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Shelfbreak frontal structure and processes north of Cape Hatteras in winter

by Glen Gawarkiewicz[1,2], James Churchill[1], Frank Bahr[1], Chris Linder[1] and Craig Marquette[1]

ABSTRACT

The circulation north of Cape Hatteras is complicated by the proximity of the shelfbreak front, the Gulf Stream, and convergent shelf flow from the Middle and South Atlantic Bights. A three-week cruise in this region in January/February, 2005 was undertaken in order to study the structure of the shelfbreak front as it terminates near Cape Hatteras and to quantify the freshwater transport from the Middle Atlantic Bight shelf into the Gulf Stream. Two strongly contrasting conditions were identified. Early in the cruise, the Gulf Stream directly abutted the shelfbreak at Cape Hatteras and drove a northward flow over the continental shelf as far north as 35°45′N. All of the Middle Atlantic Bight shelf water terminated by 35°30′N. Ten days later, the Gulf Stream had moved away from the shelfbreak south of Cape Hatteras and strong winds from the north were present. During this time, the shelfbreak frontal jet was strong (maximum southward velocity of approximately 0.5 m s^{-1} with a Rossby number of 2) and abruptly turned eastward and offshore between 35°35′N and 35°45′N. Freshwater transport eastward from the shelfbreak jet was 7.4 mSv and southward over the shelf was 19.9 mSv, giving a total freshwater transport of 27.3 mSv. This likely represents an upper bound due to the strong wind forcing. Implications of these results for the freshwater budget of the Middle Atlantic Bight shelf, stability properties of the shelfbreak front in this region, and the formation of "Ford water" in the Gulf Stream are discussed.

1. Introduction

Richard W. Garvine was a coastal oceanographer with a broad range of interests. His work covered a wide variety of topics stretching from estuarine flows to Gulf Stream instabilities. A unifying theme of his research was the circuitous path which freshwater takes from inland rivers, through estuaries to the continental shelf, and across the shelfbreak into the deep ocean. As presented in the Biographical Notes and Bibliography earlier in this issue, Rich made major contributions in understanding estuarine dynamics, estuary shelf-coupling, shelfbreak dynamics, and the variability and instability of western boundary currents.

A particularly important issue relating to cross-shelf transport of freshwater from

1. Physical Oceanography Department, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts, 02543, U.S.A.

2. Corresponding author. ***email: ggawarkiewicz@whoi.edu***

continental shelves is whether there are local geographical areas of enhanced offshore transport of freshwater, or whether cross-shelf flows are relatively evenly distributed spatially alongshelf. In the Middle Atlantic Bight, it appears from Lagrangian measurements (Lozier and Gawarkiewicz, 2001) that the cross-shelf flow (and presumably associated offshore transport of freshwater) is indeed evenly distributed with alongshelf distance north of the Cape Hatteras region. However, the area around Cape Hatteras is clearly a special region in terms of cross-shelf transport of shelf water to the continental slope. A Lagrangian trajectory illustrating the cross-shelf flow near Cape Hatteras, from a drifter initially launched north of Georges Bank, appears in Figure 1. The alongshelf transport from the outer shelf of the Middle Atlantic Bight north of Cape Hatteras abruptly shifts to offshore motion and entrainment into the Gulf Stream. It is well known that there is strong alongshelf convergence of flows, southward from the Middle Atlantic Bight and northward from the South Atlantic Bight, which leads to sustained cross-isobath flow in the mean over long time scales (Bigelow, 1933; Savidge and Bane, 2001; Savidge, 2002). This area is strongly influenced by advection and water masses from the South Atlantic Bight, Middle Atlantic Bight, slope waters of the Middle Atlantic Bight, and Gulf Stream. A salinity balance for the region has been constructed from a combination of moorings and hydrographic data (Kim *et al.,* 2001), and the complexity of the heat budget in spring has been described by Flagg *et al.* (2002).

Ford *et al.* (1952) identified the Middle Atlantic Bight shelf water in sections of the Gulf Stream farther east. This shelf water mass embedded in the Gulf Stream has been called “the Ford water.” This water mass has been studied in detail by Kupferman and Garfield (1977), Lillibridge *et al.* (1990), and Wood *et al.* (1996). However, processes by which the Ford water is injected into the Gulf Stream have not been well studied, nor are the dominant space and time scales of this exchange well established.

Other issues still remain unresolved relating to the flow variability and associated freshwater transport despite extensive studies in the area. Pietrafesa *et al.* (1994) have shown that Middle Atlantic Bight shelf water frequently crosses Diamond Shoals and appears in the South Atlantic Bight, particularly when winds are from the north. Churchill and Berger (1998) defined the Hatteras Front as the boundary between this southward-penetrating Middle Atlantic Bight shelf water mass and the ambient South Atlantic Bight shelf water. More recently, Savidge and Austin (2007) have made high-resolution measurements of the Hatteras Front and resolved the variability as well as the cross-isobath transport associated with the Hatteras Front. A schematic cartoon of the circulation in this region appears in Figure 2, identifying the different frontal zones.

A particularly important aspect of the fate of freshwater transport near Cape Hatteras is the role of the shelfbreak front and frontal jet in the offshore transport of Middle Atlantic Bight shelf water onto the continental slope. The mean characteristics of the jet have been quantified from a climatology (Linder and Gawarkiewicz, 1998). The region just north of Cape Hatteras is particularly complicated because of Gulf Stream incursions onto the shelf (e.g. Gawarkiewicz *et al.,* 1992, 1996). Churchill and Cornillon (1991) identified flows

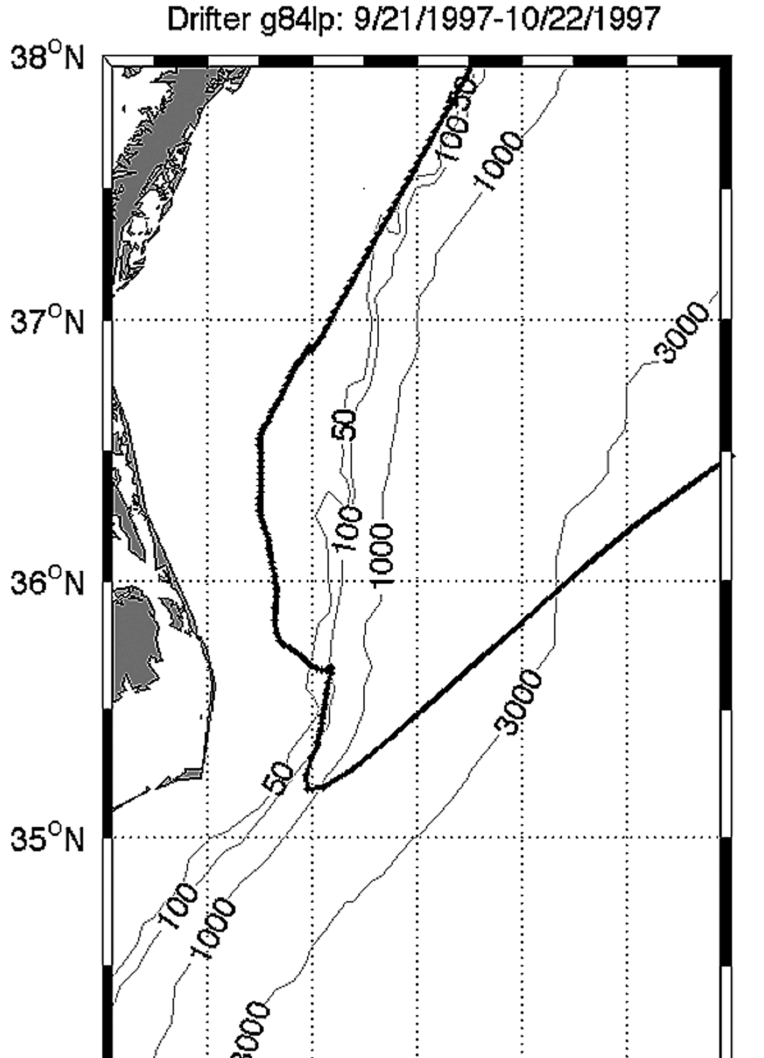

Figure 1. A drifter trajectory showing alongshelf flow from the outer shelf of the Middle Atlantic Bight extending southward to Cape Hatteras. At Cape Hatteras, the drifter abruptly moves to the northeast, entrained in the Gulf Stream. The drifter included a drogue at 15-m depth.

passing from the shelfbreak to the north wall of the Gulf Stream in this area, and Gawarkiewicz and Linder (2006) have examined the statistics of drifter trajectories crossing the 1000-m isobath and found that there were two primary modes—a slow passage across the slope region to the Gulf Stream well north of Hatteras (20–100 km),

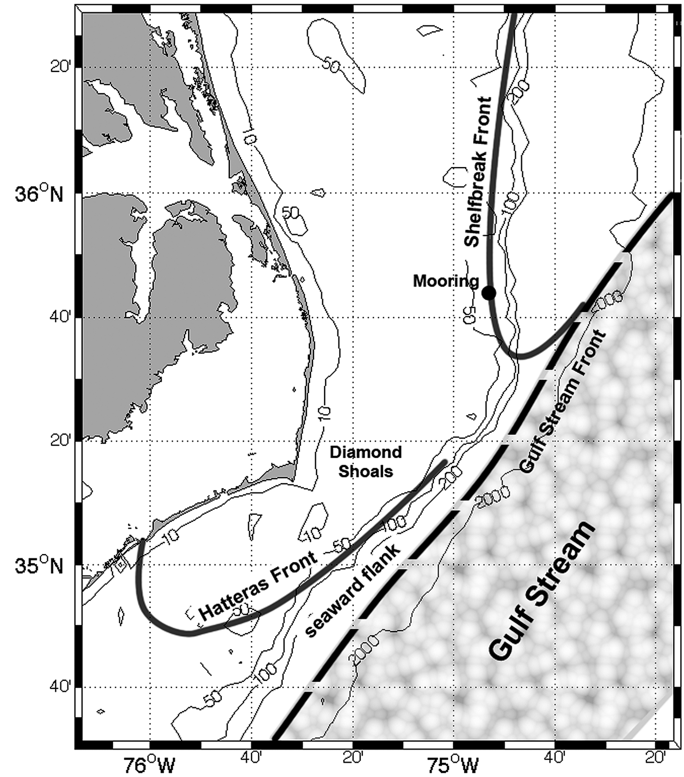

Figure 2. A schematic of the circulation in the vicinity of Cape Hatteras. Three important fronts in this region include the Gulf Stream, the Shelfbreak Front, and the Hatteras Front. The mooring at the 80-m isobath is also denoted.

with a large radius of curvature, and a rapid mode occurring in close proximity (within 20 km) of Cape Hatteras with high velocities and a small radius of curvature. Churchill and Gawarkiewicz (2009), using the same data set employed in this study, have examined the structure and dynamics of eddies over the continental slope and their origin.

In order to learn more about the offshore flow northward of Cape Hatteras and the role of the shelfbreak front in transporting the Middle Atlantic Bight water offshore and into the Gulf Stream, we performed a field experiment in January/February, 2005. The primary goal was to relate the variability of key frontal systems (the shelfbreak and Hatteras fronts) to larger-scale forcing (winds and Gulf Stream). In Section 2, we describe the study area, sampling strategy, observations and meteorological forcing. The temporal evolution of the shelf flow north of Cape Hatteras is described in Section 3, while the shelfbreak frontal structure and related dynamical properties appear in Section 4. The cross-shelf transport across the shelfbreak as well as the freshwater transport appears in Section 5. The discussion in Section 6 includes a comparison of our estimate of freshwater transport over the Hatteras region in winter with previous observations farther upstream as well as from prior studies of this region. The conclusions appear in Section 7.

2. Sampling strategy and methods

This experiment was part of the Frontal Interactions Near Cape Hatteras (FINCH) program funded by the National Science Foundation. The field work involved joint sampling with two ships during winter and summer to contrast the role of seasonal stratification in cross-shelf exchange in the region. Results from the summer cruise are described by Savidge and Austin (2007). The northern component of the observational program focused on identifying upstream conditions (water masses and frontal positions) for the Hatteras Front in addition to quantifying cross-shelf exchange with particular emphasis on the offshore transport of Cold Pool water (subsurface cold shelf water in the Middle Atlantic Bight; e.g. Houghton *et al.,* 1982). The scientific goals for this component included studying the evolution of the shelfbreak front as it approached Cape Hatteras in order to determine the role of the front in both Ford water formation and, more generally, offshore transport of freshwater.

The data were collected from January 15 to February 4, 2005 using the R/V *Oceanus.* The sampling was concentrated between 35°15′N and 36°45′N, essentially extending from Diamond Shoals to 150 km northward. The hydrographic sampling was conducted with both a Scanfish towed undulating vehicle as well as traditional vertical casts. The initial sampling was to include four cross-shelf sections, two north of 35°30′N and two south of 35°30′N, using the Scanfish. However, after completing the first two grids, the winch for the Scanfish was drenched by a large wave and malfunctioned. This necessitated sampling with traditional vertical casts using the CTD.

Transects were sampled primarily along the 1000-m isobath using the CTD in order to resolve the offshore flow of shelf water and compute the freshwater transports. On January 22, a strong winter storm with forecasts for thirty foot waves threatened the region, prompting the Oceanus to steam to Morehead City, NC for shelter and to obtain a part for repair of the winch. From January 24 to February 3 the sampling continued with a mixture of Scanfish sections across the shelf and CTD transects along the 1000-m isobath. In the second half of the cruise, sampling was extended farther to the north in order to resolve strong offshore flows, which further analysis revealed to be due to a series of shelfbreak eddies over the continental slope (Churchill and Gawarkiewicz, 2009).

The Scanfish vehicle contained a SeaBird 911+ CTD. The CTD sensors were post-calibrated after the cruise and were accurate to 0.05 in salinity and 0.002°C in temperature. The undulations of the Scanfish were processed such that one up and one down cycle were averaged into a single vertical profile. Further details on the processing (for a SeaSoar vehicle) appear in Gawarkiewicz *et al.* (2004). Horizontal spacing between profiles varied with water depth, but was typically between 0.5 and 2.0 km. The Scanfish undulated between 2 m and 120 m (or to 5 m above the bottom in water shallower than 120 m). The data were processed to provide 2-m resolution in the vertical. The ship was equipped with a

150 kHz RDI Acoustic Doppler Current Profiler (ADCP) which was in use throughout the cruise. The vertical bin size was set at 4 m. The shipboard ADCP data presented here were de-tided using harmonics generated from a barotropic finite element model of the east coast of the United States kindly provided by Dr. B. Blanton of the University of North Carolina. The average of the absolute value of the tidal corrections while in the study area was 0.028 m s^{-1} and the majority of the corrections (63%) were less than 0.05 m s^{-1}.

In addition to the hydrographic sampling, we deployed a thermistor chain and one bottom-mounted ADCP. The ADCP was a RDI 300 kHz Workhorse model. Vertical bin size was 1 m with blanking of the lowest 3 m and the uppermost 8 m. The expected instrumental error in the velocity measurements based on the choice of bin size and temporal averaging was less than 0.01 m s^{-1}. The ADCP was deployed at the 80-m isobath at 35°44.60′N and 74°52.93′W. The thermistor chain was deployed next to the ADCP at 35°44.50′N and 74°52.95′W. All thermistors were Onset Tidbit models with an accuracy of 0.2°C and a time constant of 10 minutes. Their sampling interval was set to 15 minutes. Each mooring was outfitted with thermistors at 5-m intervals through the water column.

The meteorological forcing during the cruise was as follows. Winds were strong with a mean stress of 0.17 Pa (Fig. 3) and predominantly from the north throughout the cruise, with brief periods of southwesterly wind (Fig. 4). Latent and sensible heat fluxes varied significantly during the cruise. As shown in Figure 3, the maximum hourly-averaged sum of the latent and sensible heat losses from the ocean were 1080 W m^{-2}, during which the air temperature was −4°C. However, over the two-week time period of the mooring deployment, the mean measured sum of the latent and sensible heat losses were 270 W m^{-2}. It should be noted that the ship passed from the extremely cold surface waters over the shelf to the extremely warm surface waters of the Gulf Stream during the study and so there are substantial spatial differences in the heat losses.

3. Temporal variability north of Cape Hatteras

During the course of the cruise, there was substantial variability due to both wind forcing as well as Gulf Stream variability in the vicinity of Cape Hatteras. We will first describe the temporal variability of the shelf flow through presenting the currents and temperature time series at the 80-m isobath at 35°45′N. We will then present two large-scale surveys and describe the alongshelf variability north of Cape Hatteras.

The velocities from the ADCP at 80-m depth offer a detailed view of outer shelf currents in the northern portion of our study region. The ADCP velocities presented here (Fig. 4) have been rotated into a local coordinate system oriented so that the time- and vertically-averaged mean cross-shelf velocity is zero. The alongshelf axis of this system is oriented at 356°T, which is roughly aligned with the 80-m isobath (Fig. 2).

The rotated velocities show strong alongshelf flows frequenting the outer shelf over the duration of our study (Fig. 4b). Averaged over depth and time, the mean alongshelf velocity was directed to the south with a magnitude of 0.2 m s^{-1}. Superimposed on this mean flow were strong alongshelf velocity fluctuations, which produced frequent reversals

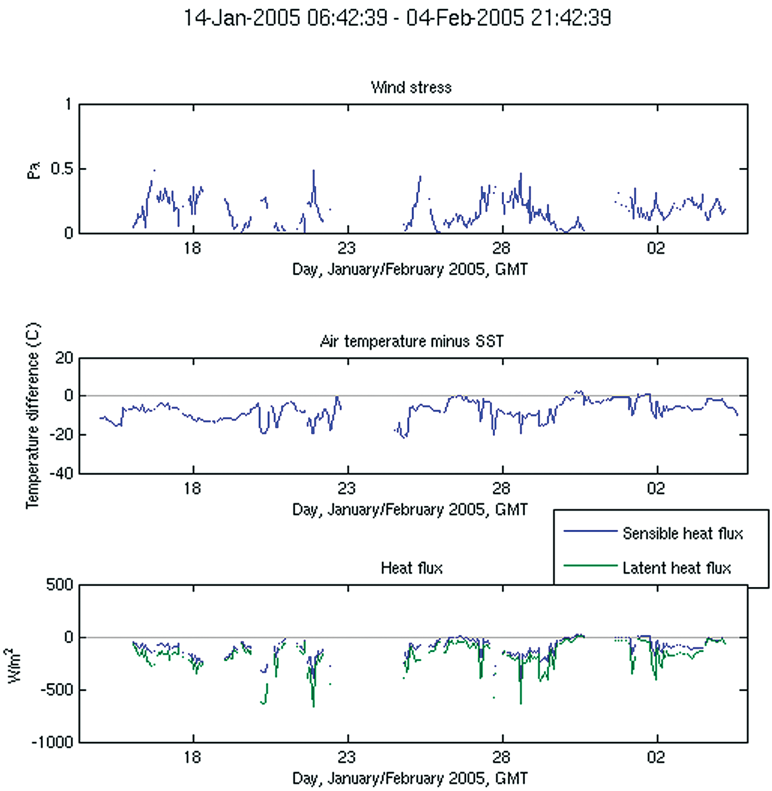

Figure 3. The meteorological forcing during the winter cruise. The upper panel is the wind stress measured on the R/V *Oceanus,* the middle panel shows the air-sea temperature difference, and the bottom panel shows the latent and sensible heat fluxes measured from the ship.

in the alongshelf flow. The strongest alongshelf velocities were directed to the south and appeared during periods of strong (>10 m s^{-1}) southward or southeastward wind. Southward alongshelf flows exceeding 0.5 m s^{-1} were recorded on four separate occasions. Northward flows were almost always associated with northward winds and did not attain the magnitudes of the southward currents.

To quantify the relationship between the wind and alongshelf velocity, we carried out a correlation analysis relating the depth-averaged velocity with the alongshelf wind stress (computed from wind speed by the method of Large and Pond, 1981). A 50-hr half-power point filter was applied to both time series used for the analysis. The maximum correlation of $R^2 = 0.79$ was seen at a lag of 14 h, with the wind leading. This correlation estimate was significantly in excess of zero at the 95% confidence level, indicating a strong relationship between the wind and alongshelf flow.

At most times, the temperatures measured at the 80-m depth site showed little vertical variation (Fig. 4d). However, relatively warm water appeared near the bottom on three

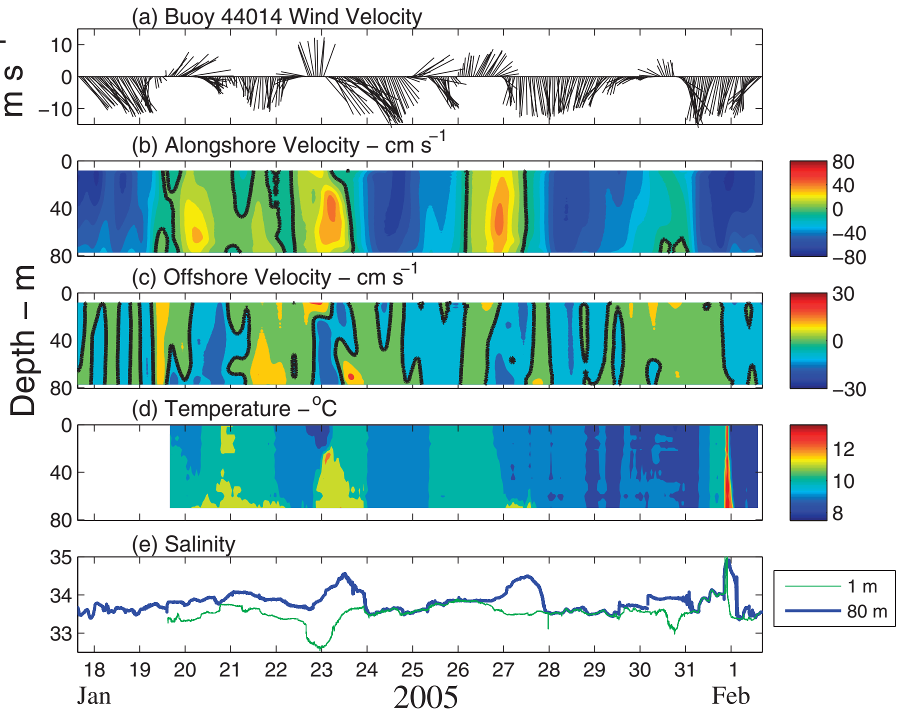

Figure 4. Temporal variations of (a) wind speed, (b) alongshore velocity, (c) cross-shelf velocity, (d) water column temperature, and e) surface and bottom salinity at the edge of the continental shelf. The moorings were deployed at the 80-m isobath along 35°45′N, at the location shown in Figure 2.

occasions (January 20–21, 23 and 27). All of these events may have been the result of wind-driven upwelling, bringing warmer offshore water onto the outer shelf. On each occasion, the appearance of warmer near-bottom water followed the alongshore wind when the wind was directed northward (upwelling-favorable) and the near-bottom cross-shelf flow was directed onshore (Fig. 4). Anomalously warm water, with temperatures >13°C, appeared at the mooring on a fourth occasion, on January 31. This water was distributed over the water column and appeared during a period of strong downwelling-favorable (southward) alongshelf winds, and so was not likely to have been carried onto the shelf by upwelling circulation. This water was of high salinity, >35, and so was likely of Gulf Stream origin.

Two large-scale Scanfish surveys also show the joint impact of Gulf Stream and wind forcing. On January 19–20, five cross-shelf transects were sampled just north of Cape Hatteras (Fig. 5). The temperature (Fig. 5, left panel) and salinity (Fig. 5, right panel) fields show large gradients between 35°30′N and 35°45′N (along the mooring line). At this time, winds reversed and were from the west while the Gulf Stream directly abutted the shallow topography near Cape Hatteras. Shipboard ADCP records indicate northward flow over the

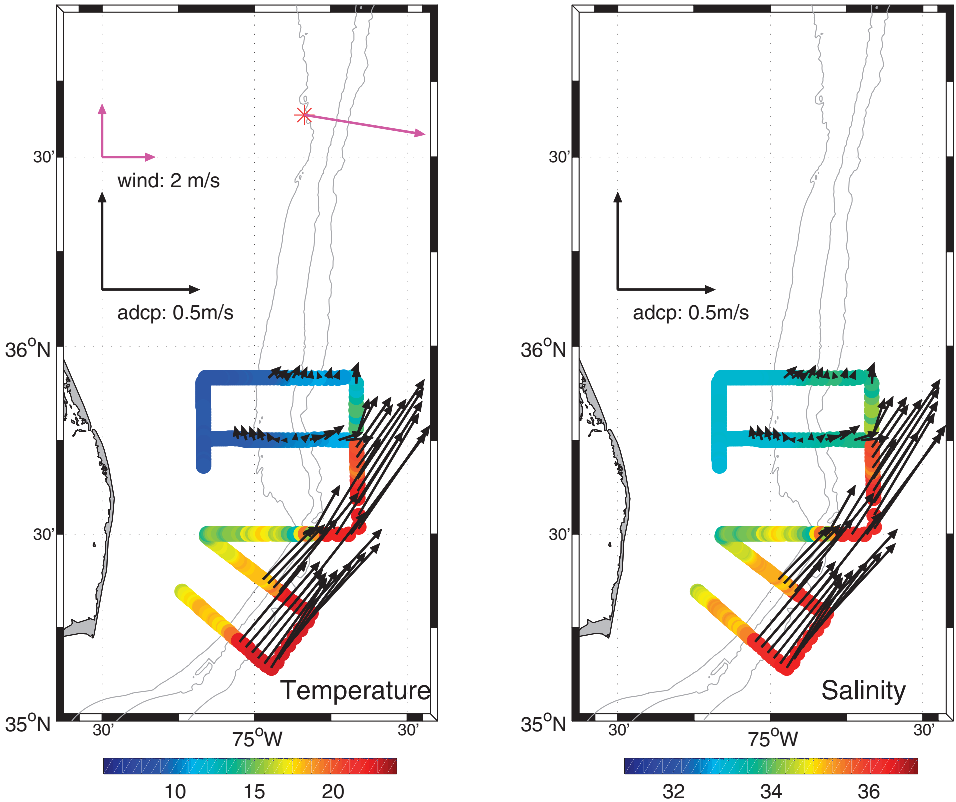

Figure 5. Plan views of (left panel) temperature and (right panel) salinity measured from a survey on January 19–20, 2005. Velocity vectors measured by the shipboard ADCP are also plotted. The velocities are from 4 m-bins centered at 25-m depth. Note the strong velocities in the Gulf Stream directed to the northeast as well as the northward flow over the continental shelf extending as far northward as 35°55′N. Winds averaged over the time period of the survey appear as the magenta arrow near 36°38′N in the left panel.

outer continental shelf even as far north as 35°55′N. Note that because of the rough seas, much of the ADCP data in shallow water was lost due to excessive noise.

In contrast, on January 29–30, after several days of winds from the north, the entire outer shelf water mass was flowing southward with a local velocity maximum in the shelfbreak jet. Both the temperature (Fig. 6, left panel) and salinity (Fig. 6, right panel) fields indicate continuity of the shelf water mass at least as far south as 35°45′N. The strong cross-shelf flows seen over the 1000-m isobath were due to shelfbreak eddies, as analyzed and reported by Churchill and Gawarkiewicz (2009). Note that there were strong cross-slope flows of up to 0.30 m s^{-1} in close proximity to the shelfbreak due to the shelfbreak eddies. For the time period of January 25–February 2, the Gulf Stream did not directly abut the continental slope, as shown in Figure 5. The shipboard ADCP-derived

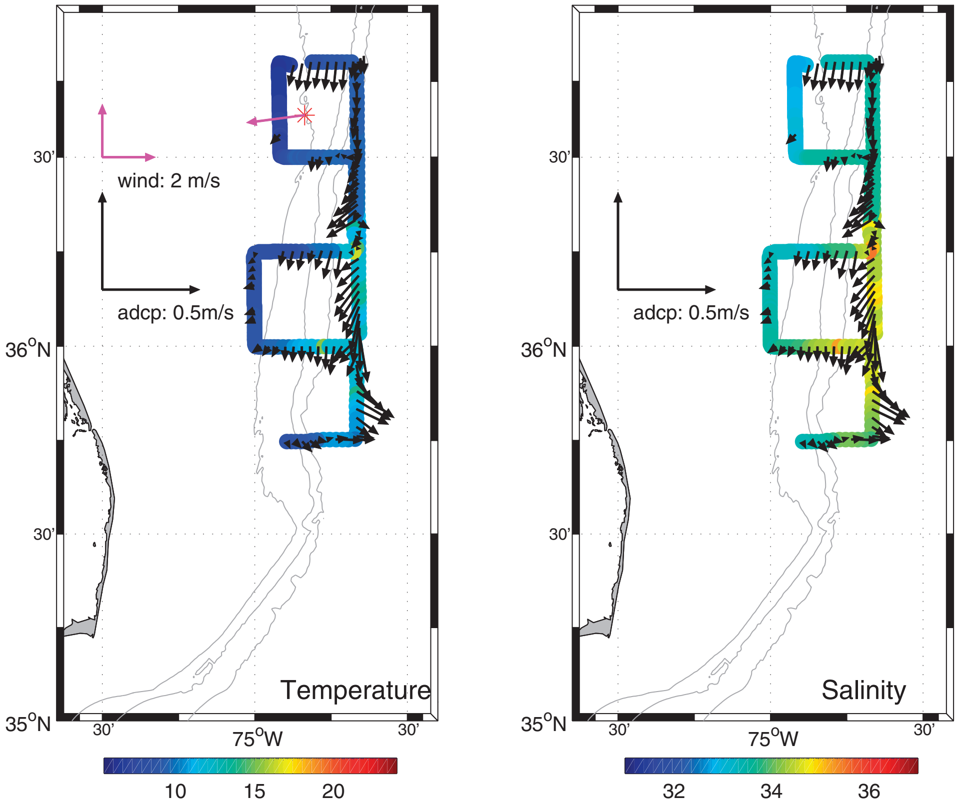

Figure 6. A plan view of (left panel) temperature and (right panel) salinity at 25-m depth from a survey conducted on January 29–30, 2005. Velocity vectors measured by the shipboard ADCP are also plotted, and are 4-m averages centered at 25-m depth. Winds averaged over the time period of the survey appear as the magenta arrow at 36°38′N in the left panel.

velocities from both January 23 (returning from Morehead City) and February 3 (recovering moorings at 35°15′N) indicated southward flow across isobaths between 35°00′N and 35°15′N, and cold Middle Atlantic Bight shelf water extending southward to 35°15′N.

4. Shelfbreak frontal structure north of Cape Hatteras

The first survey (January 19–20) found alongshelf gradients in the temperature, salinity, and density fields where the shelfbreak front terminated north of Cape Hatteras. During this survey the mean winds were from the west-northwest at 4 m s^{-1}. The cross-shelf salinity and density gradients at 35°45′N were confined to a 10-km wide frontal zone that extended upwards and offshore from the 50-m isobath (Fig. 7, upper left panel). The shelf water was well-mixed. The shelf and slope flows were both northward (Fig. 7, lower left panel) except for a narrow (5 km) and relatively shallow (25-m deep) portion of the baroclinic jet

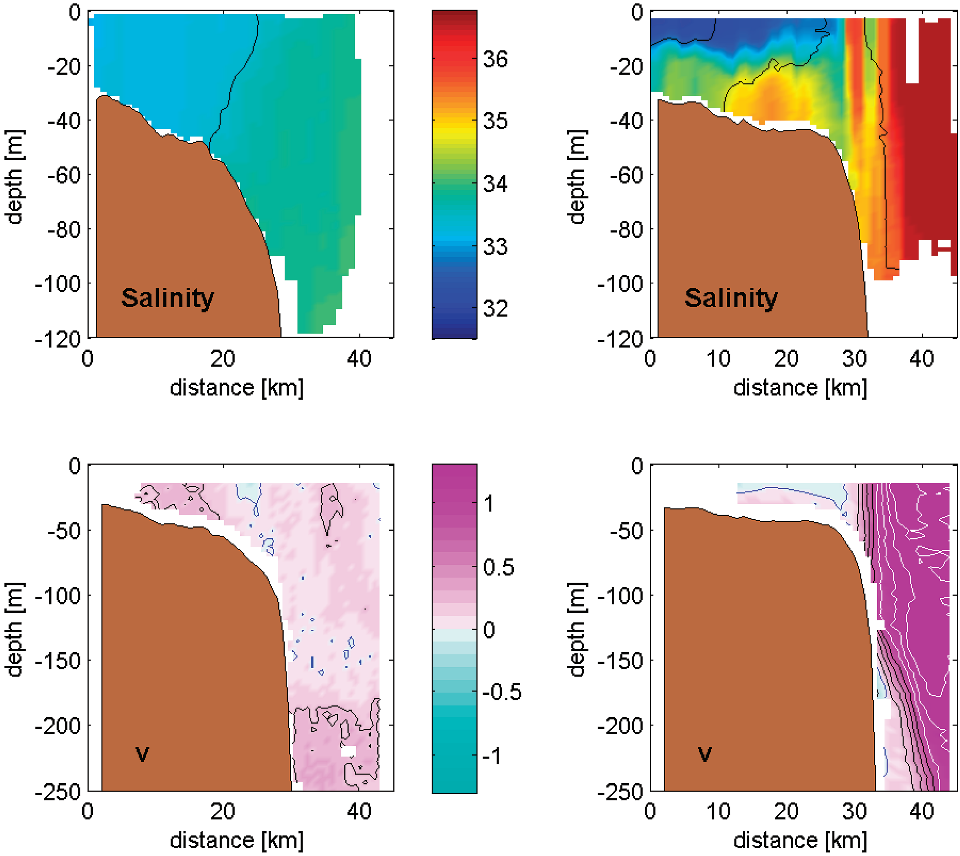

Figure 7. Two cross-shelf transects from January 19–20, 2005. The upper left panel shows the salinity field along 35°45′N. The black contour line is the 25.75 isopycnal. The lower left panel shows the southward velocity from the same section. Note that the flow is all northward except for a narrow surface-trapped band centered at x = 25 km. The upper right panel shows the salinity field along 35°30′N, and the lower right panel shows the northward velocity from the same section. The contours in the upper right panel denote the 25.0 and 25.25 isopycnals. Note the Gulf Stream directly abutting the shelfbreak in this section. The vertical scale is different between the upper and lower panels.

associated with the shelfbreak front. Farther south, at 35°30′N, there was a different set of water masses, typically associated with the Gulf Stream and South Atlantic Bight (Fig. 7, upper right panel). Over the slope, the Gulf Stream directly abutted the shelfbreak (Fig. 7, lower right panel), with extreme lateral shears (relative vorticity comparable to f, the Coriolis parameter). The maximum northward velocity was 120 cm/s. Over the shelf, a much fresher (less than 32.5) buoyant layer (which may have originated as the Chesapeake Bay plume) flowed southward over the upper 30 m of the water column. This buoyant layer was both fresher and colder than the water masses of the outer shelf.

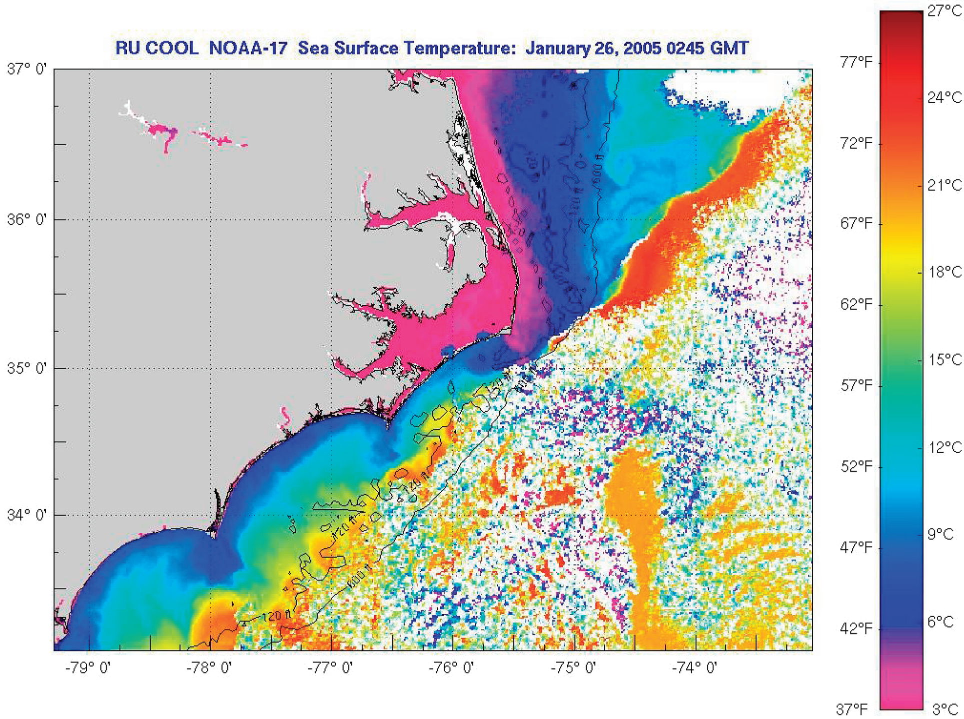

Figure 8. A surface thermal image of the Cape Hatteras region from the archive at Rutgers University. This image is from January 26. Note the band of cold water adjacent to the shore as well as the sharp thermal gradients where the Middle Atlantic Bight shelf water abuts the Gulf Stream.

The buoyant surface plume had a temperature of 6°C and a salinity of 30.5–32. Note that the layer is only 10-m thick. In contrast, the Middle Atlantic Bight shelf water just shoreward of the shelfbreak front had temperatures between 8–10°C and salinities in the range of 33–34. Satellite thermal imagery was limited during the experiment, but the few clear images indicate a cold band adjacent to the coast extending from Chesapeake Bay down to Diamond Shoals (Fig. 8). Furthermore, we did not observe a mid-shelf front as described by Ullman and Cornillon (1999). This front could have served as a possible upstream source of the Hatteras Front, particularly as it has been hypothesized to form due to cooling differences in different depth ranges, but no evidence of its presence was observed in any of the Scanfish transects.

During the second survey, on January 29–30 (Fig. 9), the shelf flow as well as the shelfbreak jet were consistently southward, with onshore or offshore flows due to the shelfbreak eddies (Fig. 6). The winds were from the north, decreasing from 8 m/s to virtually zero over the two days of the survey. At this time, the Gulf Stream was farther offshore of the shelfbreak. The shelfbreak front was present in all five cross-shelf transects

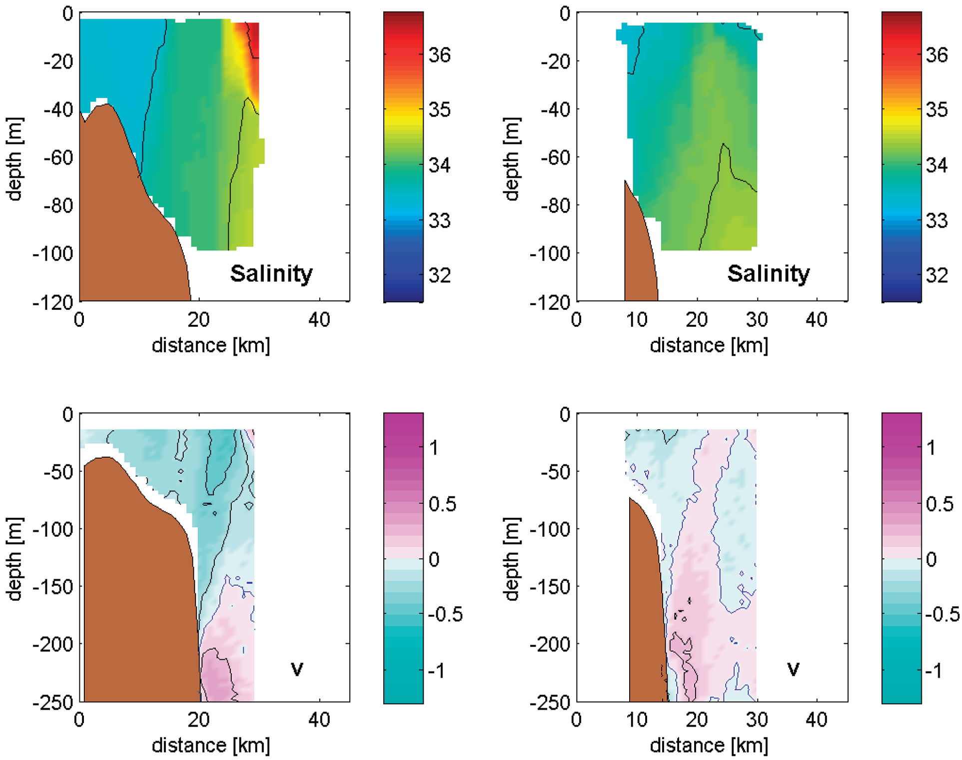

Figure 9. Two across-shelf transects from January 29–30, 2005. The salinity field along 36°15′N appears in the upper left panel, and the northward velocity appears in the lower left panel. The salinity field along 35°45′N appears in the upper right panel and the northward velocity from this section appears in the lower right panel. Note that the vertical scale is different between the upper and lower panel. The two dark contours in the upper panels are the 26.0 and 26.25 isopycnals respectively. The contour interval for the lower panels is 0.20 m s^{-1}.

in this survey, extending from 36°45′N to 35°45′N. Maximum southward velocities seen in the jet at 25-m depth in the five transects ranged from 0.49 m s^{-1} at 36°15′N to 0.22 m s^{-1} at 35°45′N. However, the maximum jet velocities varied substantially from section to section and did not monotonically decrease with proximity to Cape Hatteras. The maximum jet velocity at 36°45′N was 0.46 m s^{-1}, while the maximum at 36°30′N was 0.25 m s^{-1} and at 36°00′N was 0.47 m s^{-1}.

A particularly striking characteristic of the velocity fields of this survey were the high relative vorticities associated with the jet in all five transects. The northward velocity was smoothed with a second-order Butterworth filter with a horizontal scale of 5 km before calculating the relative vorticity. Figure 10 shows the cross-shelf distribution of the northward velocity and the relative vorticity ($\partial v/\partial x$ only—eastward shear of the northward flow) from the transects along 36°45′N, 36°15′N, and 36°00′N. Relative vorticity

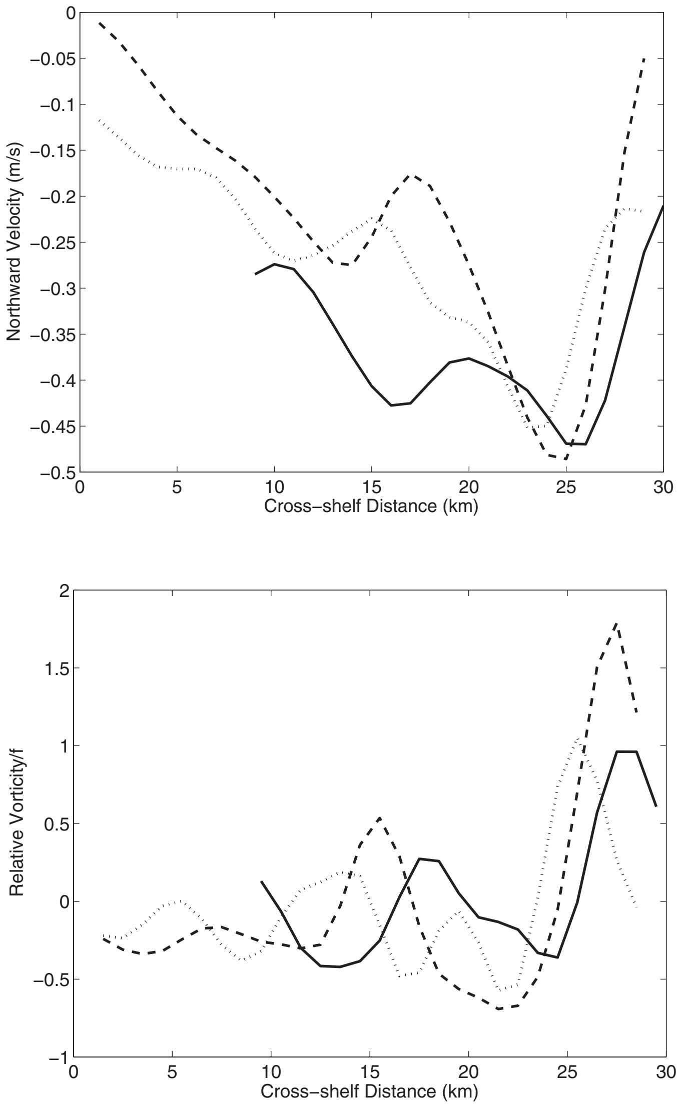

Figure 10. The southward flow at 25-m depth from three separate cross-shelf transects on January 29–30 appears in the upper panel. The solid line is from 36°45′N, the dashed line is from 36°15′N, and the dotted line is from 36°00′N. The relative vorticity (∂v/∂x) normalized by the Coriolis parameter appears in the lower panel.

(expressed relative to the local Coriolis parameter, f, of $8.365 \times 10^{-5}\ \mathrm{s}^{-1}$) was as high as 1.8f along the 36°15′N transect, and reached maxima of roughly 1.0f in two other sections. These relative vorticities are much higher than those derived from previous observations of the shelfbreak front. For example, Linder and Gawarkiewicz (1998) found a maximum

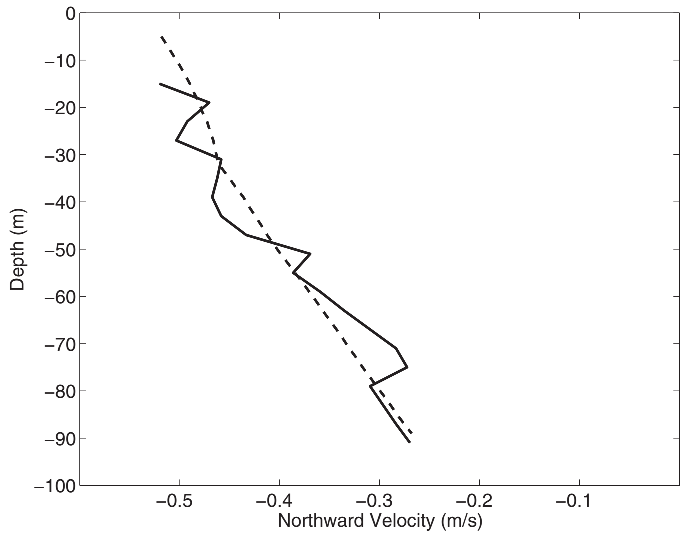

Figure 11. The vertical structure of the northward velocity down the core of the shelfbreak jet along 36°15′N from January 29–30. The solid line is the velocity measured by the shipboard ADCP and the dashed line is the geostrophic velocity. The geostrophic velocity was referenced using the average ADCP velocity between 15- and 90-m depth, which was 0.38 m s^{-1}.

relative vorticity of 0.2f in the winter-time climatological jet off New Jersey. In a synoptic study, Gawarkiewicz *et al.* (2004) found relative vorticities as high as 0.6f in a large amplitude frontal meander south of New England during the summer. In all the transects of January 29–30, the maximum shear was cyclonic and located at the offshore edge of the shelfbreak jet. In contrast, the maximum magnitude of the shear observed during January 19–20 survey was −0.5f (anticyclonic).

The southward flow in the shelfbreak jet was in a geostrophic balance in some sections. An example is the section along 36°15′N (Fig. 11). To compute the thermal wind shear for this section, adjacent density profiles 2.0 km apart (at 36°15′10′N, 74°43.31′W and at 36°15.15′N, 74°44.66′W) were used to calculate the cross-shelf density gradient. For this transect, which has the strongest jet flow at 25-m depth, the geostrophic velocities match the velocities measured by the shipboard ADCP fairly well. The average vertical shear ($\partial v/\partial z$) from the geostrophic calculations between 15- and 89-m depth was 0.0035 s^{-1}, while the shear measured by the shipboard ADCP over the same depth range was 0.0033 s^{-1}. Similar comparisons, using pairs of density profiles in the core of the jet to estimate the geostrophic shears, were done for the other four sections from January 29–30. In the section along 36°30′N, the geostrophic shears are also close to the measured shears from the ADCP between depths of 15 to 40 m, the deepest depth for the ADCP data in this profile, with a difference in shear of 0.00015 s^{-1}. However the geostrophic shear did not compare well with the ADCP shears for the other three sections. The differences between the northward geostrophic and ADCP shears between 15- and 90-m depth were 0.001 s^{-1},

$0.002\ s^{-1}$, and $0.0008\ s^{-1}$ for the jet cores along 36°45′N, 36°00′N, and 35°45′N, respectively.

Surprisingly, the shear of the southward velocity measured by the ADCP extended to depths which were significantly deeper than the shelfbreak. The depth at which the southward jet velocity declined to zero varied significantly from section to section. Along 36°15′N, the depth of zero southward velocity beneath the core of the jet was at 139 m, while at 36°00′N and 35°45′N the zero crossings were at 219 m and 71 m, respectively. This variability is likely due to the presence of large eddies offshore of the shelfbreak jet. As discussed in Churchill and Gawarkiewicz (2009), the vertical scale for eddies over the slope was as much as 300 m. However, the density differences across the front also vary between sections and contribute to the differences in the vertical structure of the jet velocity.

We can now estimate some of the important scales and non-dimensional parameters for the jet. Along the 36°15′N transect, the vertical stratification over the depth range of 5 to 80 m down the core of the jet is $N = 0.0033\ s^{-1}$. The cross-frontal density difference at 25-m depth is, on average, $0.203\ kg\ m^{-3}$. The baroclinic Rossby radius, defined by

$$r = \sqrt{(g'H)}/f$$

is 5.0 km. Here, $H = 90$ m (representing the mean depth of the foot of the shelfbreak front), $f = 8.37 \times 10^{-5}\ s^{-1}$, and the reduced gravity $g' = g(\Delta\rho/\rho_o)$, where $\Delta\rho = 0.203\ kg\ m^{-3}$ and $\rho_o = 1025\ kg\ m^{-3}$. The scaled alongshelf velocity is

$$v = \sqrt{g'H}.$$

Using the same values of g' above and $H = 90$ m gives $v = 0.42\ m\ s^{-1}$, which is similar to the shipboard ADCP value at 25-m depth of $0.49\ m\ s^{-1}$ at the maximum of the jet.

Two other important nondimensional numbers are the Rossby number, $(\partial v/\partial x)/f$, which is 1.8, and the Burger number. Assuming a frontal width L of 10 km, the Burger number is

$$Bu = (NH/fL)^2$$

and is 0.13 for the parameters of the jet along 36°15′N. As indicated earlier, the values of the Rossby number are among the highest ever measured in the shelfbreak front. The implications of these values on the potential for frontal instabilities will be examined in the Discussion.

For the other four transects, the density difference across the front at 25-m depth varies from a minimum of $0.179\ kg\ m^{-3}$ along 36°30′N to a maximum of $0.271\ kg\ m^{-3}$ at 36°00′N. The former corresponds to the weakest jet velocity at 25 m and the latter corresponds to the strongest.

5. Freshwater transports north of Cape Hatteras

During the course of the experiment, there was only one time period in which we were able to resolve the separation of the shelfbreak frontal jet simultaneously with the southward flow over the continental shelf, on February 1. As shown in Figure 5, Middle Atlantic Bight Cold Pool shelf water (salinity between 33 and 34) was not observed to pass into the South Atlantic Bight at any time, although buoyant plume water was capable of penetrating southward into the South Atlantic Bight. This was also confirmed by observations by D. Savidge and J. Austin concurrently south of Diamond Shoals. Thus, we conclude that the southward shelf flow of Cold Pool water masses over the southern transect can be used to infer the offshore transport of freshwater into the Gulf Stream. Adding the freshwater transport carried offshore in the shelfbreak frontal jet gives the total freshwater transport from the Middle Atlantic Bight shelf/slope into the Gulf Stream. The buoyant plume water and its contribution to freshwater fluxes into the South Atlantic Bight will be the focus of future work and will not be considered here.

On February 1, two cross-shelf transects were sampled at 35°45′N and 35°35′N. A north-south transect was sampled between these lines along 75°40′W. During this time (Figs. 3 and 4), the winds were from the north and over 10 m s^{-1} in magnitude. A plan view of the velocity vectors at 25-m depth from this box appear in Figure 12. In the north-south section, the baroclinic portion of the shelfbreak jet detached from the continental shelf and was directed due east into the Gulf Stream. As will be seen, shelf flows shoreward of the baroclinic jet were still strongly southward.

A vertical section showing the eastward velocity structure of the jet is shown in Figure 13. The maximum eastward velocity was 0.35 m s^{-1}, and was surface-trapped. The mean eastward velocity averaged between 15- and 100-m depth was 0.17 m s^{-1}. The vertical scale of the jet was 70 m. The jet width was 10 km.

All of these scales are consistent scales of the southward-flowing jet seen in the southernmost section of the January 29–30 survey. The eastward transport over the top 100 m in the section was 0.27×10^6 m^3 s^{-1} (Sverdrups). This is comparable to the annual climatological mean value of the buoyancy driven flow in the shelfbreak jet, which is 0.24 Sv south of Nantucket Shoals and 0.16 Sv off New Jersey (Linder and Gawarkiewicz, 1998). Similarly, a climatological transport estimate for the buoyancy-driven shelfbreak jet using only synoptic sections for the entire Middle Atlantic Bight (Fratantoni and Pickart, 2007) is 0.26 Sv. Thus the observed synoptic eastward transport is remarkably similar to previous climatological values for the alongshelf transport of the shelfbreak jet.

Because of the rough weather, the Scanfish was not deployed within this box. However, we used the average vertical structure of the salinity field from the southernmost cross-shelf section on January 30 (along 35°45′N) to calculate the freshwater transport moving eastward in the detached shelfbreak jet. The transport is calculated as

$$F_s = \int u(z)((s_o - s(z))/s_o)L_s \, dz$$

where F_s is the eastward freshwater transport, u(z) is the horizontally averaged (over 18 km) eastward velocity between depths of 15 to 89 m for the section in Figure 13, and

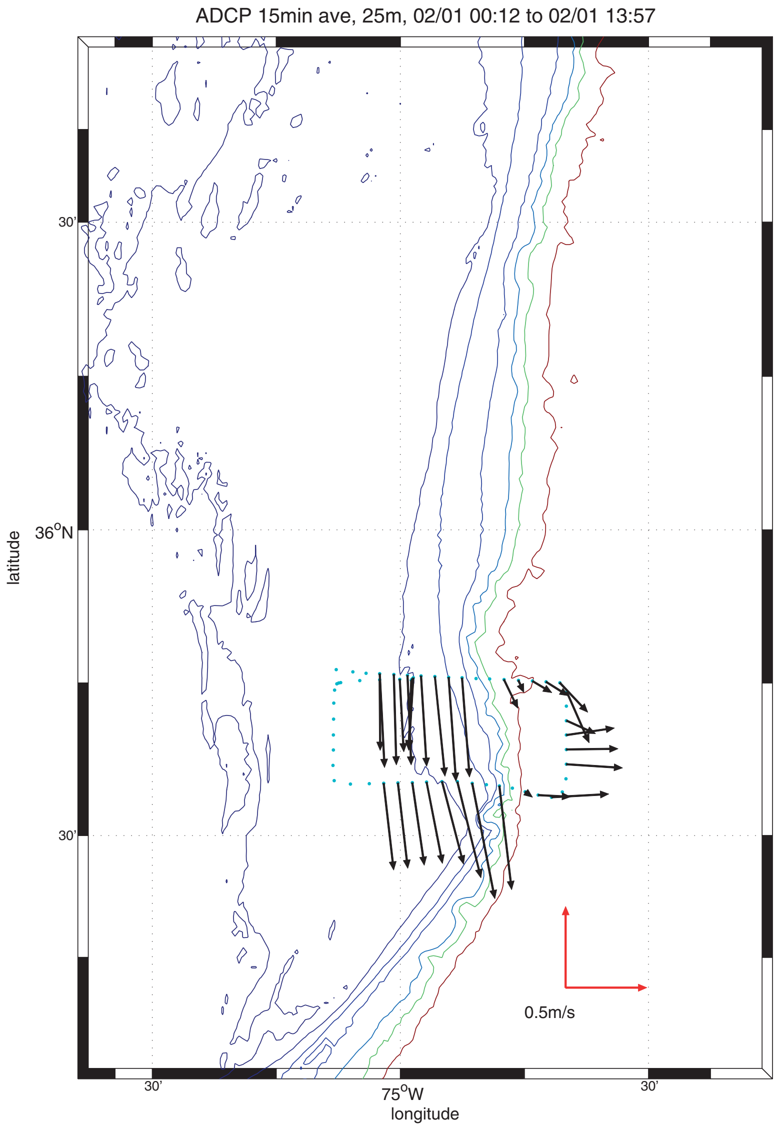

Figure 12. A plan view of velocity at 25-m depth measured on February 1, 2005.

s(z) is the horizontally averaged salinity as a function of depth for the southernmost section from January 30 along 35°45′N. L_s is the length of the section. The reference salinity s_o is taken to be 35.0. The total eastward freshwater transport is 7.4 mSv (1 mSv $= 10^3$ m^3 s^{-1}) associated with the detachment of the shelfbreak jet. Shifting the reference salinity to 35.1 results in a ten per cent increase to 8.1 mSv and shifting to 34.9 results in a ten per cent decrease to 6.6 mSv.

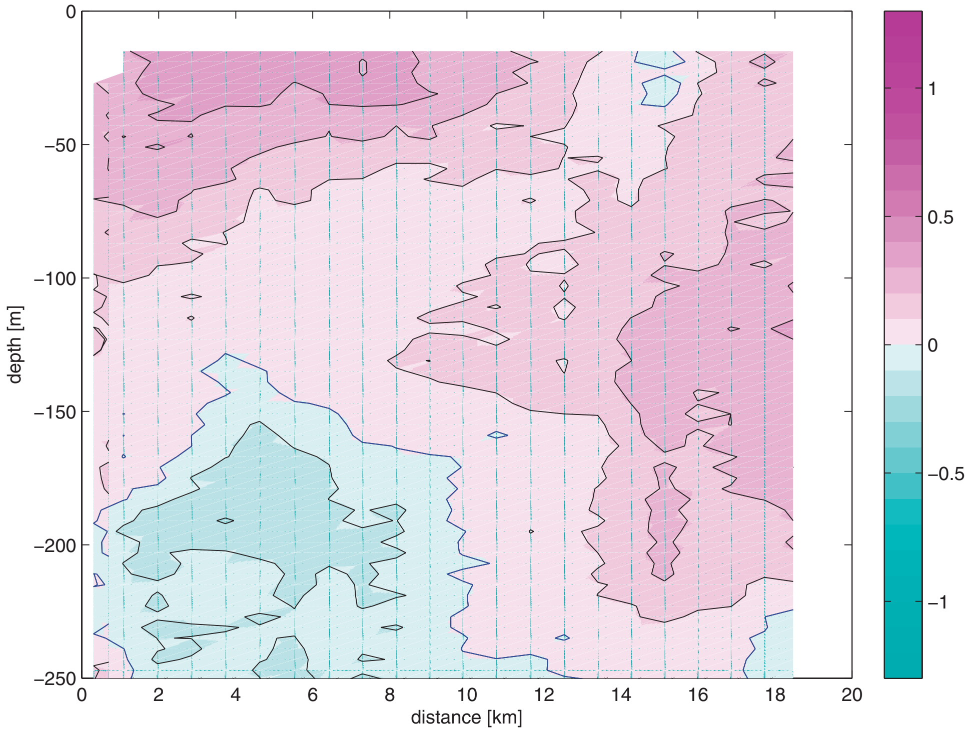

Figure 13. Eastward velocity measured along a north-south section on February 1, 2005, between 35°45′N and 35°35′N. North is to the right in this figure. The shelfbreak frontal jet is concentrated in the upper 100 m of the water column between x = 0 and x = 14 km.

The southward velocity from the southern end of the box on February 1 appears in Figure 14. The flow reached a maximum of 0.76 m s^{-1}. The lateral shear was a maximum of 0.38 m s^{-1} over 2 km in the cross-shelf direction. This gives a relative vorticity of 2.3f, comparable to the large values seen farther north on January 29–30. The total southward transport in this section is 0.61 Sv, and the freshwater transport to the south associated with the shelf flow is 19.9 mSv.

Summing the eastward and southward transports, we obtain a total of 0.88 Sv of flow departing the Middle Atlantic Bight shelf. The total freshwater transport is 27.3 mSv. As discussed below, these are large values relative to previous estimates.

6. Discussion

We will briefly discuss the implications of these results in the context of previously published work on the flow near Cape Hatteras and on shelfbreak processes farther north in the Middle Atlantic Bight. We will discuss the freshwater transports, the jet structure and

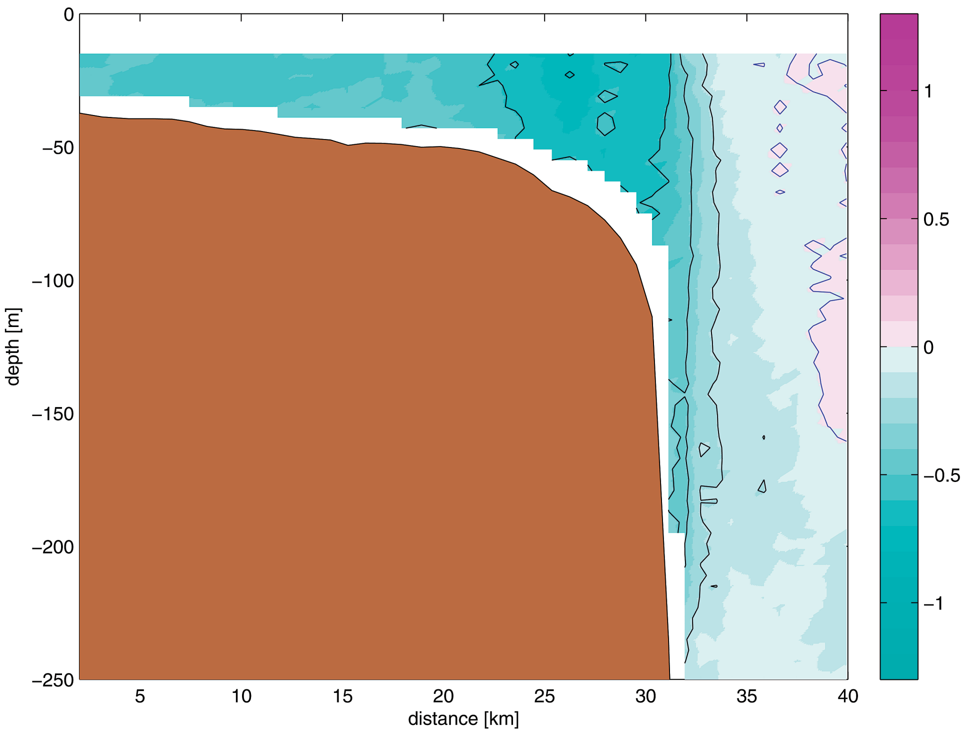

Figure 14. A section of the northward velocity along 35°35′N on February 1, 2005. The maximum southward velocity near the shelfbreak is 0.76 m s^{-1} at x = 28 km. The contour interval is 0.20 m s^{-1}.

implications for frontal instability in this region, and finally the ultimate fate of the Middle Atlantic Bight shelf water, namely the formation of the "Ford water" in the Gulf Stream.

The freshwater transport measured on February 1 is very high relative to previous estimates. Bignami and Hopkins (2002) used data from a mooring array with an one-year duration and concluded that the long-term average freshwater transport is 5 mSv. Similarly, Pietrafessa *et al.* (1994) estimated a transport of 3 mSv, a value used by Loder *et al.* (1998) in their summary of freshwater transports in the Middle Atlantic Bight. For our observations, just the shelfbreak frontal detachment contributes an additional 50% to the Bignami and Hopkins (2002) estimate and is more than double the Pietrafesa *et al.* (1994) estimate. However, the added southward freshwater transport from February 1 is 19.9 mSv, making the total freshwater transport (27.3 mSv) five times the Bignami and Hopkins (2002) estimate for a long-term mean freshwater transport. We consider the southward freshwater transport to be a reasonable upper bound due to the strong southward wind stress at the time (0.3 Pa). The southward freshwater transport was much smaller when the wind forcing was weak. Using the same method from the previous section and the velocity data from the southernmost cross-shelf transect on January 29–30 (when the wind speed

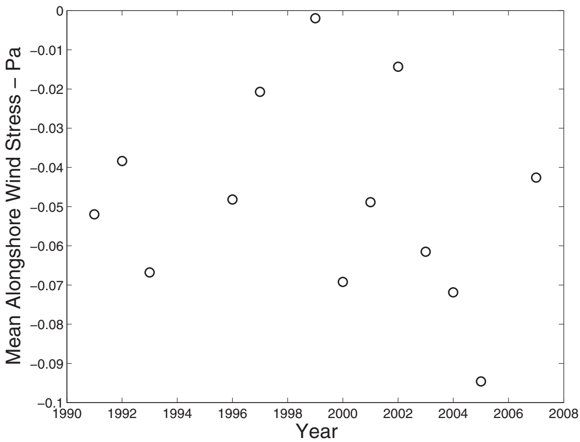

Figure 15. Mean alongshelf (northward) wind stress for the month of January for the period 1991–2007. The winds are taken from National Data Buoy Center buoy 44014.

was very weak), gives a freshwater transport of 1.6 mSv. Thus the total transport of 9.0 mSv (southward over the shelf plus detached shelfbreak jet) when the wind stress is weak is still nearly double the Bignami and Hopkins (2002) estimate for the long-term mean.

It is clear that wind-driven flow is an important factor in high freshwater transport observed during our study, as the velocity data used in determining the transport are from a time of strong and persistent southward winds and when the maximum southward flow appears in our 80-m mooring record (Fig. 4). It is important to note that the time of our study, January 2005, is a period of unusually strong southward winds in the southern Middle Atlantic Bight. Our analysis of the wind data from NDBC buoy 44014 show that a mean southward alongshelf wind stress is the norm in the southern Middle Atlantic Bight during January (Fig. 15). However, the mean alongshelf wind stress of January 2005 is the strongest of all the mean January alongshelf wind stresses computed from the 40014 data record, which begins in 1991, and exceeds the overall mean January alongshelf wind stress by a factor of 2.

The value for the total transport of shelf water is large (0.88 Sv) when compared to recent estimates for the shelf flow without the shelfbreak jet. For example Lentz (2008) uses long-term average velocities from moored current meters and a simple model of alongshelf flow and obtains a value of 0.09 Sv at 36°14.7′N. However, this only extends to the 45-m isobath. Synoptic transport estimates which include the shelf and shelfbreak jet report transports as large as 1.0 Sv (Rasmussen *et al.,* 2005).

There are two major implications of these results that should be considered in deriving future estimates of freshwater transport (both offshore and alongshelf) in the Middle Atlantic Bight. The first is that the shelfbreak jet carries a considerable amount of freshwater and must be considered when addressing the freshwater balance of the Middle

Atlantic Bight shelf and slope. Because of the small across-shelf scale of the jet, as well as its extreme variability, it is necessary to resolve motions on the horizontal scale of the baroclinic Rossby radius, which for this case is 5 km, as well as the temporal variations, which may be on the time scale of a day. This argues strongly for the use of Autonomous Underwater Vehicles for future transport measurements, as they can resolve jet motions on the necessary spatial scales and also sample more frequently. Second, close comparisons between numerical models of the Middle Atlantic Bight shelf and slope with high-resolution observations such as these will be necessary to determine the time-dependent nature of the freshwater balance over the shelf. This is particularly important as coastal oceanographers address the impacts of global change on shelf circulation and shelf ecosystems. For the shelf north of Cape Hatteras, two key elements of model-data comparison would be the joint impact of wind forcing and the Gulf Stream influence, and the time-dependence of the buoyancy-driven flow within the shelfbreak jet. For the former, determination of the impact of Gulf Stream variability near Cape Hattteras, such as due to meandering (Savidge, 2004), on the northward flows over the shelf north of Cape Hatteras particularly merits attention.

The shelfbreak frontal structure observed in this study contains the highest Rossby number measurements for the shelfbreak front reported in the literature. This has important implications for the stability characteristics of the front. Lozier *et al.* (2002) describe the stability characteristics of the shelfbreak front in the Middle Atlantic Bight. They consider a case with a maximum jet velocity of 0.6 m s^{-1}, and a Rossby number of 0.96. (This is the highest value of the Rossby number treated in their study). They find growth rates shorter than a day over a wide range of wavelengths. Thus the present study reinforces the point made in Churchill and Gawarkiewicz (2009) that the shelfbreak frontal instabilities are a potential cause for some of the eddies observed over the continental slope. For our observations, the typical cross-shelf scale of the shelfbreak jet, 10 km, is double the value of the baroclinic Rossby radius. Lozier *et al.* (2002) note previous studies suggesting that this generally leads to baroclinic instability being the primary contributor to instability relative to barotropic instability. Further work is needed to extend the analysis of frontal instability of the shelfbreak front to higher Rossby numbers.

Finally, an obvious question that we have implied but not directly addressed in previous sections is, what is the ultimate fate of the Middle Atlantic Bight shelf water? As mentioned in the Introduction, Middle Atlantic Bight shelf water has been observed as the "Ford water" in the Gulf Stream. However, this study provides direct evidence of the entrainment of the Middle Atlantic Bight shelf water into the Gulf Stream. Figure 16 shows a cross-shelf temperature and salinity section from 35°45′N on January 27. Note the continuity of the shelf water with salinities lower than 34.0 extending from the continental shelf and subducting into the Gulf Stream. Further work is necessary to examine the dynamics of this subduction and entrainment into the Gulf Stream, which was seen in other sections of this study.

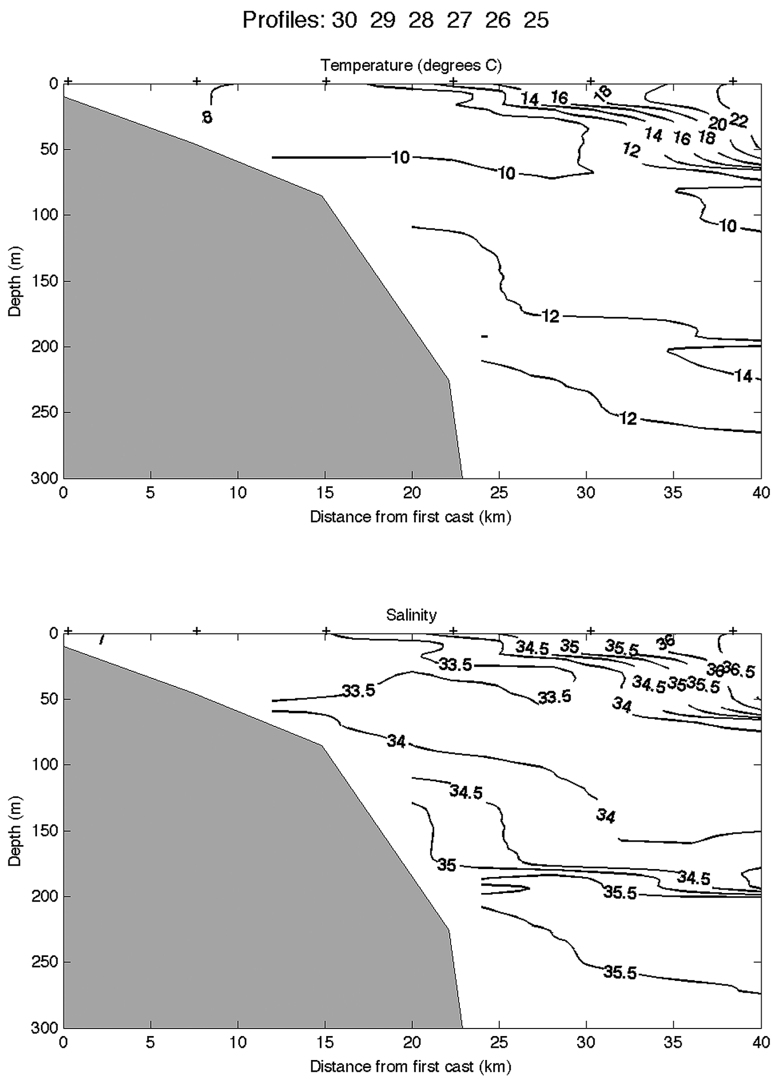

Figure 16. A cross-shelf transect from January 27, 2005, showing the offshore transport of Middle Atlantic Bight shelf water and entrainment into the Gulf Stream. Note the continuity of the shelf water (<34.0 salinity) between the continental shelf and the Gulf Stream (offshore edge of the transect).

7. Conclusions

A high-resolution synoptic hydrographic survey north of Cape Hatteras in January/ February, 2005 shows evidence of extreme variability over the shelf. Sections taken on January 19–20, while the Gulf Stream directly abutted the shelf, indicate northward flow over the shelf significant distances north of Cape Hatteras. However, sections taken ten days later, after the Gulf Stream had retreated, show that a strong southward flow was established over the shelf. The maximum jet velocity in the shelfbreak front was nearly 0.5 m s^{-1} at 25-m depth. The jet was characterized by extremely high Rossby numbers (up to 1.8) and a relatively small Burger number (0.1).

Freshwater transport associated with an eastward flowing shelfbreak front as it detached from the continental shelf was 7.4 mSv, while concurrent estimates of southward freshwater transport was 19.9 mSv. This is substantially larger than previous estimates from mooring arrays. Further work is necessary to quantify the freshwater transports as they are affected by both Gulf Stream forcing of the shelf flow north of Cape Hatteras as well as the wind-driven flow over the continental shelf.

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