so that the drawdowns preceding recharge events are due to the combined effects of ET plus drainage. Thus, if we used total recharge as an estimate of actual ET, the result would be an overestimate because some of the recharge replaces water lost by drainage. Therefore, unlike the western side of the basin, the actual ET can not exceed the observed total recharge shown on Fig. 7. Indeed, if ET actually is equal to PET, salts should be precipitating in the soils of the eastern marsh because PET is roughly 50 cm greater than total recharge. In addition, there would be no water available to support lateral seepage from up slope, which at D248 amounts to about 120 liters/yr (based on the horizontal component of the hydraulic gradient of 0.0061 and an average hydraulic conductivity of 0.0017 cm/sec). If ET is equal to PET, total recharge at the eastern end of the transect would have to be on the order of 150 cm/yr in order to balance losses by ET and down-gradient seepage and to produce the observed soil salinity. This could be supported by the mean annual rainfall if all of it were to infiltrate. But this is unlikely because the lower elevation and moister soil conditions on this end of the transect should produce greater loss of rain water as surface runoff than in the western forest.

There are two possible solutions to the water balance problem posed by the eastern high marsh. First, our sampling of recharge events on the east side of the basin may have been too sparse to give accurate annual recharge estimates. On average, the length of record for the eastern stations is about 30 percent shorter than for the western stations (Fig. 3). Although it is not clear that this should lead to an underestimate of total recharge, it does seem strange that the total recharge decreases towards the end of the transect after reaching a peak at station D207. While one would expect tidal recharge to decrease with increasing elevation, as observed, one might also expect this decline to be compensated, or even exceeded, by an increase in rain recharge. Also the last four stations on the east end of the transect lie at about the same elevation as D34 and have about the same amount of recharge. In this regard one might expect the eastern stations to have greater recharge because downward seepage combined with ET should produce greater drawdowns than ET combined with upward seepage. In any case, if ET is actually equal to PET, then, as indicated above, we have grossly underestimated recharge by rainfall.

An alternative explanation is that, as a result of high soil salinity, actual ET is substantially lower than PET on the east side of the basin. As described by Thibodeau et al. (1998), salinity in the upper 20 cm of the soil in the eastern marsh typically ranges between 30 and 50 ppt whereas in the western marsh it is less than 20 ppt. As a result of the higher salt stress, the eastern marsh has broader swaths of the salt tolerant succulent, *Salicornia* (Fig. 2). Succulents such as *Salicornia* and *Borrichia* adapt to salt stress by minimizing transpiration (J. Morris, personal communication). Similarly *Juncus* seems to have a mid-summer dormancy strategy to deal with water stress (J. Morris, personal communication). Therefore, as a result of such adaptations and their substantially lower biomass, these plants transpire much less water than *Spartina*. In addition, *Spartina* productivity and transpiration decrease as salinity exceeds 30 ppt (Giurgevich and Dunn, 1979). Thus, as a result of greater tidal recharge and the absence of upwelling fresh water from forested uplands, soil salinity in the eastern marsh is high despite its lower actual ET.

To test the idea that ET is less than PET in the eastern marsh, we estimated ET from salinity data assuming that the near surface salinity is controlled solely by the interaction of ET with recharge by rain and tides. This assumption neglects the possible mixing of recharge with seepage from topographically higher areas of the marsh and thus will produce a minimum estimate for ET. Since groundwater flow in the eastern marsh has a strong downward component, salinity in the near surface soil should be controlled largely by the interplay of ET and local recharge. If this is so, then the time-averaged salinity, S, of the near-surface soil is given by:

$$S = \frac{R_t \times S_t}{R_t + R_r - ET} \tag{1}$$

where S is the time-averaged salinity (measured in the 0.6 m piezometers, Fig. 2), S_t is the average salinity of the flooding tide (assumed to be 32 ppt based on measurements at the Oyster Landing meteorological station), and R_t and R_r are estimated annual tidal and rain recharges (as shown on Fig. 7). Using data for S (Fig. 2), S_t , R_t and R_r, Equation 1 can be solved for ET, giving the results shown by the open diamonds on Figure 7. These results suggest that actual ET in the eastern marsh is about half of the computed PET and similar to measured transpiration by short *Spartina*, *Salicornia*, *Batis* and *Borrichia* at Sapelo Island, Georgia (Antlfinger and Dunn, 1979), but far lower than measured *Juncus* transpiration (Giurgevich and Dunn, 1982). These estimates also provide 10 to 30 cm of net recharge to balance downward seepage to deeper portions of the aquifer. Thus the hypothesis that ET is substantially lower than PET in the eastern high marsh is not unreasonable.

Flux model

Based on this discussion we propose the schematic model shown on Figure 8 for average annual water fluxes across the soil surface along transect D. Limits to ET and recharge by rain are imposed by the interaction of topography, tidal regime and climate. Actual ET and recharge events, however, are driven by stochastically distributed weather events. The mean recurrence interval for observable tidal recharge events ranges from months or years in the forest and low marsh to days in the mid marsh. Similarly the recurrence intervals for observable ET and rain recharge events range from days in the forest to months or years in the low marsh.

In the western forest, we postulate that rain recharge exceeds ET by about 10 cm and that ET is slightly lower than PET because during occasional summer droughts the water table may drop to a depth of a meter or so below ground, thereby stressing shallow rooted shrubs and grasses on the forest floor. The excess of rain recharge over ET in the forest allows for substantial seepage of fresh water to the marsh, where that portion not consumed by ET may discharge to the surface. Maximum tidal recharge in the western marsh occurs near station D40 in the high marsh (Figs. 2 and 8) because it floods more often than higher sites and, on average, is less water saturated prior to recharge events than lower stations. On the other hand, the maximum discharge to the ground surface of fresh ground water from the forest occurs at a lower elevation (near station D52) because this station exhibits the strongest and most persistent upward hydraulic gradients and is more frequently saturated to the soil surface. This station is also located at the concave upward break in slope between the high and low marsh where our preliminary numerical modeling of this system suggests maximum discharge to the ground surface. As a result of

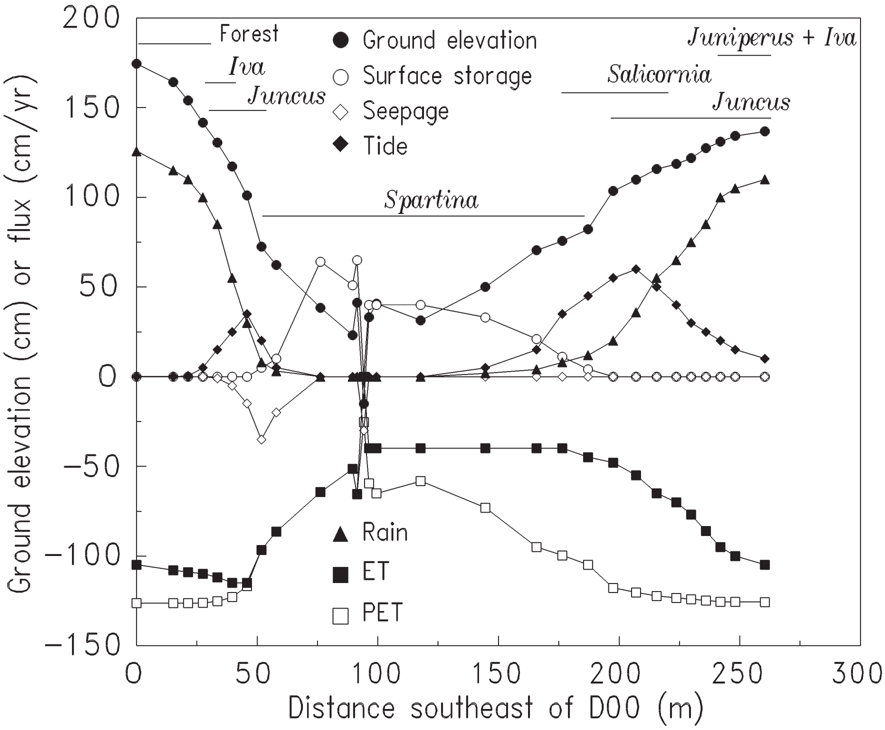

Figure 8. Schematic model for water fluxes across the soil surface along transect D as suggested by partial data shown in Figure 7. The reader should focus on the spatial patterns in the fluxes and not on the actual values which are preliminary estimates. Botanical zonation is also shown.

the offset between the sites of peak tidal recharge (D40) and peak groundwater discharge (D52), peak salinity in the 0.6 m piezometers is found at D40 rather than lower in the marsh (Figs. 2 and 8).

In the eastern marsh, we anticipate that tidal recharge should peak at station D207, which is at about the same elevation as the peak station in the western marsh. Again, this is due to the fact that this station is flooded more frequently than higher stations but has greater infiltration capacity than moister, lower lying sites. We also postulate that it should have a greater infiltration capacity than the peak western station because it looses water by both ET and downward seepage whereas ET at the western station is partly offset by upward seepage. The greater tidal recharge in the eastern marsh in conjunction with the discharge of fresh ground water in the western marsh should generate higher soil salinity in the eastern marsh. Furthermore, we suspect that the estimated rain recharge at the eastern end of the transect is probably an underestimate of the actual rain recharge because of inadequate sampling and should be closer to 100 cm/yr, as at D27 which is at a similar elevation. If so, ET would have to be somewhat higher than that estimated on Figure 7 in order to produce the observed salinities in the eastern high marsh. Finally in the low marsh, where the effects of ET and recharge are not manifested in the water-level time series, we can only assume that ET is equal to the submergence-adjusted PET and that this loss is largely balanced by infiltration from depression storage.

If this hypothetical scheme of soil water fluxes vary across the forest-marsh landscape is accurate, then several interesting ecological questions arise. Does the stress relief provided by seepage of fresh water from the forest result in higher productivity by plants in the western marsh or is it squandered by less efficient water use during photosynthesis? How does nutrient supply via lateral water flux compare to *in situ* regeneration? How important is the input of O_2 via recharge and aeration in controlling plant distribution and growth?

Conclusions

1. Long-term records of watertable elevations can be used to estimate annual recharge by tides and rainfall along forest-marsh transects.
2. Where downward gradients in groundwater hydraulic head are absent, minimum estimates of evapotranspiration can also be made.
3. For marsh sites at equal ground elevations, recharge by tides and rainfall is greater where downward gradients in head (and thus drainage) prevail than

where upward gradients (and thus upwelling) prevail. As a result, soil salinity, and thus plant assemblages and productivity, may vary among sites at a common elevation.

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References

Antlfinger, A. E. and E. L. Dunn, 1979. Seasonal patterns of CO_2 and water vapor exchange of three salt marsh succulents. Oecologia **43:** 249–260.

Bowden, W. B., 1986. Nitrification, nitrate reduction and nitrogen immobilization in a tidal freshwater marsh sediment. Ecology **67:** 88–99.

Bradley, P. M. and J. T. Morris, 1991. Relative importance of ion exclusion, secretion and accumulation in *Spartina alterniflora* Loisel. Journal of Experimental Botany **42:** 1525–1532.

Brosemer, K. M., W. K. Nuttle, E. L. Winstead, J. S. Levine and W. R. Cofer, 1987. Denitrification activity and biogenic emissions of N_2O from wetlands: The effect of tidal action. Eos Transactions of the American Geophysical Union **68:** 1224.

Chalmers, A. G., 1982. Soil dynamics and the productivity of *Spartina alterniflora*. In: Kennedy, V.S. (ed.), Estuarine Comparisons, Academic Press, New York, pp. 231–242.

Chambers, R. M. and W. E. Odum, 1990. Porewater oxidation, dissolved phosphate and the iron curtain: Iron-phosphorous relations in tidal freshwater marshes. Biogeochemistry **10:** 37–52.

Dacey, J. W. H. and B. L. Howes, 1984. Water uptake by roots controls watertable movement and sediment oxidation in short *Spartina* marsh. Science **224:** 487–489.

Dingman, S. L., 1994. Physical Hydrology. Prentice-Hall, Englewood Cliffs, NJ, 575 pp.

Gardner, L. R. and M. Bohn, 1980. Geomorphic and hydraulic evolution of tidal creeks on a subsiding beach ridge plain, S.C. Marine Geology **34:** 91–97.

Gardner, L. R. and D. E. Porter, 2001. Stratigraphy and geologic history of a southeastern salt marsh basin, North Inlet, South Carolina, USA. Wetlands Ecology and Management **93:** 371–385.

Gardner, L. R. and H. W. Reeves, submitted. Seasonal patterns in the soil water balance of a *Spartina* marsh site at North Inlet, South Carolina, USA. Wetlands.

Gardner, L. R., B. R. Smith and W. K. Michener, 1992. Soil evolution along a forest-salt marsh transect under a regime of slowly rising sea level. Geoderma **55:** 141–157.

Gardner, L. R., H. W. Reeves and P. M. Thibodeau, in press. Groundwater dynamics along forest-marsh transects in a southeastern salt marsh, USA: Description, interpretation and challenges for numerical modeling. Wetlands Ecology and Management.

Giurgevich, J. R. and E. L. Dunn, 1979. Seasonal patterns of CO_2 and water vapor exchange of the tall and short height forms of *Spartina alterniflora* Loisel in a Georgia salt marsh. Oecologia **43:** 139–156.

Giurgevich, J. R. and E. L. Dunn, 1982. Seasonal patterns of daily net photosynthesis, transpiration, and net primary productivity of *Juncus roemerianus* Scheele and *Spartina alterniflora* Loisel in a Georgia salt marsh. Oecologia **52:** 404–410.

Hamon, W. R., 1961. Estimating potential evapotranspiration. Proceedings of the American Society of Civil Engineering, Journal of the Hydraulics Division **87(HY3):** 107–120.

Harvey, J. W. and W. E. Odum, 1990. The influence of tidal marshes on upland groundwater discharge to estuaries. Biogeochemistry **10:** 217–236.

Hemond, H. F. and R. Burke, 1981. A device for the measurement of infiltration in intermittently flooded wetlands. Limnology and Oceanography **26:** 795–800.

Hemond, H. F. and D. G. Chen, 1990. Air entry in salt marsh sediments. Soil Science **150:** 459–468.

Hemond, H. F. and J. L. Fifield, 1982. Subsurface flow in a salt marsh peat: A model and field study. Limnology and Oceanography **27:** 126–136.

Hemond, H. F., W. K. Nuttle, R. W. Burke and K. D. Stlzenbach, 1984. Surface infiltration in salt marshes: Theory, measurement and biogeochemical implications. Water Resources Research **20:** 591–600.

Howes, B. L., J. W. H. Dacey and D. D. Goehringer, 1986. Factors controlling the growth form of *Spartina alterniflora*: Feedbacks between above-ground production, sediment oxidation, nitrogen and salinity. Journal of Ecology **74:** 881–898.

Hughes, C. E., P. Binning and G. R. Willgoose, 1998. Characterization of the hydrology of an esturine wetland. Journal of Hydrology **211:** 34–49.

Keenan, R., J. Dickerson, L. R. Gardner and H. Reeves, 1996. Inexpensive electronic water level recorders for hydrologic studies. Ground Water Monitoring and Remediation (Spring), pp. 77–83.

Kjerfve, B., 1986. Circulation and salt flux in a well mixed estuary. In: J. van de Kreeke (ed.), Physics of shallow estuaries and bays, pp. 22–29. Springer-Verlag, New York.

Nuttle, W. K. and J. W. Harvey, 1995. Fluxes of water and solute in a coastal wetland sediment. 1. The contribution of regional groundwater discharge. Journal of Hydrology **164:** 89–107.

Nuttle, W. K., H. F. Hemond and K. D. Stolzenbach, 1990. Mechanisms of water storage in salt marsh sediments: The importance of dilation. Hydrological Processes **4:** 1–13.

Osgood, D. T., 2000. Subsurface hydrology and nutrient export from barrier island marshes at different tidal ranges. Wetlands Ecology and Management **8:** 133–146.

Osgood, D. T. and J. C. Zieman, 1998. The influence of subsurface hydrology on nutrient supply and smooth cordgrass (*Spartina alterniflora*) production in a developing barrier island marsh. Estuaries **21:** 767–783.

Penman, H. L., 1948. Natural evaporation from open water, bare soil and grasses. Proceedings of the Royal Society **193:** 120–145.

Priestly, C. H. B. and R. J. Taylor, 1972. On the assessment of surface heat flux and evaporation using large-scale parameters. Monthly Weather Review **100:** 81092.

Salisbury, F. B. and C. Ross, 1969. Plant Physiology. Wadsworth Publishers, Belmont, CA, 747 pp.

Thibodeau, P. M., 1997. Groundwater flow dynamics across the forest- salt marsh interface: North Inlet, South Carolina. Doctoral Dissertation, Department of Geological Sciences, University of South Carolina, Columbia, SC, 278 pp.

Thibodeau, P. M., L. R. Gardner and H. W. Reeves, 1998. The role of groundwater flow in controlling the spatial distribution of soil salinity and rooted macrophytes in a southeastern salt marsh, USA. Mangroves and Salt Marshes **2:** 1–13.

Tobias, C. R., J. W. Harvey and I. C. Anderson, 2001. Quantifying groundwater discharge through fringing wetlands to estuaries: Seasonal variability, methods comparison and implications for wetlands-estuary exchange. Limnology and Oceanography **46:** 604–615.

Valiela, I., J. M. Teal, S. Volkman, D. Shafer and E. J. Carpenter, 1978. Nutrient and particulate fluxes in a salt marsh ecosystem: Tidal exchanges and inputs by precipitation and groundwater. Limnology and Oceanography **23:** 798–812.