Transverse structure of subtidal flow in a weakly stratified subtropical tidal inlet

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A B S T R A C T

The transverse structure of exchange flows and lateral flows as well as their relationship to the subtidal variability are investigated in a subtropical inlet, Ponce de Leon Inlet, Florida. Two surveys were executed during different phases of the tidal month to determine the spatial structure of subtidal exchange flows. Data from fixed moorings were used to depict the temporal variability of the spatial structure established in the surveys. The data suggested a tidally rectified pattern of net outflow in the channel and inflow over shoals with a negligible influence of streamwise baroclinic pressure gradients on the dynamics and slight modifications due to the wind. Onshore winds strengthened net inflows but weakened net outflows, rarely reversing them, while offshore winds increased net outflows and weakened net inflows. Curvature effects were found to be important in modifying secondary circulations. Slight modifications to the secondary flows were also caused by stream-normal baroclinicity during one survey. Most important, the intensity of the exchange flows was modulated by tides, with the largest exchange flows developing in response to the strongest tidal rectification of spring tides.

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1. Introduction

Subtropical inlets are unique in that they are found in regions where there is little to no net precipitation; the dynamics are therefore free from the density-driven flow common to temperate estuaries. Subtropical inlets have only sporadic freshwater influence and the residual flow is typically driven by tides and winds (Valle-Levinson et al., 2009), and modified by bathymetry (Kjerfve, 1978; Blanton et al., 2003; Li et al., 2008; Kim and Voulgaris, 2008). Subtropical estuaries are ideal sites for comparing observations on the spatial structure of tidal residual flow to theoretical results as there is little to no influence from along-channel density gradients.

Frequently, natural basins are strongly affected by channel curvature in both bathymetry and morphology. Centrifugal accelerations, produced by curvature, force a transverse flow away from the channel bends at the surface and toward the channel bends at depth (Kalkwijk and Booij, 1986). By definition, the depth-averaged transverse flow is zero everywhere within the inlet, and the transverse equation of motion in stratified flow reduces to (Kalkwijk and Booij, 1986; Geyer, 1993; Seim and Gregg, 1997; Lacy and Monismith, 2001)

\[
\frac{\partial u_r}{\partial t} + u_s \frac{\partial u_r}{\partial s} - \frac{u_r^2 - u_s^2}{R} = -\frac{g}{\rho} \int_z \frac{\partial \rho}{\partial n} dz + \frac{g}{\rho_b} \frac{\partial \eta}{\partial n} h + \frac{\partial}{\partial z} \left( A_z \frac{\partial u_r}{\partial z} \right) + \tau_{sh} + \tau_{sh} \frac{p_h}{h}
\]  

(1)

where vertical advection, depth-averaged lateral advection and Coriolis accelerations are neglected. The subscripts \( n \) and \( s \) represent transverse and streamwise velocity components, respectively, \( u \) is the velocity, overbars represent depth averages, \( R \) is the radius of curvature, \( g \) is the acceleration due to gravity, \( \rho \) is the water density, \( \eta \) is the sea surface elevation, \( z \) is the vertical direction, \( A_z \) is the eddy viscosity, \( \tau_{sh} \) is the bottom shear stress in the transverse direction and \( h \) is the depth of the water column. The radius of curvature is positive when the positive streamwise velocity (ebb, for our study) follows a clockwise trajectory due to the curvature; thus \( R \) changes sign during flood flow (Lacy and Monismith, 2001; Nidzieko et al., 2009). The third term on the left-hand side of Eq. (1) is the centrifugal acceleration term, which is generated by the streamwise velocity, \( u_r \). The first and second terms on the right-hand side of the equation are the lateral baroclinic forcing terms. Depending on their sign, these terms may act to oppose or enhance the curvature-induced lateral flows (Geyer, 1993; Dronkers, 1996; Seim and Gregg, 1997).
Observations and modeling of flow around tidal headlands (Geyer, 1993) and within curved estuaries have illustrated the existence of bathymetry-induced residual eddies (Li, 2006), enhanced vertical mixing due to channel curvature (Seim and Gregg, 1997), the importance of lateral density gradients in modifying the curvature-induced lateral velocities in weakly to strongly stratified estuaries (Dronkers, 1996; Chant and Wilson, 1997; Lacy and Monismith, 2001; Huijts et al., 2009) and channels that experience both well-mixed and stratified conditions (Nidzieko et al., 2009). Channel curvature has been found to affect transport of sediment and salt (Blanton et al., 2003; Kim and Voulgaris, 2008) while tides and winds modify the magnitude of the transverse flows (Chant, 2002).

Analytical models of residual flows within tidally driven basins (Li and O'Donnell, 2005; Winant, 2008), when adopted in a curved channel, show that lateral advection dominates in both long and short curved channels with residual eddies forming on either side of the channel bends (Li et al., 2008). In Li et al.’s (2008) model, where a cartesian coordinate system is adopted, curvature-induced flows and residual eddies are generated by the nonlinear advective terms of the momentum equations. Although advection dominates in curved channels, frictional effects of the tidal wave entering the channel may still enhance or diminish the advective terms and cannot be completely neglected, especially in long channels (Li et al., 2008). The dynamics of tidally driven residual flow can be described on the basis of a non-dimensional length. Basins whose non-dimensional lengths are < 0.6 – 0.7 are defined as short, while those above this threshold are defined as long (Li and O'Donnell, 1997, 2005; Winant, 2008). Basin length is the parameter that determines the features of residual circulation within a basin (Li and O'Donnell, 2005).

Theoretical results have been used to explain the observed lateral structure of tidal residual flows in long basins, both straight (Kjerfve and Proehl, 1979; Li et al., 1998; Cáceres et al., 2003) and curved (Li et al., 2008; Hench and Luettich, 2003), as well as in short straight basins (Winant and Gutiérrez de Velasco, 2003). Less is known, however, about the subtidal variability or modulation of the lateral structure of tidal residual flows in short or long basins under the influence of curvature. Curvature-induced secondary flow in a strongly stratified estuary increases with increasing tidal range while the secondary flow decreases with increasing river discharge (Chant, 2002). Li (2006) suggests that for bathymetrically induced residual eddies in a short, weakly stratified basin (where curvature was not included), both wind and tidal range have little effect on the structure or location of the residual eddies formed as a result of these bathymetric variations.

The purpose of this study is to document, with observations, the structure and variability of the along-channel tidal residual flows and whether the variability of the cross-channel flows help in the lateral redistribution of momentum in a long, subtropical, curved inlet, Ponce de Leon Inlet, in East Central Florida. This study also seeks to determine whether the modulation of the residual structure is produced by either tides or winds, or both. The subtropical nature of the inlet allows for direct observations of the tidal residual flows as along-channel density gradients are not influential to the dynamics.

Fig. 1. Study area of Ponce de Leon Inlet, Florida which connects the Intracoastal Waterway (IWW) to the Atlantic Ocean. The moored ADCP sites are denoted with diamonds (Stations 1, 2 and 3) and the towed ADCP survey line is denoted by a solid line across the inlet (Transect A). The CTD locations are denoted with filled circles (Stations A and B). The thalweg of the inlet follows from CTD Station A, ADCP Station 2 and CTD Station B. The two regions of curvature are North of CTD Station B (RB = 500 m) and South of CTD Station A (RA = 420 m). The meteorological station (New Smyrna Beach, Station 722 361) is denoted by the black square, 4 km to the South-West of Ponce de Leon Inlet.
2. Study area

Located on the Central East coast of Florida at 29 4.5°N, Ponce de Leon Inlet is a shallow subtropical inlet (Fig. 1). Flow within the inlet is forced by the semi-diurnal (M2) tide with a mean tidal range of 0.6 m and typical tidal currents exceeding 1 m s⁻¹. The inlet has been described as having a strong ebb channel adjacent to the North jetty (Militello and Hughes, 2000), which has been found to be responsible for considerable scour along the northern side of the inlet (Militello and Zarillo, 2000). Fresh water inputs from precipitation vary seasonally, with the strongest influences during May through September (Schwartz and Bosart, 1979).

The entrance to the inlet is 350 m wide and 12 m deep in a channel that is < 100 m wide. The thalweg of the channel decreases to 3.5 m within 750 m inside of the inlet. This narrow channel is flanked, asymmetrically, by shoals that are typically 2–4 m deep. Ponce de Leon Inlet has two regions where the channel curvature is large (in both bathymetry and morphology) with the first curved spit occurring on the northern shore, inside the inlet (see Fig. 1), with a radius of curvature $R_A$ of 200 m. The second curved spit is seaward of the first, on the southern shore of the inlet (see Fig. 1, $R_B$ = 420 m). This spit is accreting northward on both ebb and flood tides due to the curvature effect on sand transport (Militello and Zarillo, 2000). For ebbing flows, the curvature at $R_A$ is clockwise and $R_B$ is counter-clockwise. A two-layer curvature-induced transverse flow and possible formation of residual eddies, downstream and upstream of the region of curvature, are expected (Li et al., 2008).

Ponce de Leon Inlet is one of several coastal inlets connecting Florida’s Intracoastal Waterway (IWW) to the Atlantic Ocean (Fig. 1). The IWW, a long and narrow coastal lagoon, extends along the entire eastern coast of Florida with a mean width of 55 m, a mean depth of 3.5 m and a maximum depth of 5 m (Smith, 1983; Kenworthy and Fonseca, 1996). Within the IWW significant tidal attenuation occurs, 50 times greater attenuation than at the inlet (Militello and Zarillo, 2000). Due to the extent of the IWW along the East coast of Florida, the length of the Ponce de Leon Inlet-IWW system, as defined by Li and O’Donnell (2005) and Winant (2008), is difficult to determine and hence, will be determined based on the observations of the tidal wave type at the entrance to the inlet (Li and O’Donnell, 2005).

Given the shallow mean depth (4 m), Ponce de Leon Inlet can be classified as a strongly frictional inlet with a frictional constant (Winant, 2008) of

$$\delta = \frac{2A_v}{\omega H^2} = 0.94$$  \hspace{1cm} (2)

where $A_v$ is a constant vertical eddy viscosity, estimated to be 0.001 m² s⁻¹ for this case, $\omega$ is the frequency of the semidiurnal (M2) tide and $H$ is the mean depth. This frictional constant can be calculated for an inlet to determine the frictional influence on flow dynamics whether strong ($\delta = 1$), moderate ($\delta = 0.5$) or weak ($\delta = 0.1$). Therefore, Ponce de Leon Inlet is a highly frictional inlet with tidally induced residual flows that may be described for flows at the upper end of the frictional limit as considered by Winant (2008).

3. Data collection

In order to characterize the lateral structure of tidal residual flows as well as the temporal variability of this structure, several types of observations were obtained. Measurements of underway current velocity profiles and water density profiles were combined with time series of current profiles at the entrance to Ponce de Leon Inlet.

3.1. Underway surveys

Underway surveys were carried out at the entrance to Ponce de Leon Inlet on September 5, 2007 and on February 21, 2008 (Transect A, Fig. 1). Both surveys stretched for nearly a semi-diurnal cycle (11.2 h) during a neap tide in September and during a spring tide in February (Fig. 2). During both surveys, a boat-mounted 1200 kHz Acoustic Doppler Current Profiler (ADCP) with bottom-tracking capability was used to measure current profiles along cross-channel transects. The transects were sampled 15 and 17 times during the September and February surveys, respectively. Full across-inlet coverage was ensured by sampling within 3 m of the inlet perimeter or until the depth limited navigation. Given the accretion on the southern spit, depths on the southern side of the transect were very shallow. The temporal coverage during each survey allowed for the separation of tidal flows from non-tidal signals through a least-squares fit to the semi-diurnal (M2) harmonic (Lwiza et al., 1991) after rotating the currents to the principal axis of maximum variance considering ebb and flood tides separately.

During the surveys, water density profiles were measured with a SeaBird SBE19-Plus conductivity–temperature–depth (CTD) profiler at the deepest part of the survey transects during every other transect repetition (CTD Stations A and B, Fig. 1). This allowed for the estimation of mean horizontal density gradients and their influence on the mean flows, which was found to be small. A surface conductivity and temperature (CT) sensor, SeaBird SBE37, was mounted on the survey boat to capture the lateral surface variability in the density over the tidal cycle, recording at 0.2 Hz. Lateral density gradients proved to be larger and more important than the longitudinal density gradients.

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{fig2.png}
\caption{Tidal amplitude (dashed line) and subtidal water level (solid line). Periods of the towed ADCP transects are (○) September 5, 2007 and (■) February 21, 2008. Water level data was obtained from NOAA Tide Station 8721147 (29 3.8°N, 80 54.9°W) and filtered using a low pass Lanczos-cosine filter with half-power of 60 h.}
\end{figure}
3.2. Moored survey

Time series of current velocity profiles were obtained with bottom-mounted ADCPs, equipped with pressure sensors, deployed at three locations across the width of the inlet. Two instruments (Stations 1 and 3 in Fig. 1) were moored over the shoals flanking the channel, and the third instrument was installed in the channel (Station 2 in Fig. 1). These instruments recorded data for 78 days distributed in two periods, from January 14, 2008 to February 25, 2008 and from February 25, 2008 to April 2, 2008. The two data sets were joined using a spline fit between the currents and water level. During the second period the instrument at Station 3 was damaged and the record was unavailable. Instruments recorded the average of 400 pings distributed over 10 min intervals at 0.5 m bins. The currents were rotated to the principal axis of maximum variance (Emery and Thomson, 2004) for both ebb and flood tides. Positive along-channel currents indicated flow out of the inlet. Angles of rotation for Stations 1, 2 and 3 were 17.5 and 20.0, 39.2 and 38.1, 41.9 and 39.1 T for ebb and flood tides, respectively. Positive cross-channel currents were directed 90° counter-clockwise from the principal axis.

For the purposes of investigating the secondary flows associated with curvature effects in both the moored and towed surveys, the lateral residual flow was calculated as the anomaly of the vertical profiles relative to the depth averaged value (e.g. Geyer, 1993). The deployment-long average was calculated for each mooring to characterize the lateral structure of the mean flow. The subtidal variations were determined using a low-pass Lanczos-cosine filter with half-power of 34 h to eliminate tidal and inertial variations.

3.3. Atmospheric data

Local wind observations were obtained from New Smyrna Beach meteorological station (Station 722361 from the National Climatic Data Centre) located at 29°3’N, 80°57’W, 4 km to the South-West of Ponce de Leon Inlet (Fig. 1). The data were collected at a height of 3 m and then corrected to 10 m using the log-law velocity profile. The wind velocities, filtered with a 34-h Lanczos filter, were then rotated to be aligned with the coast, the log-law velocity profile. The wind velocities, filtered with a spectral analysis between sea level and streamwise currents gave friction (Parker, 1991). The coherence squared from a cross-varying the flow characteristics across the width of the inlet. For the purpose of investigating the secondary flows associated with curvature effects in both the moored and towed surveys, the lateral residual flow was calculated as the anomaly of the vertical profiles relative to the depth averaged value (e.g. Geyer, 1993). The deployment-long average was calculated for each mooring to characterize the lateral structure of the mean flow. The subtidal variations were determined using a low-pass Lanczos-cosine filter with half-power of 34 h to eliminate tidal and inertial variations.

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4. Tidal and subtidal flow

4.1. Tidal information

Tidal constituents from the streamwise velocity \(u_t\) and surface elevation \(h\) were obtained from the moored ADCPs using T_TIDE (Pawlowicz et al., 2002; Table 1). The quarter-diurnal amplitude was greater than the diurnal amplitude from the velocity records at both shallow moorings (1 and 3), consistent with nonlinearities due to the depth effect on bottom friction (Parker, 1991). The coherence squared from a cross-spectral analysis between sea level and streamwise currents gave an average phase of −149° at the semi-diurnal tidal frequency (1.9 cpd) indicating that tidal currents lead the water level by 1.04 h (Fig. 3b). This phase lead, confirmed by the phase difference from the harmonic analysis \((ΔΦ, Table 1)\), indicated that the tide in Ponce de Leon Inlet was closer to a progressive than a standing wave. Due to the progressive-like nature of the wave at the mouth of the inlet, the lagoon-inlet system was likely behaving as a long channel as defined by Li and O’Donnell (2005).

From the large \(Δ\) calculated in Eq. (2) and the large quarter-diurnal constituent \((M4)\), it follows that frictional effects were likely important. Since this inlet-lagoon system has small along channel baroclinic pressure gradients, and the tide is the dominant forcing mechanism, the dynamics are likely represented by the models of Li and O’Donnell (2005), Li et al. (2008) and Winant (2008), for tidally driven flow through a highly frictional channel of variable depth. Whether this inlet showed similar mean flows as predicted by the theory was explored using both moored and survey data. Given the strongly curving nature of the inlet, a streamwise and stream-normal coordinate system was adopted (unlike the analytical models above) in order to fully isolate the curvature induced secondary flows and to determine whether these lateral flows redistribute streamwise momentum in the transverse direction.

4.2. Long term mean flows

The average streamwise flow over the entire 78 day record at each of the mooring stations indicated vertically averaged outflow in the channel (Station 2) and inflow over the shoals at Stations 1 and 3 (Fig. 4). Lateral velocity showed vertical shears with bottom flow toward the northeast (NE) and surface flow toward the east-northeast (ENE) at Station 2. The vertical structure of the lateral velocities was consistent with curvature-induced secondary flow of Kalkwijk and Booij (1986) around a negative bend on ebb. A similar helical structure was observed at Station 1, with less shear in the lateral flows.

The differences between mean ebb and flood flows illustrated the tidal evolution of the streamwise and lateral flows. At Station 2, typical curvature-induced lateral flows occurred during ebb, with surface flows pushed southward (away) from the inshore bend \(R_b\) and bottom flows directed back toward the bend. However, during flood, the flow was likely still under the influence of \(R_b\) as the lateral flows showed surface flows directed northward at the surface, and return flows toward the spit, \(R_b\), at the bottom. These lateral patterns were observed throughout nearly the entire moored observation period and will be discussed further in the next section. To investigate the consistency of both the streamwise and lateral structure of the flow, the survey across the mouth offered a more detailed spatial resolution of the mean...
Fig. 3. (a) Coherence squared and (b) phase of the streamwise velocity and water surface elevation from Station 1 (thick line, o’s) and from Station 2 (thin line, x’s) from the bottom bins of each ADCP. Confidence interval (95%) is 0.26 for both current meters and is represented by dashed line in (a). Phase at 1.9 cpd indicates a −147° phase lag between streamwise currents and sea surface elevation.

Fig. 4. Deployment-long mean flow vectors at the three ADCP mooring sites in Ponce de Leon Inlet, Florida. Stations 1 and 2 show transverse shear with depth where the thick solid arrows represent the surface flow and the thick dashed arrows represent the bottom flow indicating the importance of channel curvature. For Station 3 (with only one bin depth), the mean flow is represented by a single vector. Mean flows during maximum flood and ebb are shown in the two lower panels and illustrate the difference in lateral velocities between the two phases of the tide. A stronger lateral shear is observed during ebb showing the classical helical structure of curvature-induced flows.
flow structure (where the radius of curvature is positive for ebb flows).

4.3. Tidal and residual flows

Transects across the mouth of the inlet were measured during one semi-diurnal tidal cycle at different phases within the fortnightly tidal cycle and under distinct wind forcing conditions (Fig. 2). These conditions influenced the residual flows across the mouth of the inlet during the two sampling periods in September and February.

4.3.1. Tidal flows

The transects at the mouth of the inlet were separated into maximum ebb and flood (Figs. 5 and 6) to determine how the lateral structure of the flow differed, depending on the phase of the tide. The average at maximum ebb and flood phases showed a distinct lateral structure, similar between the two surveys. In both surveys, maximum flood currents occurred mid-way through the channel during flood while during ebb, streamwise velocities were strongest along the northern edge of the inlet, as observed by Militello and Zarillo (2000), down to a depth of 6 m. Although both surveys occurred during different stages of the tidal cycle, tidal currents averaged over the channel width (from 0 to 110 m) were similar between the two surveys (Figs. 7f and 8f).

Lateral flows differed between the phases of the tide. For both surveys, lateral currents during maximum ebb showed flow away from the bend (northward) at the surface and flow toward the bend (southward) close to the bottom in the channel (Figs. 5b and 6b). During maximum flood currents, both surveys showed two counter rotating cells of weak lateral flow across the inlet. Curvature effects likely had less effect on the flow during flood as the across channel transects were carried out at the entrance to the inlet where flooding currents had not yet encountered strong bathymetric curvature. The effect of channel curvature, indicated by the vertically sheared lateral flows visible during ebb currents, were investigated by calculating the centrifugal acceleration term from the streamwise velocities, averaged over the channel width (from 0 to 110 m).

Centrifugal acceleration $\left(\frac{u^2 - \bar{u}^2}{R}\right)$ over the tidal cycle was on the order of $10^{-4} \text{m s}^{-2}$ for both surveys with positive (surface) and negative (near bottom) components occurring at maximum ebb and flood (1300/1900 h, Fig. 7a and 1800/1300 h, Fig. 8a) as observed in other studies (Lacy and Monismith, 2001; Kim and Voulgaris, 2008; Nidzieko et al., 2009). Although the September survey occurred during a neap tide, centrifugal accelerations were stronger and more well organized during this survey than during the February spring tide survey. This was attributed to lateral baroclinicity.

Lateral baroclinic pressure gradient accelerations were calculated as in the second term on the right-hand side of

![Fig. 5. Streamwise and lateral currents (cm s$^{-1}$) during (a) maximum flood and (b) maximum ebb across Transect A from Ponce de Leon Inlet on September 5, 2007. Asymmetry over the tidal cycle is noted by the change in position of the velocity maxima between flood and ebb tides. North is on the left-hand side of the figure, and positive velocity is toward the north (in the lateral direction).](image)

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\[ \text{Centrifugal acceleration} \left(\frac{u^2 - \bar{u}^2}{R}\right) \text{ over the tidal cycle was on the order of } 10^{-4} \text{ m s}^{-2} \text{ for both surveys with positive (surface) and negative (near bottom) components occurring at maximum ebb and flood (1300/1900 h, Fig. 7a and 1800/1300 h, Fig. 8a) as observed in other studies (Lacy and Monismith, 2001; Kim and Voulgaris, 2008; Nidzieko et al., 2009). Although the September survey occurred during a neap tide, centrifugal accelerations were stronger and more well organized during this survey than during the February spring tide survey. This was attributed to lateral baroclinicity.}
\]```
Eq. (1), $g/\rho \Delta \rho / Bh$, where $B$ is the width of the inlet. Given the outflow along the northern side of the inlet during ebb tide as observed by Militello and Zarillo (2000) and in Figs. 5 and 6, the location of the density minimum was likely a result of differential streamwise advection (Dronkers, 1996; Lacy and Monismith, 2001). Differential advection caused the minimum in density to be located along the northern edge of the inlet. The actual density minimum varied in position between towed surveys from the northern side of the inlet (80 and 60 m from the northern edge of Transect A for September and February surveys, respectively). In both surveys, a marked minimum in density occurred at the end of ebb (Figs. 7b and 8b) and the surface baroclinic pressure gradient was stronger during September than in February (Figs. 7c and 8c).

Further inside the inlet, the baroclinic effect on the lateral flows was observed from the mooring data at Station 2 (Fig. 9) which was under the effect of $R_b$ on ebb and $R_s$ on flood. The vertical shear in the lateral velocities, $\Delta \dot{u}_l$, was calculated as the difference in lateral velocity between the top and bottom bins of the ADCP. Low density water, at the moorings, exits the inlet following the northern edge of the shore. As ebb progresses, lateral baroclinic pressure gradients develop, acting in the same direction as the centrifugally forced flow (at the surface at $R_s$) causing an increase in the vertical shear in the lateral currents ($\Delta \dot{u}_l$) from mid-ebb until the beginning of flood.

Although streamwise friction was likely a strong influence on the streamwise dynamics within the inlet, lateral bottom friction was an order of magnitude smaller than streamwise bottom friction as calculated using $C_D \dot{u}_l |V| / H$, where $|V|$ is the magnitude of the velocity and $C_D$ is the drag coefficient ($C_D = 0.0025$; Figs. 7f and 8f). Modifications to the flow over the tidal cycle by the processes mentioned above were linked to the observed residual flows in both the streamwise and lateral velocities as explained next.

### 4.3.2. Streamwise residual flows

Streamwise residual velocities had a similar structure during both surveys (Figs. 10 and 11). Streamwise and lateral residual velocities were calculated by removing the tidal signal from the observed data through a least-squares fit to the semi-diurnal (M2) tide (Lwiza et al., 1991). During September, weak inflow of $5\text{ cm s}^{-1}$ was present in the first 50 m of the transect with maximum inflow velocities confined to 3 m depth. Southward of 50 m along the transect, the streamwise residual flow was directed into the inlet where maximum inflow currents occurred over the shoals up to $20\text{ cm s}^{-1}$. During February, the streamwise residual flow showed outflow on the northern side of the inlet (0–75 m across the inlet) with maximum currents of $20\text{ cm s}^{-1}$ constrained between 2 and 6 m depth (Fig. 11). The shift toward inflow, although relatively weak at $-5\text{ cm s}^{-1}$, occurred on the slope between 75 and 150 m. Weak, < $6\text{ cm s}^{-1}$, outflow was observed south of the slope.

The differences in the streamwise flow between the two surveys were due to external forcing mechanisms such as changes in stratification, the fortnightly tidal cycle and wind effects. Hydrographic data obtained from both surveys showed that the water column was mixed during flood and weakly stratified during ebb. Profiles from the end of ebb and flood are shown in Figs. 7d and 8d. At the end of ebb, the vertical density gradient $\langle \partial ho / \partial z \rangle$ at the mouth of the inlet was $1.0$ and $0.2\text{ kg m}^{-3}$ (for September and February surveys, respectively) while further inside the inlet at Station 3, the vertical density gradient $\langle \partial \rho / \partial z \rangle$ at the end of ebb tide decreased to $0.7$ and $0.01\text{ kg m}^{-3}$ (for September and February transects, respectively). The tidally averaged accelerations produced by the baroclinic pressure gradient, $\langle g \rho \partial \rho / \partial z \rangle$ (where angle brackets denote the tidal average), were small ($0(10^{-4})\text{ m s}^{-2}$). Over the tidal cycle, baroclinic pressure gradients never exceeded $0(10^{-5})\text{ m s}^{-2}$ for
both surveys. Therefore, the along-channel baroclinic pressure gradient was neglected as a mechanism affecting the tidal and residual flow dynamics within the inlet.

In density-driven estuaries, the fortnightly spring–neap cycle results in stronger residual flows during neap tides due to reduced vertical mixing (Haas, 1977). Conversely in tidally driven subtropical estuaries, the fortnightly spring–neap cycle results in stronger residual flows during spring tides rather than during neap tides (Valle-Levinson et al., 2009). Li et al. (2008) suggested that the location and strength of residual eddies in curved tidally driven inlets vary little with tidal amplitude. The September survey occurred during a neap tide with an increasing mean water level while the February survey occurred during a spring tide with a decreasing water level over the sampling period (Fig. 2). Although streamwise velocities averaged across the channel were similar (Figs. 7f and 8f), the residual flow in September showed unequal exchange with significantly weaker outflow than inflow which corresponded to a net volume flow into the inlet in agreement with increasing water level. Similarly, given the decreasing water level during the February survey, the residual exchange flow showed strong outflow and weak inflow resulting in net volume flow out of the inlet.

### 4.3.3. Lateral residual flows

In both surveys, transverse residual flows showed depth variable lateral flow with a two-layer structure (average of every other vertical profile plotted in Figs. 10 and 11). In both cases, lateral flow at the surface was directed toward the North (away from the bend $R_A$) while lateral flow below 4 m and below the surface layer along the southern shoal was directed southward, toward the bend $R_A$.

Tidally averaged lateral baroclinic pressure gradients were stronger during the September survey while lateral residual velocities were stronger during February. This was likely a result of there being no lateral density gradient to counteract the centrifugally generated flow. Weak lateral residual flows during September were due to strong baroclinic pressure gradients suppressing centrifugal accelerations (Seim and Gregg, 1997; Chant, 2002; Lacy and Monismith, 2001; Nidzieko et al., 2009).

Therefore, $\partial p/\partial y$ was likely playing a role in modifying the residual flow within the inlet, even though the along-channel baroclinic pressure gradient was small. Fresh water input to Ponce de Leon Inlet is sporadic and flow forced only by centrifugal accelerations will also likely be sporadic, depending on the external freshwater sources. Given the strong lateral shear, the presence of strong lateral velocities due to a weak baroclinic

![Fig. 7. Centrifugal acceleration ($u^2/C_0$) (a) variability over the tidal cycle and (b) average over the tidal cycle; (c) surface $\sigma_I$ (20 m, 60 m and 360 m from northern edge of transect A, respectively); (d) profiles of density at end of ebb (dotted) and flood (solid) at A (thick) and B (thin); (e) lateral baroclinic pressure gradient calculated at the surface across transect A from 60 m to 360 m and streamwise (dashed-dot) and stream-normal (dotted) bottom friction parameterizations; and (f) streamwise tidal velocity, averaged over the channel width (0 to 110 m), across transect A from Ponce de Leon Inlet on September 5, 2007.](image-url)
pressure gradient may affect the redistribution of momentum within the inlet which will be explored using the mooring data. Also to examine the details of the modulation of residual flows suggested by the surveys, the time series measurements were examined next.

5. Subtidal modulation

With the knowledge that curvature is an important driving mechanism in this inlet, the cartesian coordinate system of Li and O’Donnell (2005) and Winant (2008) is used to study how lateral flows helped in the redistribution of streamwise momentum. This was examined with the variation of the advective terms in the streamwise direction, \( \frac{\partial \mathbf{u}}{\partial \mathbf{y}} \), where \( \mathbf{u} \) and \( \mathbf{v} \) are along and cross-channel velocities, respectively. Subtidal variations of the streamwise exchange flows inside the inlet confirmed survey observations that the pattern of outflow in the channel and inflow over shoals persisted throughout almost the entire observation period at all depths for Stations 1 and 2 (Fig. 12c). The subtidal flow at Station 3 (not shown) oscillated between inflow and outflow although the mean flow over the entire record was negative (inflow) and small. Due to the very close proximity of this mooring site to Station 2, the residual currents were likely under the influence of the flow in the channel.

During positive cross-shelf winds (offshore winds on February 14 and March 8, 2008 in Fig. 12b), the subtidal flow responded with an increase in magnitude of the outflow in the channel (Station 2) and a corresponding decrease of the inflow over the shoals (Station 1). Strong northerly along-shore wind pulses modulated the residual flow in the opposite direction with decreased outflow and increased inflow corresponding to an onshore Ekman transport (January 21, 2008 in Fig. 12b and c).

Modulation of magnitude of the residual flow throughout the observation period indicated a response of the residual circulation due to wind effects. Variability in the subtidal water level within the inlet was correlated with along-channel wind stress (correlation coefficient, \( R_c \), of 0.41), while cross-channel wind stresses were less correlated (\( R_c = 0.12 \)). Wind effects caused water level set-up during onshore winds due to a net transport of water into the entire cross-section of the inlet. Correspondingly, a water level set-down during offshore winds was related to a net seaward transport of water throughout the cross-section of the inlet. Remote wind effects, not related to observed winds, would cause similar transport into the inlet (out of the inlet) during rising (falling) water level.

Fig. 8. Centrifugal acceleration (\( u^2 / R \)) (a) variability over the tidal cycle and (b) average over the tidal cycle; (c) surface \( \sigma_t \) (20, 60 and 360 m from northern edge of transect A, respectively); (d) profiles of density at end of ebb (dotted) and flood (solid) at A (thick) and B (thin); (e) lateral baroclinic pressure gradient calculated at the surface across transect A from 60 to 360 m and streamwise (dashed-dot) and stream-normal (dotted) bottom friction parameterizations; and (f) streamwise tidal velocity, averaged over the channel width (0–110 m), across transect A from Ponce de Leon Inlet on February 21, 2008.
Fig. 9. (a) Lateral velocities, $u_n$ (m/s), from the surface (solid) and bottom (dotted; positive northward) from Station 2; (b) Vertical shear in the lateral velocities ($\Delta u_n = u_{\text{surface}} - u_{\text{bottom}}$) from Station 2 (m/s). Negative lateral shear means centrifugally forced flow for ebb flows around the inshore bend ($R_B$), and positive lateral shear means centrifugal type flow from $R_A$ with weak surface lateral flows. Shading along the zero line indicates flood (gray) and ebb (black).

Fig. 10. Contours of residual streamwise flow (cm s$^{-1}$) as calculated from the least squares fit to the semi-diurnal tidal constituent across Transect A from Ponce de Leon Inlet on September 5, 2007. Lateral velocities (arrows) are the anomaly of the vertical profiles relative to the depth averaged profile and the average of every other profile is shown. North is on the left-hand side of the figure, and positive velocity is out of the inlet (in the streamwise direction) and toward the North (in the lateral direction).

Fig. 11. Contours of residual streamwise flow (cm s$^{-1}$) as calculated from the least squares fit to the semi-diurnal tidal constituent across transect A from Ponce de Leon Inlet on February 21, 2008. Lateral velocities (arrows) are the anomaly of the vertical profiles relative to the depth averaged profile and the average of every other profile is shown. North is on the left-hand side of the figure, and positive velocity is out of the inlet (in the streamwise direction) and toward the North (in the lateral direction).
The strength of the net exchange flow (Fig. 12 d), as determined by the difference between the subtidal outflow in the channel and inflow over the shallow station, was modulated by the fortnightly tidal cycle. The strongest exchange flows occurred during the largest spring tides of January 23 and March 9. Other large exchange flows developed during the spring tides of February 9, 21 and March 22. In general, the strength of the net exchange flow did not show evidence of wind effects as the wind velocity was not strongly correlated to the magnitude of the exchange flow ($R_s < 0.17$ for both cross-shore and along-shore wind stress compared to $R_s = 0.63$ with tidal amplitude variations). The observed modulation of the exchange flow was consistent with observations from a single moored time-series in a subtropical inlet that showed strongest exchanges during spring tides (Valle-Levinson et al., 2009). This study provides more complete observations of the lateral structure and the source of the modulation of these residual exchange flows.

Observations from events where the flow within the inlet was completely unidirectional (reversed flow at Station 2) tended to occur more frequently during neap tides with sustained winds. During the three neap tides, January 31, February 29, March 30, flow reversal occurred in each case. During spring tides (January 23, February 9 and 21, March 9 and 22), wind stresses were not large enough to overcome the tidally induced residual flow. As a result, no reversal of the flow at Station 2 was observed during spring tides. Therefore, during neap tides when wind stress exceeded bottom stress, flow reversal in the channel was likely overwhelming the tidally induced residual circulation.

A basic, vertically averaged, dynamical balance in the streamwise direction within the inlet, using the Boussinesq and hydrostatic approximations, can be approximated as

$$0 = -g \frac{\partial \bar{u}}{\partial x} + \frac{1}{\rho} \left( \nabla \cdot \bar{u} \right) + \frac{\tau_x}{\rho H} - \frac{\tau_y}{\rho H}$$

where the first term is the barotropic pressure gradient related to sea level slope, the second is lateral advective acceleration, and the third and fourth are the surface and bottom stress terms using parameterizations for wind and bottom frictional stresses. Although the sea level slope was not measured during this experiment, it was likely an important dynamical component.

In the streamwise direction, tidal curvature-induced flows will modify the streamwise momentum equation through lateral advective accelerations. In fact, results from the straight part of a curved estuary by Lacy and Monismith (2001) showed that the lateral advective term was high in regions where strong lateral gradients in streamwise flows coincided with strong lateral velocities. Although the magnitude of the streamwise shear ($\partial \bar{u}/\partial y$) was dependent on the spring neap cycle, depth averaged lateral velocities were not but were small given the curvature-induced flow yielding a vertical average close to zero. Depth-averaged lateral advection ($\langle \partial \bar{u}/\partial y \rangle$) was stronger in the channel than on the shoal, as expected (Li and O'Donnell, 2005; Winant, 2008; Fig. 12 e). However, lateral advection was strongest at the beginning of the record, prior to the first spring tide. The peak in lateral advection occurred in conjunction with strong NE winds followed by a peak in cross-shore (westward) winds. The peak in lateral advection occurred as a result of the wind rather than from the spring tides. Other small pulses in the lateral advective term (in the positive direction) corresponded with strong positive cross-shore (or negative along shore) wind events. Therefore, although streamwise exchange flows increased during spring tides, fortnightly variability in lateral advective accelerations was masked by the wind stress (second and third terms in Eq. (3)). Advection accelerations and were not sufficient to redistribute streamwise momentum laterally. Otherwise, the lateral shears associated with spring tides would have decreased as a result of the lateral redistribution of streamwise momentum.

Comparing the lateral advection accelerations to streamwise frictional accelerations ($\left( \zeta_u \nabla \bar{V} / H \right)$; Fig. 12 f), friction was the most dominant mechanism on the shoal while both terms were important in the channel. This was in agreement with Li and O'Donnell (2005) and (Winant, 2008) in driving a residual flow in a tidally dominated channel. Despite the strongly curving nature
of the channel, residual flows were generated mostly through the interaction between the tidal wave and depth variations. Frictional accelerations were strongest during spring tides.

6. Conclusions

In the highly frictional Ponce de Leon Inlet, observed residual flows were consistent with theoretical residual flows in a tidally dominated, curved channel. Residual flow, showing moderate recirculation, was persistent at two bends in the topography: on the lee of the upstream bend of the inlet as well as on the downstream bend further inside the inlet. A laterally sheared pattern of net outflow in the channel and net inflow over the shoals in the streamwise velocity persisted under different tidal amplitudes and most wind forcing conditions. Long term current measurements, moored across the inlet, provided novel observational information on the variability and modulation of the lateral flow within these subtidal curved inlets.

Observations indicated that strength of the curvature-induced residual circulation was affected by both winds and tides. The strength of the streamwise exchange flow across the inlet was modulated by tidal forcing from the spring–neap cycle. Unlike temperate estuaries, the strongest exchange flows occurred during spring tides confirming recent observations from a single moored time series (Valle-Levinson et al., 2009).

Along-channel winds caused unidirectional flows throughout the inlet. The direction of these flows was dependent on whether coastal set-up or set-down was observed, thereby reducing or increasing the inflow or outflow component of the tidal residual circulation. Under specific criteria of wind speed during neap tides, unidirectional flow was observed throughout the inlet with a reversal in the channel. In the lateral direction, centrifugal accelerations and lateral density gradient were influential in the dynamics. Lateral baroclinic pressure gradients acted to reduce lateral flows at the mouth of the inlet. Advection of streamwise shear by the lateral velocities was found to be related to the wind (and its effect on water level) masking the variability due to the spring–neap cycle.

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References