Tidal variations of flow convergence, shear, and stratification at the Rio de la Plata estuary turbidity front

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[1] Intratidal variability of density and velocity fields is investigated at the turbidity front of the Rio de la Plata Estuary, South America. Current velocity and temperature-salinity profiles collected in August 1999 along a repeated transect crossing the front are analyzed. Horizontal and vertical gradients, stability of the front, convergence zones, and transverse flow associated to the frontal boundary are described. Strong horizontal convergence of the across-front velocity and build up of along-front velocity shear were observed at the front. In the proximity of the front, enhanced transverse (or along-front) flow created jet-like structures at the surface and near the bottom flowing in opposite directions. These structures persisted throughout the tidal cycle and were advected upstream (downstream) by the flood (ebb) current through a distance of ~10 km. During peak flood, the upper layer flow reversed from its predominant downstream direction and upstream flow occupied the entire water column; outside the peak flood, two-layer estuarine circulation dominated. Changes in density field were observed in response to tidal straining, tidal advection, and wind-induced mixing, but stratification remained throughout the tidal cycle. This work demonstrates the large spatial variability of the velocity field at the turbidity front; it provides evidence of enhanced transverse circulation along the frontal boundary; and reveals the importance of advective and frictional intratidal processes in the dynamics of the central part of the estuary.


1. Introduction

[2] The Rio de la Plata Estuary, located at 35°S on the eastern coast of South America (Figure 1a) is one of the largest estuaries in the world and a major contributor of freshwater to the global oceans [Leopold, 1994]. An estuarine turbidity maximum (ETM) and a sharp surface front defining its seaward edge are distinctive features of this large-scale system. The turbidity front marks the transition between the freshwater outflow, rich in suspended sediment, and the brackish water, constituting the dynamical boundary between the upper and lower estuary [Framiñan et al., 1999].

[3] The estuarine turbidity maximum (ETM) is a complex feature of many river-dominated estuaries. It is an area where suspended sediment concentration and turbidity are markedly higher than landward or seaward [Officer, 1976; Dyer, 1986; Geyer, 1993; Burchard and Baumert, 1998; Uncle, 2002]. The ETM is considered the heart of the estuarine ecosystem; trapping of particles by convergent suspended sediment transport promotes biogeochemical, microbial and ecological processes [Simenstad et al., 1994; Le Bris and Glémarec, 1996]. Also, it is a fundamental factor in the distribution and fate of anthropogenic inputs to the estuarine systems. In the Rio de la Plata, numerous studies have analyzed the role of the turbidity frontal zone in the estuarine ecosystem [Acha et al., 1999; Lasta et al., 1996; Mianzán et al., 2001; Berasategui et al., 2004; Jauregúzar et al., 2003], as well as its significance for pollutant retention [CARP, 1989; Acha et al., 2003].

[4] Framiñan and Brown [1996] studied the characteristics of the turbidity frontal zone in the Rio de la Plata Estuary based on satellite imagery. Using remotely sensed Advanced Very High Resolution Radiometer (AVHRR) reflectance images they addressed questions about the location and spatiotemporal distribution of the frontal zone and ETM, and demarcated the boundaries of this large estuarine system.

[5] Satellite images provide appropriate horizontal and temporal coverage of the estuary, however, they only give information from the sea surface and are of limited use to study the vertical structure and its variability. It is clear that to fully understand the turbidity frontal dynamics we need
Figure 1. (a) Area of study and location of the ADCP/CTD transect ($C_W$, red line). Bottom topography (thin gray) interval is 1 m and a star represents the tide-gauge location. (b) NOAA-AVHRR visible image showing the ETM during the survey: blue-green-yellow (high reflectance) corresponds to turbid water with high concentration of suspended sediment and orange (low reflectance) indicate brackish water with low sediment load; clouds are masked in black. (c) View of the Río de la Plata estuary and the turbidity front from the R/V “Comodoro Rivadavia,” 16 August 1999. (d) Location of the ADCP and CTD survey along the central transect ($C_W$), winter cruise. ADCP track is shown in blue and CTD stations with red squares. Isobaths interval is 1 m (dashed lines).
to identify the processes that control the vertical structure and the diffusion of properties throughout the water column. On the basis of the mean distribution of the fronts derived from the satellite imagery and bulk stability parameters estimated using historical hydrographic data available in the area, Framiñan and Brown [1996] found a good correlation between the frontal mode locus and the head of the salt wedge in the Río de la Plata. However, a comprehensive analysis of the (in situ) characteristics of the turbidity zone and the association with the salt-wedge dynamics was not possible at the time due to inadequate resolution of the density data and the lack of in situ information of the velocity field.

[6] To further investigate the characteristics of the turbidity frontal zone, an observational study, the RIOPLA project ("RIO" for Río de la Plata and "PLA" for "plataforma," the Spanish word for shelf) was developed to obtain high-resolution data of the velocity and density field in the estuary and adjacent shelf [Framiñan et al., 2000; Framiñan, 2005]. At the ETM, data were collected at the northern, central and southern areas using towed acoustic Doppler current profiler (ADCP) and a conductivity-temperature-depth (CTD) recorder along repeated transects during a 25-h tidal cycle. Remote sensing data were used to guide shipboard sampling. A description of the data and the analysis of the tidal and subtidal flow observed during the experiment were presented by Sepúlveda et al. [2004] and Framiñan [2005].

[7] The objective of this study is to investigate the vertical structure and circulation associated to the turbidity front with focus on the intratidal variability. We analyze high-resolution data of the velocity and density field collected along one transect located across the front in the central part of the estuary. This unique data set allows characterization, for the first time, of the circulation associated to the frontal boundary. Along and across-front flow, horizontal and vertical gradients, stability, tidal straining conditions and the development of convergence zones and frontal jets are investigated.

2. Study Area

[8] The Río de la Plata estuary is located on the east coast of South America (Figure 1a), between 34°S, 36° 20'S latitude and 55°W, 58°30'W longitude. It is 320 km long and its width varies from 38 km in the upper region to 230 km at the mouth, and covers and area of ~3.5 × 10^6 km². This shallow (<20 m), large-scale estuary drains the second largest basin in South America with an area of 3.1 × 10^6 km² and it is one of the major contributors of freshwater to the global oceans [Leopold, 1994]. The two major tributaries are the Paraná River and the Uruguay River, with a combined mean annual river discharge of ~25,000 m³ s⁻¹ and maximum and minimum monthly discharge in May and January, respectively. The tidal regime is semidiurnal with influence of the O₁ diurnal component that results in diurnal inequality. Atmospheric forcing presents a marked seasonal variability with predominance of northeasters and easterlies during the summer months, northwest-westerlies and southwesterlies during late fall and winter and southeast-southerlies during winter. A description of the physical characteristics and forcing in the Río de la Plata region can be found in the work of Framiñan et al. [1999], Lopez-Laborde and Nagy [1999], and Framiñan [2005].

[9] A characteristic feature of this estuary is the turbidity front (Figures 1b and 1c) that delimits the “transition zone” between the upper and lower estuary. Large seasonal variability characterizes the frontal zone, especially on the northern coast, and the mean location is related to the estuarine geometry with the most frequent position coincident with the estuary’s constriction [Framiñan and Brown, 1996]. This area is located approximately 200 km from the estuary head where the estuary is ~100 km wide, and presents large differences in suspended sediment load, marked changes in water color, and surface salinity varying from freshwater to ~15.

[10] Seaward of the transition zone, the buoyant plume spreads offshore over a large geographical area in the outer estuary and the shelf. The plume has marked seasonal variability [Guerrero et al., 1997] controlled by river discharge, predominant winds and ambient flow [Framiñan, 2005]. The influence of the Río de la Plata freshwater outflow extends offshore beyond the 50 m isobath and northward along the South American coast as far as 28°S [Piola et al., 2005].

[11] When studying the Río de la Plata estuary, one must differentiate between the estuarine turbidity front (the “inshore” front which is the focus of this study) that separates the freshwater from the estuarine brackish water, and the plume front (the “offshore” front) that divides the brackish from the shelf water and is located more than one hundred kilometers seaward on the shelf. Our research focuses on the estuarine dynamics at the transition zone, rather than on the buoyant plume that spreads over the outer estuary and continental shelf.

3. Data and Methodology

[12] The ADCP and CTD data used in this study were collected at the turbidity maximum zone in the central part of the estuary during August 1999 in the Austral winter (Figure 1, transect Cw). The transect was oriented perpendicular to the front, on the basis of the NOAA-AVHRR satellite image of 15 August 1999, afternoon pass (1905 GMT), acquired a few hours before starting field operations in the area. The reflectance image presented in Figure 1b is a product derived from the visible band of the AVHRR and provides a proxy measurement of water turbidity associated with suspended sediments. In situ sediment data were not available to develop a calibration algorithm to estimate sediment load from the reflectance values, however the image provides valuable information to qualitatively describe the sediment distribution: high sediment concentration are shown in green-yellow tones and low suspended sediment load in orange-red. Suspended sediment values larger than 150 mg/l are common in the turbidity maxima zone and concentration drops to <50 mg/l downstream the front [Framiñan and Brown, 1996]. NOAA-AVHRR images were acquired and processed at the HRPT receiving station “Villa Ortizar” operated by the Argentine Meteorological Service, in Buenos Aires-Argentina, and the coordinates of the turbidity front were transmitted in real time to guide shipboard sampling.
The survey crossed the turbidity front (Figure 1c), in the area of Barra del Indio where the bathymetry changed from 8 to 10 m depth along the 15 km transect (Figure 1d). The transect was sampled from 16 August, 0013 GMT to 17 August, 0148 GMT. We completed 12 repetitions, but data from the last repetition were not included in the analysis because of the high noise level due to strong winds and rough seas.

Concurrently to the ADCP survey, 38 CTD casts were taken from a second vessel following a track parallel to the ADCP survey (Figure 1d). Stations at the center and endpoints of the transect were occupied on all repetitions. Higher resolution sampling was done during the seventh, eighth and ninth repetitions when a sharply delineated front on the surface was observed in the area. The front was clearly defined with the strong change in water color and a foam line (Figure 1c). The ship crossed the front at 35° 21.37'S, 56° 24.83'W, on 16 August 1999 at 1416 GMT.

Tidal forcing information consisted of sea level hourly records from nearby station Torre Oyarvide, provided by the Argentine Naval Hydrographic Service. A lag of 4 s between the time series of water level at Torre Oyarvide, located upstream approximately 40 km to the northwest of the transect (Figure 1a), and the tidal flow observed at the transect is expected. Still, the sea level record provided a good reference of tidal conditions in the area. Winds were measured onboard, and monthly mean river discharge from the Paraná and Uruguay rivers was used to characterize the freshwater outflow at the time of the survey. Forcing information is displayed in Figure 2.

Data from the underway ADCP were recorded every 5 s while cruising at a speed of ~2.5 m s⁻¹. The data were then averaged in bins of 16 profiles, i.e., every 80 s or 200 m in spatial resolution. The spread of the ADCP beams causes the instrument to instantaneously record different water types; averaging several pings per ensemble reduces this effect. Despite the averaging process, the frontal features were clear in the results portrayed here. After calibration [Joyce, 1989] and quality control, the velocity data were interpolated onto a regular grid of 0.25 m and 200 m of vertical and horizontal resolution, respectively. The nearest surface observation was centered at 1 m, and data in the lower 15% of the water column were discarded due to interference from sidelobe effects [RD Instruments, 1996].

A least squares fit to harmonic signals with periods of 12.42 h and 23.92 h was used following the procedure described by Lwiza et al. [1991] to estimate the subtidal or residual flow and the amplitude and phase of the semidiurnal and diurnal components [Sepúlveda et al., 2004].

The velocity data are presented as along and across estuarine components (U, positive seaward; V, positive to Uruguayan coast), which are oriented in the across-front and along-front direction, respectively. Each transect repetition took nearly 2 h to complete. Despite the apparent aliasing with respect to tidal phases, the observed flow and the calculated gradients were coherent with those derived from the fit to semidiurnal and diurnal harmonics. The 200 m horizontal average gives a temporal resolution for the averaged profiles of approximately 100 s. Thus, higher frequencies are filtered and the data can be considered turbulent-mean components of the velocity. The data provides a quasi-instantaneous representation of the velocity field at the front.

### 4. Results

#### 4.1. Mean Density and Velocity Field

The salinity, temperature, and density field across the turbidity front were described with the CTD profiles acquired along the transect (Table 1). Salinity ranged from freshwater to 21 while the temperature field was relatively homogeneous; density is controlled by salinity in this estuary [Guerreiro et al., 1997]. The mean density field (Figure 3) clearly showed a two-layer structure with the salt-wedge intruding underneath a fresher upper layer. The buoyant layer had density <1008 kg m⁻³ (salinity <10), and the lower layer showed maximum salinities of 21 at the seaward end. This value corresponds to dilute shelf water (characterized by salinities of 33). The upper layer thickness

<table>
<thead>
<tr>
<th>Temperature (°C)</th>
<th>Salinity</th>
<th>Density (kg m⁻³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minimum</td>
<td>7.72</td>
<td>0.00</td>
</tr>
<tr>
<td>Maximum</td>
<td>10.72</td>
<td>20.99</td>
</tr>
<tr>
<td>Mean</td>
<td>9.98</td>
<td>12.57</td>
</tr>
<tr>
<td>SD</td>
<td>0.27</td>
<td>4.12</td>
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Table 1. Statistics of Salinity, Temperature, and Density at the Turbidity Front (August 1999) Based on a Total Number of 35 CTD Stations
was \( \sim 4 \) m and the head of the buoyant flow was located at 14 km from the transect origin, where the surface front was observed (Figure 1).

[19] The along-estuary residual flow (Figure 4, top left) showed the buoyant layer moving seaward (positive U) with surface residual velocities of 0.20–0.30 m s\(^{-1}\) and the lower layer advancing upstream at \( \sim 0.05 \) m s\(^{-1}\). The 0-m s\(^{-1}\) isotach occurred at about 5 m depth. Near the surface, the across-front residual velocity decelerated close to the seaward end of the transect (\( x = -14 \) km) where the turbidity front was observed with an across-front residual (indicated by brackets) shear \( < \frac{\partial U}{\partial x} >, < \frac{\partial V}{\partial z} > \) of \(-10^{-4} \) m s\(^{-1}\) and \(-0.6 \times 10^{-4} \) m s\(^{-1}\), respectively, indicating convergence at the front. The across-front residual flow was not zero at the boundary but dropped to \(< 0.10 \) m s\(^{-1}\). Large vertical shear was associated to the buoyant layer (\( < \frac{\partial U}{\partial z} > = 0.1 \) s\(^{-1}\)). The along-front residual flow (\( < V > \)) is shown in Figure 4 (top right). There is evidence of a strong subsurface jet in the upstream side of the front with velocity reaching \( \sim 16 \) cm s\(^{-1}\) toward the Uruguayan coast and enhanced flow (\( \sim 4 \) cm s\(^{-1}\)) in the opposite direction near the bottom. A vertical shear of \( < \frac{\partial V}{\partial z} > = 0.03 \) s\(^{-1}\) was observed at the jet location.

[20] Semidiurnal and diurnal amplitudes of the tidal current are presented in Figure 4 (center) and Figure 4 (bottom), respectively. Tidal currents were stronger at the surface than near the bottom. Mean amplitudes for the semidiurnal and diurnal components were 0.29 ± 0.05 m s\(^{-1}\) and 0.25 ± 0.02 m s\(^{-1}\) at the surface and 0.22 ± 0.06 and 0.12 ± 0.02 m s\(^{-1}\), near the bottom; deeper semidiurnal and diurnal currents led the surface. A marked influence of stratification and the frontal structure was observed; the tidal ellipses changed the direction of rotation following the pycnocline geometry and were larger and more elliptical in the proximity of the front and at the frontal jet [Framiñan, 2005]. This spatial variability in the semidiurnal and diurnal components seems to be associated to frontal features instead of astronomical forcing.

4.2. Variations During a Tidal Cycle

4.2.1. Velocity Field

[21] The velocity field measured along the transect is presented in Figure 5. On the left-hand side, the component U represents the along estuary (across-front) flow positive seaward; on the right-hand side, the component V describes the across estuary (along-front) flow. Acquisition of ADCP data started on 16 August 1999 at 0013 GMT at the shallower point (\( x \) axis origin in Figure 5), during strong ebb flow, 4 h after the high tide at Torre Oyarvide. The ship headed seaward in the odd number repetitions; during even number sections travel was upstream. Operations started after sunset and therefore, direct observations of the surface expression of the front were not made until dawn, which occurred during the fifth repetition (Cw-5).

[22] The first repetition was surveyed during strong ebb and showed outflow from top to bottom. The second repetition marked the transition from ebb to flood: transect began at the deepest end with downstream flow at the surface and upstream in the lower layer and ended with upstream flow in the entire column as a result of the strong flood. Maximum high water during the survey (1.15 m) was recorded at Torre Oyarvide at 0400 GMT and flow in transects Cw-3, Cw-4, and half of Cw-5 corresponded to peak flood. Flow was upstream in the whole water column, indicating reversal of the freshwater surface outflow. During this flood tide, the along-estuary velocity (U) showed a maximum at \( \sim 4 \) m depth.

[23] In the first transect, a maximum in the along-front (V) velocity of \(+0.45 \) m s\(^{-1}\) at 2 m occurred at 13 km, with strong gradient and sign change \( \sim 2 \) km downstream of it. This positive maximum was noticeable in the following four repetitions, even when the flow changed to strong flood. It had a progressive upstream position and was located at 8 km in repetition Cw-5. In the cross-front component (U), minimum upstream velocity during flood was associated to the same location. This velocity structure was consistent with a surface jet along the front. The upstream movement of this jet indicated advection of the frontal structure at a speed of \( \sim 0.20 \) m s\(^{-1}\), with a total excursion of approximately 8 km. Inspection of the density field (Figure 6) shows that this frontal structure was associated to the isopycnal of 1010 kg m\(^{-3}\), which corresponded to the 12 isohaline. The velocity (and density) structure during the first five (Cw-1 to Cw-5) repetitions clearly suggested the presence of the front. Unfortunately, direct observation of the sea surface was not possible at nighttime, so there was no visual confirmation of the surface front.
Repetitions Cw-6 and Cw-7 were surveyed during ebb flow. There was outflow throughout the water column over the shallowest part, and two-layer flow in the deepest part of the transect. Along-estuary outflow at the surface was $U > 0.60 \text{ m s}^{-1}$ and inflow was $U \approx 0.15 \text{ m s}^{-1}$ near the bottom. During Cw-7, approximately 14 km from the transect’s origin a marked surface front was observed (see photograph in Figure 1c). This front was also identified at Cw-8 near the seaward end of the transect. The flow associated to the front showed increased gradients at the surface, with across-front velocity ($U$) dropping from 0.65 to 0.20 m s$^{-1}$ in less than 1 km. The along-front component ($V$), dropped from the maximum $\sim 0.45 \text{ m s}^{-1}$ at the subsurface jet to zero at a distance of 1 km downstream. The along-front velocity exhibited a sharp vertical gradient with maximum ($\sim 0.45 \text{ m s}^{-1}$) positive values at the surface.

**Figure 4.** Residual, semidiurnal, and diurnal flow across the turbidity front, winter 1999. Along and across-estuary ($U$, $V$) components are shown on the left and right, respectively.
Figure 5. Velocity field across the turbidity front, winter 1999. Along (U, left) and across-estuary (V, right) components for each repetition of the transect. Repetition number is indicated on the lower left. The x-axis origin corresponds to the upstream point, and along- and across-estuary velocities are positive seaward and northward, respectively.
(toward the northern coast) and negative values near the bottom. The null velocity level was found at 4 m. In Cw-7, flows with maximum speed of $-0.15 \text{ m s}^{-1}$ observed near the bottom, just downstream of the surface front, indicated enhanced flow near-bottom toward the Argentinean coast.

Intensification of the along-estuary density gradient was also associated to the front. Surface density was 1007.9 kg m$^{-3}$ (salinity 10.48, temperature 9.71$^\circ$C) and 1011.6 kg m$^{-3}$ (15.30, 10.00$^\circ$C) upstream and downstream of the front, respectively. These values were measured at two CTD stations less than 2 km apart. Same as before, the isopycnal of 1010 kg m$^{-3}$ and the isohaline of 12, both gave a good representation of the frontal boundary.

Transect Cw-8 was transited at the end of ebb and beginning of flood, and Cw-9 was surveyed under flood tide but less intense than the earlier one. The surface front was observed at the beginning of Cw-8 but its visual surface expression vanished and was not detected along repetition Cw-9. The two-layer flow structure persisted through the flood, with the upper fresh layer, $\sim$3 m deep, moving seaward at $\sim$0.40 m s$^{-1}$ (Cw-8). The surface outflow decelerated to $\sim$0.20 m s$^{-1}$ through the flood, while the upper layer became thinner. There was significant along-front flow. At the surface it had positive velocity values of 0.20 m s$^{-1}$ and reached a negative velocity maximum of $-0.45$ m s$^{-1}$ near the bottom in the shallower part.

Ebb flow developed during the last three repetitions. In the tenth repetition (Cw-10), the along-estuary velocity was seaward in the entire water column except for a thin bottom layer. Outflow dominated in the last two repetitions in response to ebb and wind stress forcing. Before the survey and until Cw-10, light winds (speed <5 m s$^{-1}$) blew from the north-northwest (Figure 2). Upon starting Cw-10, winds increased to moderate-strong (>10 m s$^{-1}$) from the

Figure 6. Variation of the density field at the turbidity front during a tidal cycle.
Changes in the vertical stratification at the turbidity front during a tidal cycle can be observed in this series of consecutive density profiles at the central point of the transect. Labels indicate the repetition number.

4.2.2. Density Field

The evolution of the temperature, salinity, and density field was investigated. Vertical profiles were interpolated to a regular grid. Low horizontal resolution resulted in very smooth fields, except for repetitions Cw-7, Cw-8, and Cw-9 where CTD casts were taken every ~2 km and provided sufficient data to describe the variability across the front. These smooth fields do not properly represent high spatial variability and can underestimate horizontal gradients. Nonetheless, they provide information on the general changes observed along the transect and allow determination of the movement of the buoyant flow and the underlying salt wedge. The density sections for each repetition are presented in Figure 6.

4.2.3. Velocity Field

The direction of the surface current was then ~40° to the left of the southwestward wind, which suggests that wind was reinforcing the outflow. The velocity decreased and veered counterclockwise (anticyclonically in the southern hemisphere) with depth. The along-estuary component (U) decreased with depth to 0.05 m s⁻¹ near the bottom, and the across-estuary (V) component was zero at 3 m, and reached a maximum of +0.25 m s⁻¹ at 4 m.

4.3. Velocity Gradients: Flow Convergences and Shears

Fronts are by definition zones of locally intensified horizontal gradients in water properties. In general, the velocity field at fronts presents high variability. Convergence at the frontal boundary results in enhanced along-front (transverse) circulation and significant vertical motion [O’Donnell et al., 1998; Trump and Marmorino, 2003], which affects the distribution of water properties and suspended particles. Although the data available allows estimating only the across-front terms of the shear tensor (\(\partial U/\partial x\)), the information collected along the transect perpendicular to the front allows us to evaluate for the first time the strength of the convergence associated with the turbidity front in the Rio de la Plata and the spatial variability associated to it.

The across- and along-front velocities (U, V in Figure 5) showed high variability both in space and time. The across-front and vertical gradients were estimated to quantitatively evaluate the importance of across-front convergence/divergence and horizontal and vertical shear associated to the turbidity front. The along-estuary gradient of the across-front velocity (\(\partial U/\partial x\)) is presented in Figure 8a. Negative values correspond to across-front convergence. A zone of locally intensified across-front gradient (\(\partial U/\partial x\)) was observed at the surface coincident with the front and jet structure during the first eight repetitions (Cw-1 to Cw-8), suggesting convergent flow at the boundary. The maximum change in across-front velocity (\(\partial U/\partial x\)) was observed on repetition Cw-7 close to the surface, where the surface front was detected. There were no areas with intensified horizontal gradient in the last three transects. The across-front shear of the along-front velocity (\(\partial V/\partial x\)) also showed largest
Figure 8. (a) Across-front convergence of the along-estuary velocity. (b) Across-front shear of the across-estuary velocity.
Figure 9. (a) Vertical gradient of the along-estuary velocity (across-front component). (b) Vertical gradient of the across-estuary velocity (along-front component).
values at the surface, which were concentrated at the front. In the vertical, the maximum shear was found at the downstream boundary of the jet (Figure 8b).

34 The horizontal resolution of the averaged profiles ($\Delta x \approx 200$ m) was low compared with the actual frontal scale, yielding finite differences ($\Delta U/\Delta x$, $\Delta V/\Delta x$) that underestimated the actual gradients. As a result, the real values could be larger than the ones estimated here, and therefore the magnitudes in Figure 8a and Figure 8b should probably be considered as lower bounds.

35 Estimates of the vertical shear for the across-front ($U$) and along-front ($V$) velocity are presented in Figure 9a and Figure 9b, respectively. When the survey started, with strong ebb current during the first repetition, the across-front velocity was uniform from top-to-bottom and no significant vertical shear was observed. A few hours later, during the strong flood tide (repetitions Cw-3, Cw-4, and Cw-5), the vertical shear of the across-front velocity had a distinct distribution with positive values in the upper layer, negative below and the zero line located at middepth. This vertical shear distribution was related to the jet at middepth. In the sixth repetition, after ebb started, the structure remained in the center of the transect. During ebb current, in Cw-7 and Cw-8, when the surface front was sharply delineated on the surface, maximum shear was observed in the upper layer near the surface in the area upstream of the front. The shear decreased with depth. Downstream of the front, there was a sudden drop in shear magnitude, which was weak from top to bottom. In the last two repetitions, the vertical shear $\partial U/\partial z$ showed a three-layer structure, with positive shear at the surface layer and near the bottom and negative shear at middepth. The intensity and thickness of the pycnocline layer varied throughout the repetitions, but there was a persistent difference between surface and bottom salinity of 12–14 units, with a maximum observed vertical salinity gradient of 10 m$^{-1}$ at 4 m at the end of Cw-10. The three-layer structure of the shear resulted from different forces acting on the stratified water column: wind stress at the surface accelerated the current and increased positive shear; a weak negative shear at middepth coincided with the pycnocline layer; and bottom stress decelerated the ebb flow near the bottom also increasing positive shear.

36 The distribution of vertical shear of along-front velocity (Figure 9b) also showed increased shear along the jet boundaries during the first half of the survey. Strong shear at the surface resulted in the upper layer in Cw-7, Cw-8 and Cw-9 upstream of the front. In the last two repetitions, under moderate-to-strong winds a three-layer structure developed, with negative shear in the upper and bottom layers and positive at middepth. In this case negative shear in the upper layer responded to southward (negative) flow decreasing with depth.

4.4. Stability

37 The stability at the front was investigated based on the gradient Richardson number

$$R_{ig} = \frac{N^2}{S^2}$$

where $N^2 = -\frac{g}{\rho} \frac{\partial \rho}{\partial z}$ is the buoyancy frequency and $S^2 = \frac{1}{2} \left[ \left( \frac{\partial U}{\partial z} \right)^2 + \left( \frac{\partial V}{\partial z} \right)^2 \right]$ is the total velocity vertical shear. On the basis of laboratory and observational studies [Simpson and Britter, 1979; Geyer and Smith, 1987], a threshold for development of shear instability of $R_{ig} = 0.25–0.33$ has been suggested. At smaller $R_{ig}$, a stratified shear layer develops where density and velocity show a linear variation with depth [Geyer and Smith, 1987]. During the seventh and eighth repetitions (Cw-7 and Cw-8), when the front was sharply delineated at the surface close to the seaward end, the buoyant layer upstream of the front had low $R_{ig} < 0.25$. Values $R_{ig} > 0.5$ were computed.
at middepths, with maximum values $R_{ig} > 5$ at a layer 4–5 m deep. Downstream of the front, the layer with $R_{ig} < 0.25$ was thicker and no middepth maximum was observed.

Brubaker and Simpson [1999] observed a clear shear layer structure upstream of a tidal intrusion front in the lower James River, Chesapeake Bay. The density and velocity profiles across the front are examined here to investigate the occurrence of such layer of stratified shear. The vertical structure at both sides of the surface front is described in Figure 10 with data from stations located at 0.5 km upstream and 3.5 km downstream of the surface front, respectively. The density and velocity profile in the station upstream the front showed a marked linear variation with depth in the upper 4 m, where large total shear resulted in very small Richardson numbers, $R_{ig} < 0.25$. Below that upper layer, a maximum $R_{ig}$ occurred at 4.5 m. Downstream of the front, the pycnocline was deeper in contact with the bottom, and there was a secondary pycnocline at 4 m. The across-front velocity ($U$) had a two-layer structure. It decreased linearly with depth in the upper layer down to 5.5 m, where it changed sign and increased slightly with depth. There is a jump in current direction from $\sim 30^\circ$ (measured clockwise from x axis) to $270^\circ$ and veering counterclockwise below. Buoyancy frequency had its maximum value at 8 m, and a secondary maximum at 5 m. The total shear also was maximum at 5 m, but it was small top to bottom. Therefore, the Richardson number was small with values below the threshold in the upper layer and increased with depth below 5 m to a maximum of $R_{ig} \approx 5$ near the bottom. The Richardson numbers indicated that the area in the proximity of the front was very active, with favorable conditions for development of stratified shear layers and shear instabilities.

The stability parameters at peak flood and ebb were compared to assess differences in vertical structure during the tidal cycle. In Figure 11, density, buoyancy frequency ($N^2$), velocity components $U, V$ (black solid and dashed lines), speed and direction (gray lines), total shear ($S^2$) and Gradient Richardson Number ($R_i$) during (a–e) flood and (f–j) ebb flow.

Figure 11. Pycnocline, velocity, and stability at the salt-wedge: Flood and ebb condition. Profiles of density ($\sigma$), buoyancy frequency ($N^2$), velocity components $U, V$ (black solid and dashed lines), speed and direction (gray lines), total shear ($S^2$) and Gradient Richardson Number ($R_i$) during (a–e) flood and (f–j) ebb flow.
The pycnocline was shallower during ebb and there was a sharp change in velocity direction at the depth of the maximum density gradient (Figures 11g and 11h). During peak flood, with upstreamflow from top to bottom, there was a distinct middepth velocity maximum at the depth of the pycnocline (Figure 11c), just above the maximum density gradient. Maximum shear occurred right on top of the maximum velocity and corresponded to minimum values of gradient Richardson number. Richardson numbers during flood were one order of magnitude smaller than the values during ebb.

5. Discussion

The estuarine circulation in the Río de la Plata can be understood as a result of the balance among buoyancy (river outflow), wind and tides. Ambient shelf circulation has been proved to play an important role along the northern coast especially in the summer months, but we can exclude this effect in this analysis of the winter conditions in the central part of the estuary. The study area was surveyed in August 1999, during austral winter, 3 days before neap tide (Figure 2). River discharge was typical for winter season: a comparison with the total mean annual cycle (computed based on ~90 years of daily records) showed that the river input to the estuary was close to the average in the semester before August 1999, with no drought or flood affecting the area. Light winds (~5 m s\(^{-1}\)) blew during the first nine repetitions (before 1900 h) and increased to moderately strong winds (8–10 m s\(^{-1}\)) from the northeast toward the end of the 25-h observation period.
[41] To investigate the influence of wind on the circulation, we performed complex regression analysis between the observed current velocity and winds. Results indicate that the correlation improved significantly after 1900 h, with the amplitude of the complex correlation coefficient between surface current (at 1 m) and wind changing from 0.40 to 0.76. Therefore, we can speculate that buoyancy and tidal dynamics dominated during the first part of the survey on 16 August, until late afternoon, when wind forcing became important.

[42] Results indicate that there is a marked intratidal variation, with significant differences in velocity and density structure. During peak flood tide, the upper layer flow was reversed, and upstream flow occupied the entire water column. This condition was not observed during the lesser flood, when the two-layer bidirectional current persisted. The along-estuary velocity presented a marked subsurface maximum. Even though the midwater flood maximum may develop in homogeneous water columns [Prandle, 1981], in this particular case the flood maximum was associated with the vertical migration of the pycnocline. This indicates the upward limit of the bottom boundary layer [e.g., Chant et al., 2007; Stacey and Ralston, 2005].

[43] The observed variability in density field suggested dependence with the tidal cycle. To evaluate the presence of tidal modulation in stratification, we computed a bulk-stratification parameter as the density difference between surface and bottom for each CTD cast. The time series obtained is presented in Figure 12, together with the water level recorded at the nearby station and the wind speed and direction. During the first part of the survey, the stratification was maximum at the end of ebb and beginning of flood. Stratification decreased through flood tide and reached its minimum at the end of flood. The amplitude of this semidiurnal fluctuation of stratification was \( \Delta \sigma \approx 8 \text{ kg m}^{-3} \).

[44] Evidence of stability fluctuations caused by tides has been found in many estuaries and regions with freshwater influence around the world, e.g., the Hudson River estuary [Nepf and Geyer, 1996; Geyer et al., 2000], the Tamar Estuary [Uncles, 2002], the York River estuary [Sharples et al., 1994], San Francisco Bay [Stacey et al., 1999], Liverpool Bay [Sharples and Simpson, 1995; Rippeth et al., 2001] and Rhine region [Simpson et al., 1993; Simpson and Souza, 1995; Fisher et al., 2002]. In coastal regions, where horizontal density gradients are significant, the vertical shear of the tidal current acts on those density gradients to produce semidiurnal oscillations in water column stability. Simpson et al. [1990] investigated this mechanism known as “tidal straining” and the switching between stratified and mixed states over the tidal cycle defining the conditions for “strain-induced periodic stratification” (SIPS). They proposed a criterion based on the minimum horizontal density gradient necessary to produce SIPS. Considering that the average input over the ebb half cycle should be greater than the mean tidal stirring power over the same period, they obtained the following condition:

\[
\frac{1}{\rho} \frac{\partial \rho}{\partial x} > \frac{2 \varepsilon k}{3} \frac{\sigma}{\rho} \left( \frac{u_t}{h} \right) \approx 2.2 \times 10^{-5} \left( \frac{u_t}{h} \right)^2 \tag{1}
\]

where \( k \) is a nondimensional drag (0.0025), \( \varepsilon \) is a nondimensional mixing efficiency (0.004), \( u_t \) is the tidal current amplitude, and \( h \) is depth.

[45] Evaluation of this criterion in the area of study was accomplished using the ADCP and CTD observations along the central transect. With along-estuary tidal amplitude of 25 cm s\(^{-1}\) and 8 m depth, the term on the right side in (1) equals \( 2.15 \times 10^{-8} \text{ m}^{-1} \). The density gradient was estimated as the total difference between upstream and downstream points, yielding \( 6.25 \times 10^{-4} \text{ kg m}^{-2} \) (larger values occurred, but we used a minimum to compare with the threshold value). Using the mean density from Table 1, the left term in equation (1) equals \( 6.17 \times 10^{-7} \text{ m}^{-1} \), and

\[
\frac{1}{\rho} \frac{\partial \rho}{\partial x} \gg 2.15 \times 10^{-8} \tag{2}
\]

Therefore, the conditions were favorable for SIPS development.

[46] The available data do not provide information on lateral variations, so we cannot evaluate the importance of lateral straining in the dynamics. However, in this part of the estuary the bottom topography is smooth with the isobaths oriented perpendicular to the along-estuary axis and following the geometry of the Barra del Indio shoal (Figure 1). Consequently, we do not expect to observe channel-shoal lateral effects affecting the density and velocity field. Although other factors such as tidal advection of the front and increased mixing may be affecting the water column, the estimate above indicates that SIPS contributed to the changes in stratification observed during the first part of the survey, before the wind strengthened. To evaluate the potential contribution of straining to the temporal variation of the vertical stratification, we estimated the ratio \( \left[ \partial \bar{U}/\partial z \right] \left( \partial \rho/\partial x \right) / \left( \partial \rho/\partial t \right) \). Tidal straining explained between 20 and 100% of the tidal variability in density stratification. The highest variability explained by tidal straining (approaching 100%) appears near the bottom during ebb and near the surface during flood.

[47] Toward the end of the survey, under strengthening wind conditions, the stratification increased (Figure 12). This result contradicts the idea that wind stress enhances mixing at the surface and tends to destroy stratification. However, in such a large environment as the Río de la Plata with a fetch of >100 km, the southwestern wind stress had two effects. First, it advected fresher water from the northern part of the estuary to the area causing an increase in bulk stratification. Just before the survey at the central transect, increased freshwater outflow was observed along the northern coast following a 2-day pulse of strong southerly winds [Framiñan et al., 2000; Framiñan, 2005]. Enlarged flushing resulted in fresher water (salinity <10) occupying the entire water column in a 20-km band along the Uruguayan coast, approximately 60 km northeast of the area of study. The second effect of the southwestward wind was to increase mixing at the surface layer causing the deepening of the pycnocline (Figure 7, profile 12). The mechanism is similar to the effect of winds on stratification studied in river plumes and coastal currents [Münchow and Garvine, 1993; Souza and James, 1996; Fong and Geyer, 2001]. In general, in small estuaries this process may be less significant because of limited lateral scales restricting advec-
tion. In the wide Río de la Plata this effect needs to be considered.

[48] A strong near-surface frontal jet flowing toward the Uruguayan coast was observed on the upstream side of the turbidity front. The jet persisted throughout the tidal cycle, advected upstream (downstream) by flood (ebb) current by a distance of ~10 km. The jet showed semidiurnal and diurnal modulation with its maximum signature during ebb, when the front was clearly defined at the surface, and reached ~0.45 m s$^{-1}$. Near-bottom along-front flow in the opposite direction (southwestward) was observed during part of the tidal cycle. This frontal jet structure was evident in the residual field, with velocities of $+0.15$ m s$^{-1}$ near the surface and $-0.05$ m s$^{-1}$ near the bottom. Although the lateral structure of the flow at the front qualitatively resembles geostrophic balance, a comparison with the thermal wind equation indicates that the observed vertical shear is smaller than the one estimated from the horizontal density gradient. This result was anticipated because of the importance of friction in the area that contributes significantly to the lateral momentum balance [Framinhan et al., 2006]. Wind-driven flow, especially toward the end of the survey, and tidal effects may also add to the ageostrophic characteristics of the jet.

[49] Frontal jets have been observed in tidal mixing fronts on the shelf around the British Islands [Simpson, 1976; Simpson, 1981; Lwiza et al., 1991; Kasai et al., 1999]. However, in those shelf features the observed across-front flow was much smaller than the along-front velocities, which were in good agreement with geostrophic balance. At the Río de la Plata front, the flow decelerated and converged at the front, but the along-estuary (across-front) residual flow was significant and larger than the lateral velocity (along-front).

6. Summary

[50] Density and velocity observations collected across the turbidity front in winter 1999 in the central part of the Río de la Plata estuary characterized horizontal and vertical gradients, transverse flows and jets associated with the frontal boundary. A sharp surface front was observed at the end of ebb, close to the seaward end of the transect. Strong horizontal convergence of the across-front velocity and build up of along-front velocity shear indicated the ageostrophic nature of the front dynamics caused by frictional and advective processes. In the proximity of the front, enhanced transverse flow created jet-like structures flowing in opposite directions: near the surface toward the northeast (to the Uruguayan coast) and near the bottom toward the southwest. These observations provide the first direct evidence of transverse circulation associated with the turbidity front in the Río de la Plata, which is considered a key issue to sediment dynamics and the estuarine ecosystem.

[51] This study shows the significance of the spatial variability in the velocity and density fields and the importance of intratidal phenomena. The presence of flood-ebb stratification and shear asymmetries has been proposed as the driving mechanism of residual flow [Jay and Smith, 1990; Geyer et al., 2000; Prandle, 2004; Simpson et al., 1990; Stacey et al., 2001]. Our results support the hypothesis that these mechanisms might contribute significantly to the estuarine circulation in the central part of the estuary.

[52] The in situ data at the estuarine turbidity maximum validate the working hypothesis of Framinhan and Brown [1996] in their remote sensing study of the turbidity front. The surface signature observed in visible images is indeed associated with the saltwater intrusion and constitutes the surface signature of the complex frontal boundary.

[53] The analysis of this high-resolution current data set revealed features of the velocity field at the turbidity front in the Río de la Plata estuary that resemble certain characteristics observed in tidal mixing fronts found in the European continental shelf. This result is intriguing considering the different environments where these frontal zones developed. Future work will focus in investigating the mechanisms that result in similar frontal behavior.

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