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A Benthic Front in the Straits of Florida and Its Relationship to the Structure of the Florida Current

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ABSTRACT

Observations from CTD tow-yos and microstructure profiles indicate the presence of a benthic front near 450-m depth on the western side of the Straits of Florida at 27°N. The front is at midslope and approximately beneath the axis of the Florida Current. There is a 0.2 kg m^{-3} jump in density across the front, and the bottom boundary layer changes thickness from 10–20 m upslope to 50–70 m downslope of the front. Axial currents exhibit very strong vertical shear and cyclonic vertical vorticity upslope of the front in contrast to the significantly weaker vertical shears and anticyclonic vertical vorticity observed downslope. All these features, with the exception of the anticyclonic vorticity downslope of the front, are consistent with a model of barotropically forced flow through a stratified channel with a sloping sidewall. An arrested boundary layer regime is predicted to develop on the sloping sidewall upslope of a height $h_p \equiv fL/N$, where f is the Coriolis parameter, L is the width of the flat-bottomed portion of the channel, and N is the buoyancy frequency. The arrested boundary layer supports little cross-slope transport. Below h_p a cross-channel circulation cell develops in the model, driven by the bottom Ekman layer.

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Transition between the boundary layer regimes leads to convergence and frontogenesis near a height h_p on the sloping sidewall. Strong mixing at the front acts to destroy the salinity minimum signature of Antarctic Intermediate Water moving into the North Atlantic. The development of anticyclonic vorticity in the Florida Current as it flows through the Florida Straits forces the strongest flow onto the western slope, where an arrested boundary layer develops. This configuration supports strong vertical and lateral shears, all largely in geostrophic balance.

1. Introduction

While tow-yoing a CTD package eastward along 27°N in the western half of the Straits of Florida, we noticed that each downcast terminated in a thin boundary layer of nearly identical density, despite an increase in depth from 300 to 450 m, and considerable changes in the overlying waters associated with the thermal wind balance of the Florida Current. An abrupt 0.2 kg m^{-3} increase in the density of the boundary layer near 450 m came as a great surprise (Fig. 1). The boundary layer maintained this higher density to the bottom of the channel. At this latitude the western side of the Straits of Florida is of remarkably constant slope, and the abrupt change in bottom boundary layer density was perched in the middle of this slope. Why was it there?

In this paper we briefly describe the benthic front and advance an explanation for its existence. After presenting some background information on the Florida Current, the observations, which give some idea of the structure of the front, are presented next. A mechanism that generates boundary layer convergence, based on Seim and Gawarkiewicz (1999, manuscript submitted to *J. Phys. Oceanogr.*, hereafter SG99), and which may explain the existence of the front, is outlined in section 3. We note those features common to the observations and the frontogenesis mechanism and discuss how this may relate to the structure of the along-channel flow. We briefly discuss processes that may give rise to anticyclonic vorticity in the Florida Current and how, together with the frontogenesis mechanism, these processes may shape the Florida Current.

2. Background

The Florida Current is the extension of the Loop Current in the Gulf of Mexico and the origin of the Gulf Stream. The current is confined within the Straits of Florida, a channel 300 km-long between Florida and Cuba and the Bahamas Banks (Fig. 2). Decreasing depth and width in the straits cause current speeds to increase from the gulf to 27.5°N where the Bahamas Banks end abruptly and the current widens and becomes the Gulf Stream (Richardson et al. 1969).

The Florida Current plays an important role in the wind-driven circulation of the North Atlantic (Molinari 1983) and in the thermohaline circulation of the Atlantic basins (Schmitz and McCartney 1993). Of the 29 Sv (Sv $\equiv 10^6$ m^3 s^{-1}) carried by the Florida Current, Schmitz and Richardson (1991) estimate that 45% of the flow entering the Straits of Florida comes from the South Atlantic and compensates for North Atlantic Deep Water exported into the southern oceans by the thermohaline circulation. Because of the proximity of different water types, strong mixing in the straits could bring about significant water mass modification.

Despite being one of the best observed currents in the world (Leaman et al. 1987) the processes that shape the current and control its structure have not been clearly identified. Attention has been focused on quantifying transport and its fluctuations (Larsen 1992) and identifying the nature and frequency of meandering and eddy formation (Boudra et al. 1988).

3. Observations

In June 1990, M. Gregg and T. Sanford led a field program to study turbulent mixing in the Straits of Florida. Sampling was concentrated along 27°N, a section previously occupied in the Subtropical Atlantic Climate Study (Molinari 1983). The two instrument systems operated during the cruise that will be discussed below are the Multi-Scale Profiler (MSP) and a SeaBird 911 Plus conductivity–temperature–depth (CTD) package equipped with an altimeter. The MSP is a free-falling profiler that carries a CTD suite, microstructure sensors to measure turbulent dissipation rates of turbulent kinetic energy (ϵ) and temperature variance (χ), and velocity sensors to measure finescale (1–100 m) horizontal velocity. The MSP is described in detail in Winkel et al. (1996). The MSP repeatedly occupied seven stations across the channel, collecting a total of 85 profiles, and sampled to within 5–10 m of the bottom. The SeaBird CTD package was run over full depth cycles while slowly steaming (tow-yos), providing high spatial resolution sampling of the density field.

A composite of the observations reveals the mean axial velocity and density structure of the Florida Current during the cruise (Fig. 3). Maximum velocities occur in the western half of the channel over the Florida slope. Vertical shear of the axial flow increases moving upslope. In the eastern half of the channel, velocities and vertical shear steadily decrease as the Bahamas are approached. The configuration of the density field and axial velocity are consistent with an approximate geostrophic balance, and hence horizontal density gradients are greatest in the west where the strongest shears occur. The slope of near-bottom isopycnals is nearly the same as the bottom slope over the western half of the channel and the densest water in the section lies on the lower portion of the western slope.

Temporal variability is a hallmark of the Florida Current (Larsen 1992), and the sampling period for these observations was no exception. The axis of the current shifted to the east soon after the CTD section in Fig. 1 was collected, about a week into the cruise. At stations 1, 2, and 7 there are a sufficient number of MSP profiles to construct mean profiles before and after the shift, but at the other stations only a cruise average can be constructed. Figure 3 is a composite of all the data, but all subsequent graphics use the early time means where possible, because this time period is most relevant to the CTD observations presented. The most noticeable change is at station 2 where the vertical shear is noticeably weaker after the axis of the current shifts to the east.

The average thickness of the homogeneous bottom boundary layer (HBL) (defined as $\Delta\sigma_\theta < 0.005$ kg m^{-3}) and turbulent bottom boundary layer (TBL, defined as $\Delta \log \epsilon > 2$) have been determined from the MSP profiles (Winkel et al. 1992). The HBL and TBL are thickest over the eastern half of the channel, much thinner over the western slope, and are comparable in thickness at most stations. Notable exceptions occur at station 3 where the TBL is much thicker than the HBL and at the easternmost station next to the Bahamas Banks. The proximity of the Bahamas Banks enhances finescale shear and results in stronger dissipation there (see Winkel 1998); our focus in this paper is on conditions near station 3.

There appear to be distinct boundary layer regimes, one thick and one thin, and the transition between these occurs near station 3 (Fig. 4). At station 3, the TBL extends 50 m above the HBL, forming a region of intense turbulence in an area of locally elevated stratification. Mixing in the straits is strongest in the TBL where the channel is more than 300 m deep (Winkel 1998). Estimated diffusivities rise well above 10^{-4} m^2 s^{-1} in this region and fall to less than 10^{-5} m^2 s^{-1} within the core of the current. Station 3 is located about 5 km downslope of the front seen in the CTD tow-yos of Fig. 1 and lies roughly beneath the axis of the Florida Current (Fig. 3). Both the CTD and MSP observations suggest a transition or front in the boundary layer on the western slope near a depth of 450 m.

Below 500 m temperatures (*T*) drop below 7°C and salinities (*S*) fall to 34.9 psu, properties consistent with Antarctic Intermediate Water (AAIW) (Atkinson 1983). The front separates AAIW and saltier, warmer water that originates from the subtropical South Atlantic (Schmitz and Richardson 1991) (Fig. 5). In Fig. 5, *T*–*S* plots from profiles collected at the front (CTD 17, shown in Fig. 1) and downslope of the front (CTD 29) are compared to the *T*–*S* envelope for the Florida Current defined by Schmitz et al. (1993). The *T*–*S* properties from downslope of the front closely follow those from the subtropical South Atlantic [see Fig. 4b of Schmitz et al. (1993)], whereas *T*–*S* properties at the front follow a mixing line between the densest water downslope of the front and 10°C water. The MSP profile from station 3 shown in Fig. 4 closely follows CTD 17, suggesting they sampled the same water mass.

The strong dissipation rates and *T*–*S* relationships imply mixing at the benthic front is removing the salinity minimum associated with the AAIW. This is consistent with the *T*–*S* envelope of the Florida Current of Schmitz et al. (1993) that shows the freshest water in the straits falling along the same mixing line as our profiles near the front (Fig. 5). The circulation and mixing associated with the front is apparently one of the processes responsible for removing the salinity minimum signature.

4. Comparison with a model

Seim and Gawarkiewicz (1999, manuscript submitted to *J. Phys. Oceanogr.*) present a simple model of the spinup circulation in a channel with a sloping sidewall. An aspect of the model that distinguishes it from previous spinup studies is the way in which it is forced. By specifying the value of the streamfunction at the channel walls, the flow is forced barotropically and drives a steady along-channel flow, but leaves the shear and stratification free to be determined by the model. Previous studies (e.g., Buzyna and Veronis 1971) impose a surface stress that elicits a transient response that brings the fluid back to a rest state. With barotropic forcing, a persistent cross-channel circulation develops in unstratified conditions. A transverse density gradient develops in a stratified flow, which leads to a thermal wind balance in the boundary layer and a slowdown of the secondary circulation.

In the model two distinct regimes develop during spinup, an upslope arrested boundary layer regime (Garrett et al. 1993) and an abyssal Ekman layer-driven regime (Fig. 6). The arrested boundary layer occurs where stratified flow overlies a sloping boundary; it supports almost no cross-slope transport, is thinner than an Ekman layer, and can support strong vertical shears of alongslope flows in a thermal wind balance, requiring little bed stress. This last characteristic is the origin of an alternate name for these regimes, slippery boundary layers. Strong along-channel flows drive strong cross-channel flows in the bottom Ekman layer, but not in the arrested boundary layer, and this leads to convergent flow in the boundary layer. The Ekman circulation pushes the densest water part way up the slope and draws less dense water into the downwelling corner, setting up a cross-channel density gradient. Under certain conditions, the convergence zone lies on the sloping sidewall boundary, and fluid is expelled from the boundary layer in a narrow region, forming a front. The front defines the upslope extent of a cross-channel circulation cell driven by the bottom Ekman layer; fluid is expelled from the boundary layer at the front and a middepth return flow closes the loop. In the model, convective adjustment at the front, as dense water moves over lighter water, enhances mixing near the front (shown as stippling in Fig. 6).

There are some general features common to the model and the observations. The transition between regimes should occur at a height $h_p \equiv fL/N$ above the deepest part of the channel, where f is the Coriolis parameter, L is the width of the flat-bottomed portion of the channel, and N is the buoyancy frequency. Idealizing the geometry as approximately flat between 40 and 70 km, and using an average $N = 7.5 \times 10^{-3}\ \mathrm{s}^{-1}$, $f = 6.6 \times 10^{-5}\ \mathrm{s}^{-1}$, and $L = 30$ km, we estimate $h_p = 260$ m for the Florida Straits. For a channel 700 m deep we expect the front to form at about 440 m, which is close to that observed (Fig. 1). In both the model and the observations the well-mixed bottom boundary layer is thickest downslope of the front and thins upslope of it (cf. Fig. 4 with Fig. 9 in SG99).

Upslope of h_p we observe strong vertical shear of the axial flow in the Florida Current (solid lines in Fig. 7), consistent with the model predictions of an arrested boundary layer upslope of the benthic front. Near the front (dashed–dotted line, Fig. 7) the strong shear is interrupted by a thick region of weak shear, possibly the signature of fluid expelled from the boundary layer in which momentum is well mixed. East of the benthic front (dashed lines, Fig. 7) shear is considerably smaller, consistent with model results that show weaker shear downslope of the front. The reversal in the sign of the shear in the upper 200 m near the Bahamas, as in stations 6 and 7 in Fig. 7, is not explained by the model and is likely due to other factors, for example, interaction with the Little Bahama Bank (Leaman and Molinari 1987).

Both the model and the observations display a rapid decrease in the depth-averaged along-channel current upslope of the benthic front. In the model this occurs over the arrested boundary layer and is a consequence of maintaining the same boundary layer structure as the total depth decreases (i.e., the boundary layer occupies a larger fraction of the water depth).

Interestingly, Boudra et al. (1988), in their study of instabilities of the Florida Current, use the cross-stream motion field at the end of a model initialization procedure as the small perturbation to test stability of the flow field. They vary the along-channel wavenumber of the perturbation to find the fast-growing mode of instability. The form of this cross-stream field (in particular their Fig. 13b) is qualitatively similar to that in SG99, with strong convergence occurring in the boundary layer over the middle of the slope of the Florida side and a middepth return flow, together forming a cross-channel circulation trapped in the lower half of the channel. They use this form of perturbation in recognition of the link between variations in transport of the Florida Current and the adjustment process exemplified by the cross-channel circulation.

The adjustment timescale for the secondary circulation to become established in the model is approximately the stratified spinup time $\tau = h_p/(\delta_E f)$, where $\delta_E = (2A_\upsilon/f)^{1/2}$ is the Ekman layer thickness and A_υ is the vertical eddy viscosity. Using the values of h_p and f above and $\delta_E \approx 10$ m, τ is O(5 days). Larsen (1992) observes that weekly transport variations in the Florida Current are nearly as strong as annual variations. Given the rapid pace of oscillations in the Florida Current transport, the spindown timescale suggests that the flow is often in adjustment and that cross-channel circulation should be significant.

A basic characteristic of the Florida Current not captured by the SG99 model is the decrease in depth-averaged northward flow (υ) to the east of the surface jet. The decrease is apparent in most of the comprehensive observations of the current (e.g., Richardson et al. 1969;Leaman et al. 1987). Figure 8 estimates $\upsilon(x)$ from the MSP observations. Though uncertainties are large, around 0.1 m s^{-1} owing to a lack of absolute velocity reference (Winkel 1998), it is clear that the velocity falls off rapidly to the east; $\partial\upsilon/\partial x \approx -0.6\ \mathrm{m\ s}^{-1}/5 \times 10^4\ \mathrm{m} = -1.2 \times 10^{-5}\ \mathrm{s}^{-1}$. If we use only the surface current we find $\partial\upsilon(0)/\partial x \approx -3 \times 10^{-5}\ \mathrm{s}^{-1} = -0.46f$, much closer to the value reported by Stommel (1965).

The depth-averaged MSP observations yield a shear of $-0.18f$ to the east of the axis of the Florida Current. Both the change in planetary vorticity and vortex squashing associated with shoaling bathymetry will affect the relative vorticity. The Florida Current flows eastward along 23°N for several hundred kilometers before turning northward. The bottom is also shoaling, from more than 2000 m off the western tip of Cuba to less than 800 m at 27°N (Fig. 2). However, bottom water off Cuba does not pass through the straits. The densest water observed at 27°N is AAIW and has a density of 27.46 σ_θ, and it is more appropriate to consider changes in thickness of the water column above this horizon as it passes through the straits. In an edited version of the National Oceanographic Data Center database (Curry 1996; Lozier et al. 1995), this density surface shoals from about 900 m in the western Caribbean south of Cuba to an average of about 600 m at 27°N. Similar to the cartoon of Stommel (1965), this density surface develops considerable transverse slope as it enters the Florida Straits. At 27°N the isopycnal extends from 750 m at the bottom of the channel to 450 m, terminating in the benthic front. If we assume potential vorticity is conserved along the current,

$$\frac{\partial}{\partial y}\left(\frac{f + \zeta}{H}\right) = 0.$$

If $\zeta = 0$ in the Caribbean, we find $\zeta = -f_{27°N} + (600/900)f_{21°N} = -0.46f$. This simple calculation assumes no friction or entrainment between the Yucatan Channel and 27°N, and both effects will act to reduce the magnitude of the relative vorticity, as needed to match the observations. Thus, the development of anticyclonic vorticity is consistent with potential vorticity conservation in the presence of friction and mixing.

5. Conclusions

We propose that the benthic front observed in the Florida Current is produced by a secondary (cross-channel) circulation that brings the downstream flow into adjustment with the boundary conditions. Dynamically, the benthic front marks a transition in flow regimes. Upslope of the front an arrested boundary layer regime exists in which cross-slope transport is minimal, alongslope shear is very strong, and the combination of decreasing water depth and identical vertical profiles of alongslope flow leads to pronounced cross-slope shear of (υ). Downslope of the front an Ekman layer–like boundary layer exists that supports significant cross-channel transport, is thick relative to the arrested boundary layer, and that supports much less vertical shear of the along-channel current than in the arrested boundary layer regime. The vertical shear that does exist above the Ekman layer results from a thermal wind balance. The cross-channel density gradients required for the thermal wind balance are made possible by the presence of the channel sidewalls.

The model predicts convergence rates of $O(1–2\ \mathrm{m^2\ s^{-1}})$ when the front occurs on the slope (SG99). The model parameters are very roughly consistent with the geometry and flow rates of the Florida Current, and therefore the model provides a gross estimate for cross-channel convergence rates in it. Assuming the benthic front is active over the 200-km length of the straits, 0.2–0.4 Sv will be ejected from the boundary layer into the interior. The benthic front may play an important role in water mass modification, and in particular, in modifying AAIW as it moves into the North Atlantic.

The well-documented high-frequency transport variations [i.e., weekly (Larsen 1992)] and spinup circulation we have described are intimately linked. The transport variations will continuously force the cross-channel circulation, with the likely result that the cross-channel circulation will be quite vigorous because the spinup timescale and scale of transport variations are similar.

Anticyclonic vorticity in the Florida Current, together with the frontogenesis mechanism above, define a set of processes that may shape the Florida Current. As the Loop Current flows into the Florida Straits, potential vorticity conservation leads to the development of negative vorticity. As a result, large velocities develop over the western slope, a secondary circulation driven by the bottom boundary layer develops, and the interaction of the stratification and sloping bottom produces an arrested boundary layer on the upper portion of the slope. Convergence in the boundary layer produces the observed front, which is a physical marker of the transition between boundary layer regimes.

Lateral variations in vertical shear of the axial flow are related to the transition from Ekman layer to arrested boundary layer. The region of positive vorticity to the west of the core is a reflection of the arrested boundary layer, while to the right of the core, potential vorticity conservation produces negative vorticity. The strong horizontal and vertical shears set the stage for instability of the current after it leaves the straits.

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Figures

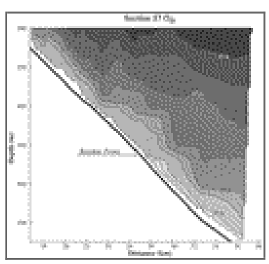

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Fig. 1. Contour plot of density as a function of cross-channel distance and depth on the western slope of the Straits of Florida at 27°N. Isopycnals nearly parallel the bottom slope. Near 450 m, the density in the bottom boundary layer abruptly increases, marking the position of a benthic front.

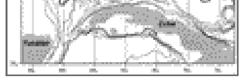

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Fig. 2. Map of the Florida Straits. The observations were collected along 27°N, shown as a heavy black line.

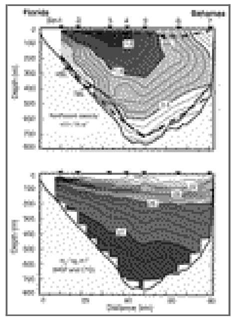

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Fig. 3. (a) Mean axial velocity structure (contours every 0.1 m s^{-1}) and (b) density structure (contours every 0.2 kg m^{-3}) of the Florida Current at 27°N based on all MSP and CTD observations. The view is to the north. The thickness of the turbulent boundary layer (TBL) is much larger than the homogeneous boundary layer (HBL) near station 3, the site nearest the benthic front, implying the ejection of boundary layer fluid into the interior near this site. The front is not apparent in the density contours because of averaging and coarse contouring.

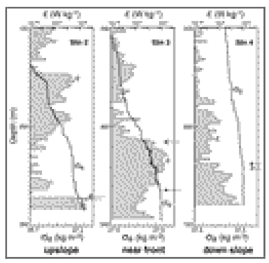

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Fig. 4. Profiles of the lowest 200 m of density and ϵ from upslope, near, and downslope of the bottom boundary layer front. Arrows mark the depths of the homogeneous (filled arrows) and turbulent (open arrows) boundary layers at each of the sites. The boundary layers are of similar thickness except near the front, where the turbulent layer is much thicker than the homogeneous layer. It is interpreted to signify fluid expelled from the boundary layer at the front. The thin boundary layer upslope of the front is consistent with an arrested boundary layer regime.

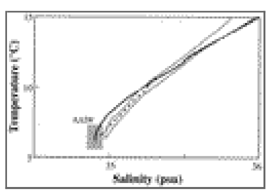

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Fig. 5. A *T–S* plot of CTD 17 (dashed) at the front (shown in Fig. 1), CTD 29 (solid black), downslope of the front, and the *T–S* envelope of the Florida Current in Schmitz et al. (1993). Mixing at the front removes the salinity minimum characteristic of AAIW.

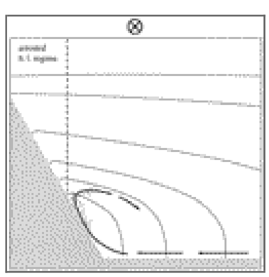

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Fig. 6. Cartoon of the cross-channel circulation in the model of SG99. Thin lines are isopycnals and axial flow is into the page. A benthic front perched on the slope separates the upslope arrested boundary layer (ABL) regime from the downslope cross-channel circulation driven by the bottom Ekman layer. The Ekman circulation pushes the densest water part way up the slope,

and draws less dense water into the downwelling corner, setting up a cross-channel density gradient. The resulting thermal wind balance supports strong vertical shear. Convection enhances mixing rates near the front (stippling). The bottom mixed layer is thick where the Ekman layer circulation exists because of convective adjustment, and is thin upslope of the front in the ABL regime.

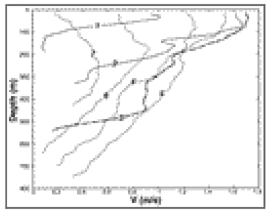

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Fig. 7. Average velocity profiles from each of the MSP stations across the strait. The two stations west of the benthic front (1, 2) show remarkably strong vertical shear, the station near the front (3) shows an upper and lower shear region, separated by a low shear region at the depth of the front, and progressively weaker vertical shear occurs east of the front (4–7).

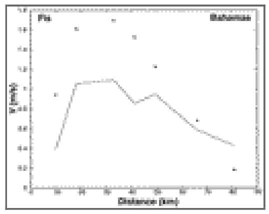

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Fig. 8. Depth-averaged (line) and surface (asterisks) velocity across the strait from the MSP observations. Errors in absolute velocity are fairly large, around 10 cm s^{-1}, but the jet is still obvious, bounded by regions of strong shear.

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