The effect of secondary circulation on the salt distribution in a sinuous coastal plain estuary: Satilla River, GA, USA

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1. Introduction

The Satilla River estuary is a relatively pristine salt-marsh estuary along the coastline of Georgia, USA (Fig. 1). Lying at the center of the South Atlantic Bight (SAB), the Satilla experiences some of the largest tides in the southeast US. The sinuous nature of the Satilla River estuary is characteristic of many of the estuaries along the SAB. The \( M_2 \) tide accounts for about 80% of the tidal energy in the Satilla (Seim et al., 2006). The tidal regime is consistent with a strongly convergent estuarine geometry. On most bends, the momentum core shifts from the inside to the outside of the bend. A strong (order 0.1 \( \text{m s}^{-1} \)) depth-averaged residual flow is produced at the bends forming counter-rotating eddies that meet at the apex of the bends. This type of residual flow is thought to be typical of sinuous, meandering estuaries (Dronkers, 2005).

Less well known is the structure and variability of the mass field in these systems. Importantly, a secondary circulation that takes the form of a helical flow pattern is generated in open channel flow around bends in unstratified systems (Smith and McLean, 1984). In stratified systems, the helical flow can be suppressed if \( \text{Fr}^2 \left( B/R \right) < 1 \), where \( \text{Fr} \) is a measure of vertical shear of the streamwise velocity \( u_z \), typically about 0.5, \( \text{Fr} = U_2 / g h \) is a Froude number, \( g \) is reduced gravity, \( h \) is water depth, \( B \) is channel width and \( R \) is the radius of curvature of the channel (Seim and Gregg, 1997). The strength and structure of the secondary circulation in estuaries responds to the neap–spring cycle and to river discharge (Chant, 2002). A helical flow pattern can be established during spring tide with flow to the outside of the curving channel at the surface and flow to the inside near bottom. This pattern can change to a two-cell structure, one cell stacked over the other, during smaller tides (Chant, 2002; Elston, 2005) with secondary flow due to channel curvature confined to the upper cell. Because the secondary circulation has the potential to overturn the water column its presence or absence may significantly impact mixing and exchange in an estuary.

The impact of curvature-induced secondary circulation on the salt balance in an estuary remains largely unexplored. Most observations and models of exchange in straight estuaries indicate the importance to the salt balance of runoff and baroclinic pressure gradients (e.g. Uncles et al., 1986; Wong, 1994). Landward flow is found in the thalweg and seaward flow in the shallows. The opposite configuration is found in barotropic
models (Li and O’Donnell, 1997). There have been studies to explore how these two modes operate together in shallow tidal channels (Li et al., 1998; Ralston and Stacey, 2005). In these simplified models the details of exchange are governed by the Ekman number $E_k = A_2/(fH^2)$ (Valle Levinson et al., 2003) which measures the relative importance of rotation based on the strength of vertical mixing ($A_2$), depth ($H$), and Coriolis effect ($f$). Using typical ranges of vertically averaged axial velocity ($U$) throughout the estuary, $C_o = 0.002$ (Seim et al., 2002), and with $A_2 = 0.05 \sqrt{C_o U}$, $A_2$ varies from 0.009 m$^2$ s$^{-1}$ at neap to 0.015 m$^2$ s$^{-1}$ at spring. The corresponding values for $E_k$ are 1.5 at neap and 2.4 at spring, indicating that frictional forces can be expected to dominate over Coriolis effects in the Satilla River estuary.

These formulations do not explicitly include the role of secondary circulation, however. Secondary circulation may decrease dispersion of salt landward (Smith, 1976; West et al., 1990). Some evidence of the impact of secondary circulation on the salt balance has been seen. Focusing attention on a single channel bend, the along-channel density gradient in the thalweg was found to be weakest during spring tide (Seim et al., 2002), suggesting that the salinity intrusion could be greatest at spring tide. At neap, vertical stratification was strong enough to raise the gradient Richardson number well above 0.25 during ebb, consistent with tidal straining. The transition was related to activation and suppression of helical circulation around the bend.

In contrast to the focus on tidal circulation by Seim et al. (2006), this paper describes the subtidal circulation, the subtidal salt field variations, and the intrusion of salt into the estuary. We provide a conceptual model to explain the observed variability. The elements of the conceptual model include a link between channel curvature and the secondary circulation is generates, which causes a tilt in the free surface at the bends, and the residual tidal eddies reported briefly by (Seim et al., 2006), which can be considered a tertiary flow. The overturning circulation at the bends promotes a horizontal salinity contrast that is carried along-estuary by the residual tidal eddies. The geometry is such that the seaward flow follows the deep channel of the estuary and proceeds largely unobstructed, whereas the landward flow follows a shallower channel and is prone to strong mixing over shoals. This circulation appears to weaken the dependency of the salt intrusion on fluctuations in river discharge because it allows a pulse of river discharge to rapidly push saline water seaward. However, salt moves landward significantly slower.

After outlining the field program and analysis performed, we describe the observed circulation features including the spring–neap differences in the structure and strength of the observed residual tidal eddies. Next we describe details in the salinity field and how its longitudinal and lateral distribution changes in response to secondary circulation and the neap–spring cycle. Finally, we discuss how the observed exchange flow in counter-rotating eddies affect how salinity intrudes into the estuary in response to fluctuations in river discharge.

2. Data description and analyses

We undertook two 2-month long field efforts in the Satilla River estuary during 1999, designed to sample high (spring) and low (autumn) river discharge conditions. In each case two sampling schemes were employed: a series of along-channel moorings, deployed for the full 2-month period, and several intensive small-boat sampling exercises. The small-boat sampling strategy was designed for limited reaches of the estuary to yield a picture of the three-dimensional variability of the circulation and density field over a tidal cycle. This effort covered four adjacent sections of the estuary (Fig. 1), once at spring tide and once at neap tides. We refer to the these intensive efforts as roving samples. These eight intensive sampling periods were split to coincide with neap and spring tides of the high and low river discharge mooring programs. Additional temperature, salinity, and pressure data were collected at two long-term moorings, one upriver from the mouth of the Satilla River at LT1 and one near river kilometer 38 at LT2.

2.1. Salinity, pressure, and velocity moorings

The mooring array used Teledyne RD Instruments, Inc. (RDI) ADCPs and InterOcean S4 current meters to measure velocity and Seabird Electronics, Inc. (SBE) sensors to measure temperature, salinity, pressure, and optical backscatter (OBS) (Table 1). As a result, in some locations full-depth velocity profiles are available, while at other locations currents are observed at a single depth. Moorings consisted of a pyramid-shaped stainless steel frame placed at each location on which a subsurface (0.5 mab) SBE SeaCAT CTD and either an InterOcean S4 current meter or an upward-looking RDI Workhorse ADCP were mounted. A SBE MicroCAT CTD was tethered with 40 ft of stainless steel cable to each pyramid to measure surface temperature and salinity.

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**Fig. 1.** Satilla River morphology showing the large intertidal areas typical of estuaries along the southeastern US. The locations of current, sub-surface pressure and salinity monitors during SAT1 and SAT2 are shown as white circles. Distances of stations from the ocean are given in Table 1.
Temperature, salinity, pressure, and OBS were recorded at 6 min intervals by the SeaCAT and MicroCAT CTDs while the ADCPs measured depth, velocity, and acoustic backscatter at 12 min intervals starting at 1.24 mab to 0.5 m below the water surface. Shipboard DGPS and in situ fathometry were used to determine the exact locations and depths of the mooring placements. See Elston (2005) for additional information on the instrument configuration at each mooring.

Two long-term monitoring stations were in place during this study. Data at LT1 will be shown for a 9-month time series (March–December 1999) of temperature, salinity, and pressure at a 6-min interval. Data collected at LT1 were used to monitor and relate changes in temperature, salinity, and pressure over the monthly, seasonal, and annual cycles to changes in freshwater discharge. LT1 pressure data were used to adjust the depth component of the moored CTDs and moored ADCPs to the mean lower low water (MLLW) datum. LT2, located farther upriver, functioned in a similar manner, and a 2-month record (September–November 1999) will be shown.

Salinity data from the moored SeaCAT and MicroCAT CTDs were of excellent quality except for Station 4 during the autumn 1999 mooring deployment. Antifoulants significantly reduced biofouling during mooring deployments. Due to rough weather and a highly corrosive environment, several of the surface-tethered MicroCAT CTDs broke free approximately 20 days after the initial deployment thus compromising most of our surface salinity measurements at the moorings. Temperature and pressure data for the SeaCAT CTDs and the MicroCAT CTDs were of excellent quality for both the spring and autumn 1999 mooring deployments.

The salinity fluctuations recorded during autumn 1999 at Station 4 (main channel) had tidal fluctuations that were too small, suggesting some clogging of the conductivity sensor. In order to better simulate the salinity for Station 4, the data at Station 4 were substituted with Station 5 salinity to which a constant value of 7.5 PSU was added, based on mean downstream salinity data from the SBE-21 unit by documenting cross-channel changes in vertical stratification.

Roving ADCP observations were analyzed spatially on a latitude/longitude grid following Seim et al. (2006) and on latitude/depth grids following Elston (2005). For the latitude/longitude grid analysis, the processed, 15-s averaged roving ADCP observations were analyzed on a 200 m grid. At each location in the sampling grid, a simple least-squares fit was used to estimate the mean, tidal current ellipse parameters at semi-diurnal and quatra-diurnal frequencies, and an error term. The fit was performed for a grid cell if there were a minimum of seven clusters of observation times over the tidal cycle. Often there was more than one 15-s average in a grid cell from a circuit, in which case multiple current estimates were included. Typically there were 9 or 10 sets of current observations per grid point over the 13 h sampling period. Root-mean-square errors average 1.6 cm s\(^{-1}\) during spring tides and 1.3 cm s\(^{-1}\) during neap tides. Differences in the spatial coverage between spring and neap tide surveys occurred because some shallows became impassable at spring tide low water. The most dramatic example was near mooring 3, where we were unable to adequately sample the shallower channel to the west of the marsh island.

For the latitude/depth grid analysis, raw, 2-s averaged irregularly spaced roving ADCP observations were reduced to the MLLW datum and interpolated onto a regular grid with 8-m horizontal and 0.5-m vertical bins. Data in a given cross-section were not demeaned, so that absolute residual values could be calculated. Next, at each survey location, the laterally gridded cross-section was rotated into along and cross-channel components by using the tidal spatially averaged depth-averaged transport. Last, for each cell in the regular latitude/depth grid with at least five velocity observations over the tidal cycle, a simple least-squares fit was used to estimate the mean, tidal current ellipse parameters for the semidiurnal and quadradiurnal frequencies, and an error term. The rms values of the errors are, respectively, on average 0.9 cm s\(^{-1}\) at spring and 0.4 cm s\(^{-1}\) at neap tides. Because of the difference in the spatial area covered between spring and neap tides and the minimum five observations required per sample cell fit, the lateral coverage in a given section varied between surveys. The intertidal area was excluded from the resulting tidally averaged cross-sections, and the number of fit boxes at spring tide was reduced from that at neap tide by approximately \(\sim 12\%\).

### Table 1

<table>
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<tr>
<th>Station</th>
<th>LT1</th>
<th>3a</th>
<th>4a</th>
<th>5</th>
<th>6</th>
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<td>1100</td>
<td>900</td>
<td>900</td>
<td>500</td>
<td>200</td>
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</table>

Station 7 is in White Oak Creek (WOC). Distance is measured from the mouth of the estuary. Station depth was calculated from bottom pressure using UNESCO (1985) protocol. LW and HW widths were calculated from regional GIS data by Alice Chalmers, University of Georgia. Parameter symbols are: \(P\) = subsurface pressure from SeaBird sensors; \(CB\) = bottom current from InterOcean S-4; \(CP\) = current profile from RDI ADCP workhorse.
other three stations were positioned where seaward flow occurred from residual tidal eddies (described below). The strength of the subtidal flow was modulated by the spring/neap cycle, but did not reverse direction. The two exceptions occurred at Stations 3 and 4a. Subtidal flow at Station 3, opposite the cusp of a large meander, had episodes of landward flow near the time of neap tide, as previously observed (Seim et al., 2002; Blanton et al., 2003). Station 4a, located in a side channel, had consistently landward flow. The fluctuations in flow at Station 4a were remarkably well correlated with those at Station 3 (Fig. 2a).

The magnitude of the observed subtidal velocities significantly exceeds that due to river discharge. A rough estimate of the momentum carrying cross-sectional near station 4, about mid-estuary, is 3.5 m deep times 1200 m across or \( A = 4200 \text{ m}^2\). Discharge during the fall period shown in Fig. 2 was \( Q = 50 \text{ m}^3 \text{s}^{-1}\) or less, and therefore the downstream velocity due to river flow is 0.01 \text{ m}s\(^{-1}\) or less. Even during the spring freshet when discharge peaked at 150 \text{ m}^3 \text{s}^{-1}, \( Q/A\) is only 0.03 \text{ m}s\(^{-1}\) and its signature is difficult to perceive in the observed subtidal velocities (not shown).

Station 3 is located near the outside shore of a channel meander where channel curvature is maximum. Given the link between channel bends and the position of residual tidal eddies (see below), the landward flow present at this station suggests minimal influence of eddies at this location.

### 3.2. Lateral flow differences

Tidally averaged maps of the vertically averaged flow field (Fig. 3) show an array of residual eddies situated between the channel meanders (Seim et al., 2006). The eddies are situated on either side of prominent cusps on the inside bend of the channel meanders. The counter-rotating flow of these eddies has been cited as the cause of the sharp cusps on the inside bend of meanders in many estuaries (Dronkers, 2005).

The tidally averaged axial flow through several cross-sections conform to the presence of the tidal eddies and illustrate the combined effect of density-driven along-channel exchange and channel curvature on the cross-sectional distribution of the flow (Fig. 4). While recent models (e.g. Valle Levinson et al., 2003) indicate landward flow in the main (deeper or ebb) channel and seaward flow on the side (shallower or flood) channel, this picture is altered by the curvature-driven residual flow between channel meanders. The most seaward cross-section (A36, lower right) was situated in the large meander in the vicinity Station 3 and its surface flow was within a clockwise eddy (partially shown in Fig. 3) downstream of the cusp. This flow overlayed landward subsurface flow which was strengthened substantially during neap where it spread across much of the channel.

To the west around the bend (A45, middle right) a counter-clockwise eddy promoted seaward flow near the inside of the bend and augmented the landward flow towards the outside of the bend, here filling most of the water column in the deeper channel. Note the spreading of landward flow into shallower areas at neap.

The third panel (A910, upper right) is located where the channel begins to straighten and bottom depths are shallower and more uniform. At spring tide, the flow was clearly separated into two parts: landward flow throughout the water column on the north side of the channel and seaward flow on the south side. The landward flow spread laterally at neap to occupy the lower half of the water column on the south side of the channel where there was seaward flow present at spring tide.

The last two cross-sections (Fig. 4) are located near the cusps of the next two meanders farther upstream. The first section (B56,
lower left) was situated near the downstream edge of a clockwise rotating residual eddy. The flow on the south side of the channel was landward. This result is supported by current data at Station 4a which also indicated persistent landward flow in this channel. Seaward flow was measured in the deeper channel on the north side, consistent with the time series measurements at Station 4 (Fig. 2). At neap, seaward flow seemed to overtop the landward flow in the southern channel.

The final section (B1011, upper left) is close to the next cusp where seaward flow along the south side to the west of the cusp crossed over from the south to the north shore. During spring tide, landward flow is confined to the northern half of the section. During neap, landward flow occupied the entire lower layer across much of the channel and seaward flow occupied almost the entire channel in the surface layer.

The structure of the tidally averaged exchange flow appears consistent with the residual tidal eddies depicted in Fig. 3. Rather than conform simply to landward/seaward exchange in deep/shallow channels (Valle Levinson et al., 2003), the lateral structure of the exchange flow appears to be governed to a large extent by the residual tidal eddies (Dronkers, 2005). The additional vertical stratification present at neap tides allows the landward flow to spread laterally along the bottom so that it occupies more of the cross-section than it does at spring tide but is not capable of altering the depth-averaged flow pattern shown in Fig. 3.

3.3. Axial distribution of salinity

Salinity was monitored at several stations (Fig. 5) along the Satilla estuary. Freshwater discharge to the estuary occurred as three main events. The first reached 150 m$^3$ s$^{-1}$ in February, consistent with the climatological maximum in winter. The other two were related to the passage of two tropical cyclones on the continental shelf.

The salinity response to discharge is clearly seen in the along-channel salinity time series (bottom panel of Fig. 5). Note the response of salinity to the discharge curve: salinity decreased rapidly by up to 15 PSU over a 20-day period in response to the February discharge event while the recovery period lasted 70 days. A similar rapid decrease in salinity was seen as a response to the October discharge event. During both discharge events the salinity response was quite pronounced at Stations 3–7 but was much less obvious at Station 2. Station 2 was located in an area of more open water (St. Andrews Sound) and suggests that a fundamental change in the dynamics may occur somewhere between Stations 2 and 3. The muted response of the salinity at LT1 to the July discharge event may indicate the change in dynamical regime is landward of its location (between Stations 2 and 3).

The axial–vertical distribution of salinity, as seen in high water surveys from a wide range of conditions, shows changes in the salt regime typical of the response of a partially mixed estuary to
changes in freshwater discharge (Fig. 6). At ~25 km from the ocean the salinity in the estuary varied between 0 and 22 PSU, and surface-to-bottom salinity difference at high water varied between ~1 and ~3 PSU throughout the year. It is significant that stratification exists at the end of flood, implying that tidal straining during each flood tidal cycle is unable to mix away the...
stratification, and is borne out in the individual profiles collected in the deep channels as part of the roving surveys.

The spring and autumn salinity regimes are compared by plotting against river discharge the distance of the 10-PSU isohaline from the ocean and the axial salinity gradient between three station pairs (Figs. 7 and 8). During spring, the 10-PSU line responded to the freshet by moving ~7 km seaward. Most of this excursion was accomplished in about 3 days, and 10-PSU water remained about 16 km from the ocean for the next 25 days, even though discharge decreased almost five-fold. The landward excursion during Days 68–72 was not accompanied by any obvious response of the salt intrusion to the spring/neap cycle.

The maximum $dS/dx$ observed in autumn was positioned between Stations 4 and 5, about 6 km farther landward than in the spring and was about $1.0 \times 10^{-3}$ PSU m$^{-1}$ before the freshet. Afterwards, $dS/dx$ increased to $1.5 \times 10^{-3}$ PSU m$^{-1}$ which was about equal to the maximum gradient observed in spring. The strength remained at this level even as discharge began to decrease to the same magnitude observed in late spring. The gradient between the seaward pair also increased in response to the freshet. Again, there is no obvious response of the salt intrusion to the spring/neap cycle.

In summary, the maximum observed $dS/dx$ was about $1.5 \times 10^{-3}$ PSU m$^{-1}$ for discharges between 50 m$^3$ s$^{-1}$ (spring) and 150 m$^3$ s$^{-1}$ (autumn) and $150$ m$^3$ s$^{-1}$ (spring). Freshets during both seasons, even though they were of different magnitudes, moved the 10-PSU isohaline about 10 km seaward. The gradient stayed remarkably constant in spring even during the low discharge periods at the beginning and end of the spring observations. At the beginning of the autumn study period, there was a smaller salinity gradient of $1.0 \times 10^{-3}$ PSU m$^{-1}$ but $dS/dx$ increased to $1.5 \times 10^{-3}$ PSU m$^{-1}$ during and after even this small freshet. A consistent and surprising result is that the near bottom salinity intrusion does not exhibit a response to the spring/neap cycle of variation in subtidal circulation.

3.4. Cross-sectional distribution of salinity

Though not as complete as for the velocity field, a representation of the tidally averaged salinity was constructed based on vertical salinity profiles near the channel sides and continuous surface salinity across the channel (Elston, 2005). The tidally averaged salinity is superimposed on the tidally averaged along-channel velocity sections in Fig. 4 to demonstrate the relationship between the flow field and mass field. During neap tides, top-to-bottom difference in salinity in the halwag was approximately 10 PSU. Moreover, the water was stratified even in the shallows by 2-3 PSU over depths less than 4 m. During spring tide, weaker stratification of about 1.5 PSU was found in the halwag. The difference decreased to almost zero in the shallows, but there was a significant lateral salinity change of ~2 PSU across the channel. Thus, the spring tide regime was characterized by lateral salinity gradients with vertical gradients confined to the main channel. During neap, the salinity regime shifts; significant lateral salinity gradients across the channel remain but strong vertical gradients are present in both deep and shallow parts of a cross-section. The cross-channel distribution of salinity is influenced by the curvature-induced secondary circulation. At neap tides the seaward flow, always surface trapped, carries the freshest water in the cross-section, switching from side to side of the channel as it moves downstream (Fig. 4). During spring tides surface water is significantly saltier, little vertical stratification exists, and lateral
salinity differences persist. Unlike neap tides when salinity differences of 5 are common, salinity variability in a cross-section at spring tide is considerably less.

The best spatial representation of the salinity field can be created for the surface from the roving samples. Removing the mean along-channel salinity gradient from the surface salinity accents cross-channel differences (Fig. 9). While the data are noisy, tidally averaged cross-channel differences in salinity reach as high as 1 at spring tide and 2 at neap. The cross-channel salinity gradient changes sign, depending upon the sense of curvature in a meander. Minimum surface salinity is most often observed in the main channel, consistent with the circulation pattern represented by the counter-rotating residual eddies. Maximum surface salinity is observed in the shallower side channels though the sampling of this branch of the circulation is more limited because the shallows prevented sampling throughout the tidal cycle at these sites.

The vertical structure of the salinity is most readily appreciated as a series of profile plots which show the tidal cycle average salinity profile for each station (Fig. 10). At each cross-section there are profile stations on either side of the meandering channel, where vertical profiles were collected over a tidal cycle. The southern station is plotted in red, the northern station in blue. Note that the range of salinity shown in each subplot increases in the seaward direction. It is uncommon to observe in the instantaneous profiles the same salinity profile at the 2 sites in a cross-section; in a tidal cycle average this results in a nearly constant offset of up to 3 PSU over the entire water column as typical. The offset between stations in a given cross-section tends to be the smallest at the surface during spring tides, indicating the surface plots (Fig. 9) may under-represent the lateral structure during large tidal range.

Greater salinity can be either on the north or south side of the channel and can be related to the residual eddy field induced by the channel bends (Fig. 3). The sections with landward residual flow (flood) channels to the south (A36, A45, B56, and B78) consistently indicate greater salinity in the side channels. Those sections with seaward residual flow (ebb) channels to the south (A910, C34 and C910) tend to show lower salinity in the main channel. The occurrence of higher salinity near the bottom in the main channels is consistent with curvature-induced secondary circulation. Stations B1011 and C12 appear to be on top of the transition zone between clockwise and counterclockwise rotating residual eddies as they show both types of behavior.

Vertical stratification is much more pronounced at neap tides, even in water as shallow as 3 m. For the sections in domain A vertical gradients in tidally averaged profiles at neap exceed 1 m$^{-1}$ and display little indication of any boundary layers; individual profiles exhibit in some instances thin bottom boundary layers (<2 m) and remarkable salinity structure—in some cases 8 PSU change in 5 m of water. Vertical stratification is greatly reduced at spring tides; individual profiles are characterized by large regions of well-mixed fluid and well mixed profiles are common for those collected in less than 5 m depth. There is also a clear tendency for stratification to decrease moving upstream. It is worth noting that the tidal current amplitude increases in the upstream direction.
Fig. 7. The intrusion distance of the 10-PSU isohaline and the axial salinity gradient compared with spring discharge of the Satilla River. The dotted line in the second graph is based on the relationship of Monismith et al. (2002). The bold line for the gradient is calculated from salinity at Stations 3 and 4. The dash-dash and dash-dot lines are calculated from salinity data at Stations 2 and 3 and Stations 4 and 5, respectively. See Fig. 1 for station locations.

Fig. 8. The intrusion distance of the 10-PSU isohaline and the axial salinity gradient compared with autumn discharge of the Satilla River. The dotted line in the second graph is based on the relationship of Monismith et al. (2002). The bold line for the gradient is calculated from salinity at Stations 4 and 5. The dash-dash and dash-dot lines are calculated from salinity data at Stations 2 and 4 and Stations 5 and 6, respectively. See Fig. 1 for station locations.
over the area under study (Seim et al., 2006) and that the decreasing stratification may reflect the increased tidal energy available for mixing.

When including all cross-sections and calculating the average horizontal and vertical density difference over all tidal cycles sampled, this difference amounts to 1 kg m\(^{-3}\) or greater throughout the first 25-km of the estuary. The mean differences are the greatest during neap tides in the lower reaches of the estuary that were sampled. The ratio of horizontal to vertical density differences is near unity on neap tides but doubles on spring tide. These observations indicate a persistent cross-channel density structure that is increasingly pronounced at spring tides.

4. Discussion

Data presented here suggest that channel curvature can strongly influence the circulation field, mass field, salt flux, and
salt balance in an estuary. A thorough investigation of the nature of each of these influences is beyond the scope of this effort, but a cursory examination of the dynamics involved in each of these processes is undertaken below.

Considering first the circulation field, the cross-channel surface gradients associated with the curving flow drive a tidally averaged barotropic along-channel residual circulation between the bends that creates the counter-rotating eddies between channel bends. The magnitude of the surface slope can be estimated as a steady balance between the cross-channel surface slope and the centrifugal acceleration:

$$\frac{\partial \eta}{\partial y} \approx \frac{\eta}{g R}; \quad \Delta \eta \approx \frac{g \pi^2}{g R}$$
where $R$ is the radius of curvature of the bend and $\rho$ is the surface elevation. Using a channel width $B$ of 1000 m, $R = 1000$ m, and $U = 0.5$ m s$^{-1}$ (a reasonable rms value) gives $\Delta v = 4$ cm. The magnitude of cross-channel setup is large enough to be observed with conventional sensors and should be readily confirmed.

The residual along-channel current driven by the cross-channel setup can be considered a balance between along-channel surface slope and bottom friction:

$$\frac{\partial q_y}{\partial t} + \frac{\partial q_y}{\partial x} = \frac{g}{h} \delta \zeta; \quad \tau_b = C_d U^2$$

Assuming the sections between bends are 2 km long, the mean depth $h = 4$ m and a drag coefficient of $C_d = 0.002$ (Seim et al., 2002) suggests the residual flows should be roughly 10 cm/s, consistent with the observed values. Obviously, the magnitude of the flow will be modulated by variations in tidal amplitude, but this should be readily re-produced in numerical simulations that adequately resolve the cross-channel flow and the channel geometry. Thus, the flow features of the residual eddies require explicit recognition of channel curvature (centrifugal accelerations) and cross-channel variations in flow.

With regard to the impact of the curving flow on the mass field, we observed persistent cross-channel salinity differences of 2–4 PSU and persistent vertical stratification. The latter is surprising considering the shallowness of the system and the strength of the tidal currents. One possibility is that the cross-channel salinity differences and cross-channel flow induced by the channel bends help maintain the vertical stratification. We estimate the importance of cross-channel processes through examination of a simplified stratification equation (neglects vertical advection and lateral dispersion of stratification)

$$\frac{\partial q_s}{\partial t} + \frac{\partial q_s}{\partial x} = \frac{g}{h} \frac{\partial \delta \rho}{\partial x} - \frac{\partial q_x}{\partial x}$$

where $N^2 = -g / \rho \cdot \partial \rho / \partial z$ is the buoyancy frequency and $K_z$ is the vertical eddy diffusivity assumed to be equal to the vertical eddy viscosity. We observe persistent stratification, suggesting a quasi-steady balance between production and destruction; assuming the first term vanishes, the destruction of stratification embodied in the last term may be balanced be either one or both of the remaining terms on the left hand side. The magnitude of each term is estimated from observations as follows: tidally averaged $u,v \approx 0.1$ m s$^{-1}$; $\partial \rho / \partial x = 10^{-3}$ kg m$^{-4}$; $N^2 = 10^{-2}$ s$^{-2}$; $K_z = 10^{-2}$ m$^2$ s$^{-1}$. Assuming a vertical length scale of 5 m, the cross-channel production term is $2 \times 10^{-2}$ m$^3$ s$^{-1}$ whereas the destruction term is $4 \times 10^{-4}$ N m$^{-3}$. The scaling suggests gravitational circulation alone is unable to maintain the observed stratification. Assuming a cross-channel length scale of 500 m, $\partial q_s / \partial y = 2$ kg m$^{-3}$ per 500 m = $4 \times 10^{-3}$ kg m$^{-4}$, the cross-channel production term is $10^{-6}$ m$^3$ s$^{-1}$, the same order of magnitude as the destruction term and a factor of five greater than the along-channel production term. Hence it appears that the variability and structure of the mass field is largely controlled by the channel curvature.

4.1. Effects of residual eddies on the axial salinity gradient

In the residual eddies the landward flows are found to be correlated with higher salinity and seaward flows with lower salinity that should provide an efficient mechanism for moving salt upstream. The cross-channel salinity difference of 2–4 PSU combined with the residual currents, estimated at 0.1 m s$^{-1}$, suggest upstream salt flux $\frac{\partial s}{\partial t}$ driven by this mechanism alone should be roughly 0.2–0.4 PSU m$^{-1}$ s$^{-1}$. The changing orientation of the bends generates a pattern of counter-rotating residual eddies and the communication between eddies requires lower and higher salinity water to switch to opposite sides of the channel. The communication between eddies occurs at the bends themselves. The existing observations allow a qualitative description of this flow feature, but do not lend themselves to calculation of the salt flux on the bends; we therefore make some inferences about the overall characteristics of the salt balance through the use of simple formulations then relate these to the existing observations.

4.2. Longitudinal dispersion

Steady-state theory for estuarine salt exchange holds that salt advected seaward by river discharge is balanced by upstream dispersive processes; a wide variety of decompositions of the processes responsible for the upstream flux have been proposed. There is also recognition of the time-dependence of the salt balance and its implications for estimating the importance of the various processes (Banas et al., 2004). We here take a very simple approach to calculate a representative axial dispersion coefficient under the range of axial salinity gradients observed in the Satilla; obviously, much more sophisticated approaches are possible. The stability of the salinity intrusion length over month-long time scales during decreasing discharge is assumed to justify a steady balance, and we choose to define a single horizontal dispersion coefficient to represent all processes responsible for the upstream movement of salt. If the influence of time dependence is similar to that seen in Banas et al. (2004) then we would expect the range of estimates of diffusivity based on a steady-state balance to be greater than estimates made that include time dependence. This balance is succinctly expressed by the following simple one-dimensional model:

$$K_x = \frac{K_s}{A \delta S / \delta x}$$

where $R$ is river discharge, $S$ is the salinity in the domain, and $\delta$ is the cross-sectional area. We use data from sections in the “A” and “B” domains (Fig. 4), collected when $R = 20$ m$^3$ s$^{-1}$, and apply Eq. (1) to the results (Table 2). The “C” sections were collected under extremely low discharge ($R = 5$ m$^3$ s$^{-1}$), and the axial salinity gradient was the same order as the neap-tide value during springtime sampling, producing $K_s = 30$ m$^2$ s$^{-1}$. The axial salinity gradient ($\delta S / \delta x$) was smaller at spring tide, resulting in $K_s \approx 100$ m$^2$ s$^{-1}$. During neap, $K_s \approx 50$ m$^2$ s$^{-1}$, roughly one-half the value at spring tide. The range observed here (30–120 m$^2$ s$^{-1}$) is consistent with other studies (see, for example, West et al., 1990; Ralston and Stacey, 2005).

### Table 2

<table>
<thead>
<tr>
<th>Section</th>
<th>X (km)</th>
<th>Area (m$^2$)</th>
<th>Salinity</th>
<th>$S_x$ (m$^{-1}$)</th>
<th>$K_x$ (m$^2$ s$^{-1}$)</th>
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</thead>
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<td></td>
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<td></td>
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</tr>
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<tr>
<td>C910</td>
<td>24</td>
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<td>14.6</td>
<td>–</td>
<td>–</td>
</tr>
</tbody>
</table>

Sections in the A and B domain were done during a discharge of 20 m$^3$ s$^{-1}$, while those in the C domain were done during a discharge of 5 m$^3$ s$^{-1}$. “X” is distance from the ocean. Sections “A” and “B” were done at spring and neap tide between 10 and 20 March 1999. The “C” section was done during a spring tide on 10 September 1999. Hurricane Floyd prohibited obtaining neap tide sections for comparison. See Fig. 4 for location of sections.
An estimate of the total upstream salt flux is simply $\mathcal{F} = K_x \frac{dS}{dx}$. For the range of axial salinity gradients observed, estimates are 0.015–0.18 PSU m$^{-1}$. Given the magnitude of the salt flux driven by the residual eddies alone (estimated as 0.2–0.4PSU m$^{-1}$ above) it would appear that the upstream salt flux is limited by processes acting at the bends in the crossover regions.

A possible explanation for the blockage of upstream salt flux at the bends is connected to the morphology surrounding the cusps of meanders, analogous to that found in meandering rivers (Ritter et al., 2002). The bends exhibit a specific geometry of detached point bars downstream of the cusp of the bend that merge with mid-channel bars which separate the channel into a main and side channel. Point bars at the cusps are displaced seaward, and these bars constrict the width and depth of the flood channels. This geometry favors the transfer of lower salinity water between eddies because it is surface trapped and because it follows the main channel of the estuary. It discourages transfer of higher salinity water between eddies because the flow is bottom trapped and must leap-frog over the point bars and because it follows the shallower side channel that will induce more vertical mixing.

The blockage of upstream flow occurs on a scale presumably limited to the eddy size ($\frac{1}{4}$ the meander wavelength), in this case $\sim 2$ km along the axis of the estuary. A portion of the salt is simply recirculated on this scale and lowers the amount of salt that can be passed farther upstream. We suggest that recirculation of salt in the residual eddies accounts for the relatively slow increase of salinity after an episode of river discharge. In other words, while an event of river discharge rapidly pushes salinity downstream, the return of salinity as discharge slackens is significantly slower in meandering channels of estuaries where channel curvature induces the recirculating eddies (Figs. 7 and 8).

4.3. Salt intrusion

The observed ranges of salt intrusion as defined by $X_s$ for the 10-PSU isoline are similar whether for the spring (Fig. 7, high discharge) or fall (Fig. 8, low discharge). While the salt intrusion distance ($X_s$) is theoretically related to $Q$ by $X_s \sim Q^{-\alpha}$ where $\alpha = \frac{1}{4}$ (Hetland and Geyer, 2004), data in San Francisco Bay yielded a smaller exponent of $\alpha = \frac{1}{5}$ (Monismith et al., 2002) suggesting a significantly weaker dependency of salt intrusion on river discharge. We have added to the intrusion sub-panels the equation $X_s = kQ^{-\alpha/2}$ with $k = 40$ for the Satilla. This constant results from a regression of log $X$ on log $Q$ for all collected mooring data and represents the approximate distance in kilometers of salt intrusion when $Q$ approaches zero.

There is a reasonably close correspondence to the equation when the 10-PSU isoline is pushed seaward during a discharge impulse. But the relationship overestimates the return landward of this isoline, suggesting that the circulation impedes its rate of return landward. Thus, the asymmetry is described by the observation that pulses of discharge move salt rapidly seaward, but that the salt moves landward at a relatively slow rate when discharge decreases.

There is no clear response of intrusion distance to the neap–spring cycle. Tidally averaged cross-sectional flow (Fig. 4) does indicate a stronger gravitational mode at neap tide. Note the single event around Days 69–73 when discharge was steady and low (Fig. 7). The 10-PSU isoline travelled landward about 5 km during the smallest neap tide of this study period when the secondary circulation weakened to a level insufficient to provide overturning. Thus, at times when secondary circulation is sufficiently small, we expect the gravitational mode to set up and efficiently push salt landward. But as a rule, the secondary circulation in channels of such strong curvature is robust enough to set up spatially trapped eddies that efficiently transport water and salt seaward and less-efficiently transport water and salt landward.

Related to the lack of response of salt intrusion to the spring–neap cycle, the time scales of adjustment for the Satilla—estimated above as 20 days in response to an increase in discharge and 70 days to return to pre-freshet conditions—are long given the scales of the system. MacCready (1990) suggests that in the diffusive limit the adjustment time varies at $K_x \tau^2 = 10^5$ s, of order 10 days for the Satilla. Other estimators yield similar values. The slow response of the system to change is consistent with the notion that poor communication between the recirculation features in the main body of the estuary limits the landward flux.

Lastly, it appears that the series of channel bands in the Satilla River estuary act to trap the salinity gradient. The region of maximum salinity gradient was consistently found in the region of the bends over a range of discharge with weaker gradients both landward and seaward. Interpreting the along-channel variations in the salinity gradient in terms of varying horizontal dispersion suggests the upstream movement of salt is a minimum in the region of the bends. Seaward of the bends the channel broadens rapidly and the topography becomes more complicated. This region was much less responsive to increases in river discharge and more akin to an open sea with more random dispersive processes. Landward of the bends the channel narrows and deepens and is characterized by a single depth maximum. Both horizontal and vertical salinity gradients weaken, more in line with predictions based on the geometric scales. This form of estuary geometry is reasonably common in undisturbed (i.e. not dredged) estuarine systems along the southeast US seaboard and may act as a natural buffer to salinity intrusion.

5. Summary and conclusions

Our findings lead to a conceptual model that takes into account the dynamics induced by channel curvature found in sinuous channels of tidally forced shallow coastal plain estuaries. Tidal flow in the Satilla around curving channels raises the water level opposite the cusps of channel meanders and generates a series of counter-rotating subtidal tidal eddies with convergent flow at the cusps and divergent flow opposite (Fig. 3). The secondary circulation at the bends tends to separate density classes across the estuary; together with the counter-rotating eddies between bends these flows promote and maintain lateral density gradients in the central estuary. The geometry of the system is such that where the system is wide enough to host multiple channels, the surface-trapped (because of its reduced density) subtidal seaward flow follows the deep channel and moves towards the sea relatively unimpeded. The bottom-trapped (because it is dense fluid) landward flow follows the shallower channel and is prone to strong mixing at the cusps of the bends where its channel shallows owing to mid-channel shoals that nearly attach to the meander cusps.

The circulation of the conceptual model described above is enhanced at spring tide when secondary circulation is strongest. Weakened residual eddies and secondary circulation at neap tide allows the setup of a stronger gravitational mode of estuarine circulation that is surprisingly well developed even in the shallows. Scaling the stratification equation indicates that such strong stratification cannot be maintained by the gravitational circulation alone, but requires the cross-channel production of stratification largely controlled by channel curvature. A stronger gravitational mode at neap is evidenced by the spread of landward flow beyond the confines of the deeper parts of the channel. Even
during prolonged periods of low river discharge, high salinity water resists travelling inland over the normal energy range of spring–neap tides (Figs. 7 and 8). Only during the weakest neap tides is there some evidence that salt travels farther inland (Fig. 7) in an enhanced gravitational flow. Thus, the role of secondary (cross-channel overturning) and tertiary flow (bend-scale horizontal eddies) in this meandering shallow mesotidal estuary is to retard the normal tidal mixing of higher salinity oceanic water landward.

Several important questions remain. While we have shown that the general regime of the salinity gradients flip from a vertically stratified system at neap to a horizontally stratified system at spring, our observations are inadequate for determining which mode is more important for maintaining the salt balance. To a large part, the strong horizontal gradients observed at spring are a product of circulation in the strongly curving channels of the Satilla River. Due to the limited dynamic range of river discharge that we observed, we are unable to assess how changes in river discharge affect the “flip-flop” in the fortnightly change in the stratified regime. We would need more detailed surveys covering a larger range of river discharge.

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