4
The Gulf Stream System

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4.1 Introduction

In the two decades since the first publication of Stommel's (1965) monograph on the Gulf Stream, our knowledge of the Gulf Stream System has been expanded dramatically through the development and application of new, powerful measuring techniques. Multiple ship surveys of the type organized by Fuglister (1963) provided the first systematic descriptions of the spatial structure between Cape Hatteras and the Grand Banks that included the surrounding slope waters to the north and the Sargasso Sea waters to the south of the Gulf Stream. Several major theoretical and interpretative studies grew from the base of data and descriptions provided by this study. During the same period, instrumented buoys, both moored and drifting, were beginning to reveal some of the complexities of the subsurface and deep fields of temperature and currents. Among the new techniques implemented in the 1960s was infrared-radiation imaging to map the thermal patterns of the ocean surface from satellites orbiting the earth (Legeckis, 1978). The two-dimensional surface thermal maps that have been obtained have added rich detail to our knowledge of the strongly varying thermal structure associated with the Gulf Stream throughout its path. Yet, despite these advances in our ability to measure, our understanding of the dynamic mechanisms by which the Gulf Stream forms, develops in intensity, decays, and finally merges into the large-scale circulation of the North Atlantic have not evolved as satisfactorily. Even the mechanism controlling the position of the Gulf Stream after leaving the continental shelf at Cape Hatteras has not yet been firmly established. Is the Gulf Stream controlled by bottom topography, by the distribution of mean wind stress, or by a mechanism yet to be determined? The dynamics by which meanders of the Gulf Stream amplify and develop into large rings and eddies and the subsequent movement and evolution of these entities are not well understood. In spite of the impressive progress of the past decades, much remains to be done to resolve and understand the particular mechanisms that determine the character and behavior of the Gulf Stream along its entire path from the Gulf of Mexico into the central North Atlantic.

In preparing material for this review, I concluded that my initial plans for a comprehensive discussion of the literature since 1958 were unrealistic. As over 200 references plus numerous technical reports and articles were identified, it became obvious that only a few aspects of the Gulf Stream System could be covered in a single short review. Given the necessity for choice, it is clear that the selection must reflect my preferences, interests, and perhaps, biases. I hope my effort to trace particular lines of research in the literature

will prove of interest to readers and will serve as a guide to a part of the rapidly growing body of literature that represents our collective knowledge of the Gulf Stream System. That other, equally important, aspects of research are omitted is unfortunate but inevitable.

4.2 The Gulf Stream System

The subdivisions of the *Gulf Stream System* proposed by Iselin (1936), reproduced in figure 4.1, although not entirely accepted in practice, serve as a convenient framework for grouping the research literature. Starting from the Gulf of Mexico, the *Florida Current* was labeled by Iselin as the portion of the Gulf Stream System flowing through the Florida Straits northward past Cape Hatteras to the point where the flow leaves the continental slope. Objections had been raised by Nielson (1925) and Wüst (1924) to using the word "Gulf" in reference to the Florida Current, as they considered that the water flowed directly across from the Yucatan Channel into the Florida Straits rather than from the Gulf of Mexico. This distinction seems less justified now because the flow through the Yucatan Channel has been observed to loop well into the Gulf of Mexico on occasion (Leipper, 1970; Behringer, Molinari, and Festa, 1977), although the Florida Current does not originate there. After leaving the Florida Straits, the Florida Current presses close to the continental slope and in the upper layers forms a relatively continuous system. The flow is augmented on the seaward side by inflow of water of essentially the same characteristics

as the Florida Current. Iselin included both sources under the same label. Oceanographers frequently refer to the Florida Current between the Florida Straits and Cape Hatteras as the Gulf Stream. However, because measurements and theoretical studies have tended to relate this portion of the Gulf Stream System more closely to the current in the Straits rather than to the currents downstream from Cape Hatteras, Iselin's nomenclature is more convenient in the present review.

North of Cape Hatteras, the current begins to flow seaward off the slope into deeper water. Freed of the constraints of the shelf, the Gulf Stream develops meanders of increasing amplitude downstream. Bowing to popular usage, Iselin retained the name *Gulf Stream* for the section between Cape Hatteras and the Grand Banks. The name *North Atlantic Current* had already been widely accepted for easterly flows at mid-latitudes beyond the Grand Banks. Even though an extension of the Gulf Stream, the North Atlantic Current, according to Iselin, becomes separated into branches and eddies to form a distinctly different regime of flow. Its eastern limit is not clearly defined, though Iselin assumed that the branches extended into the eastern North Atlantic. A composite view of the western portion of the Gulf Stream System (figure 4.2) has been assembled by Maul, deWitt, Yanaway, and Baig (1978) from infrared satellite images and surface tracking by ships and aircraft. The sharp thermal contrasts between the warm currents and the neighboring waters are detectable from space and reveal the variability of the Gulf Stream System throughout its length. The complexities introduced by the near-surface spatial and temporal variability of the Gulf Stream System are only beginning to be described. Comparatively little is known of the variable subsurface and deep structure, particularly downstream of Cape Hatteras.

4.3 The Florida Current

Iselin (1936) defined that Florida Current as all the northward-moving waters with velocities exceeding 10 cm s^{-1} starting along a line south of Tortugas and extending to the point past Cape Hatteras where the current ceases to follow the continental shelf. The three chief characteristics of the Florida Current noted by Iselin are that it greatly increases in volume as it flows north, that it flows most swiftly along the continental slope, and that over most of its length it is relatively shallow, transporting water no colder than 6.5°C until passing the northern limit of the Blake Plateau. The surface thermal structure of the Florida Current between Miami and Cape Hatteras can be seen in the infrared image reproduced in figure 4.3.
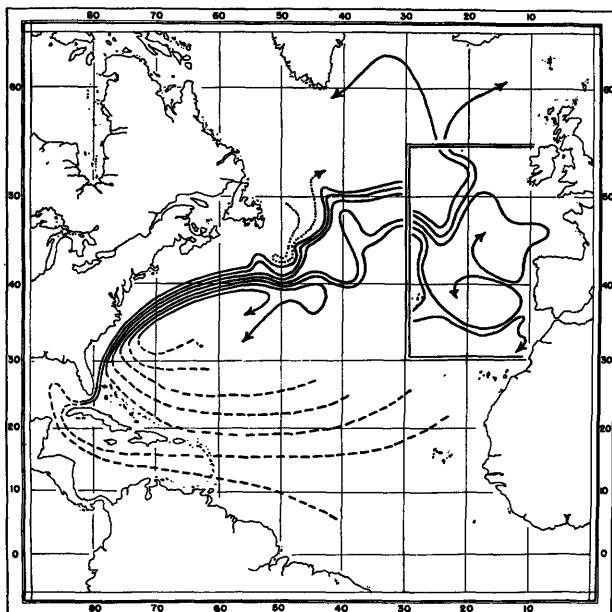

Figure 1

Figure 4.1 A schematic diagram showing the Gulf Stream System as described by Iselin (1936). Each streamline represents a transport interval of about $12 \times 10^6 \text{ m}^3\text{s}^{-1}$.
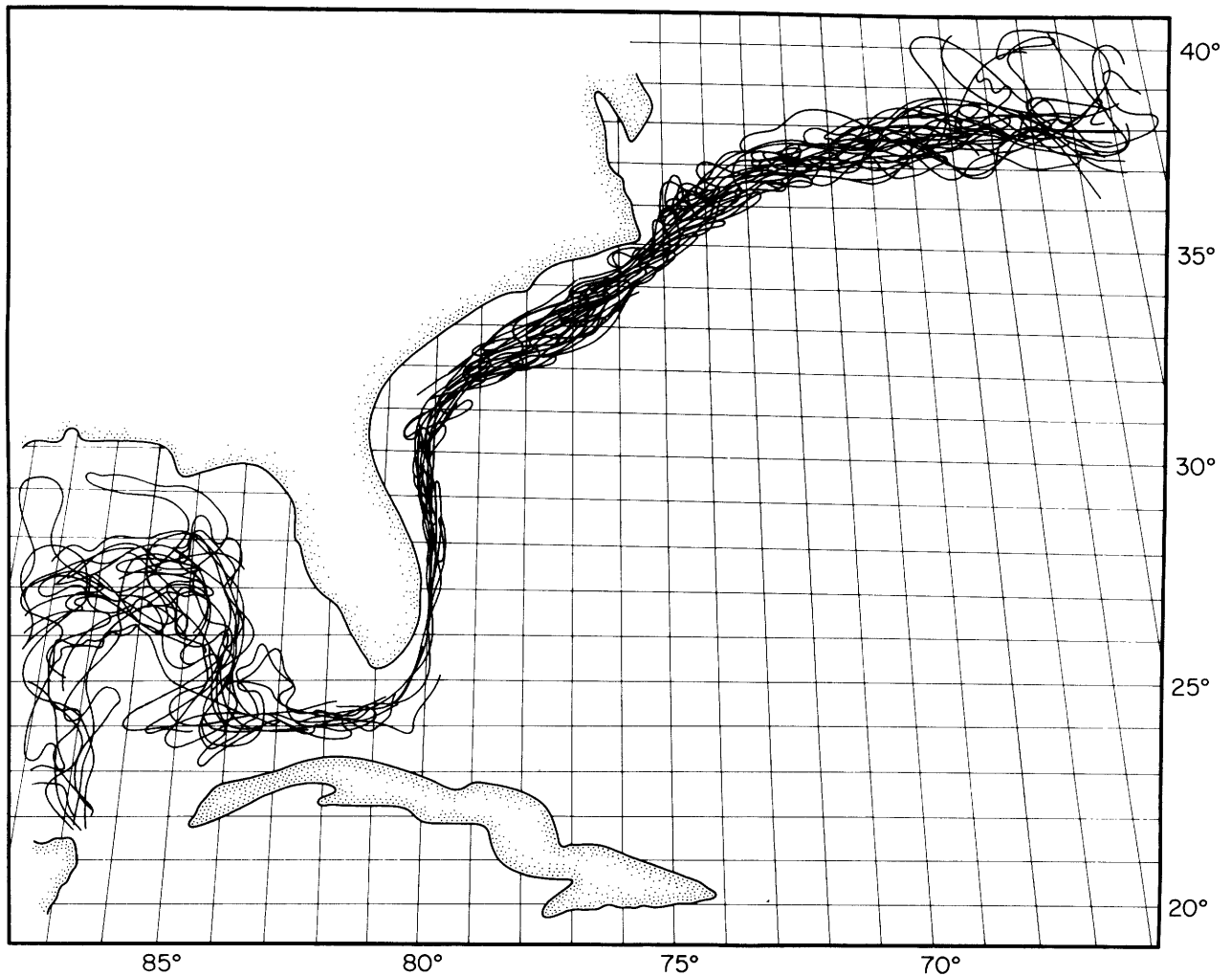
Figure 4.2 A composite of thermal fronts of the Gulf Stream System showing variability for a 9-month period between February and November 1976. [Courtesy of G. Maul from Geostationary Operational Environmental Satellite (GOES) infrared observations.]

4.3.1 Sea-Level Slope from Tide Gauges

Montgomery (1938b) assumed that the intensification of the Florida Current as it flowed into the Straits of Florida is produced by a hydraulic, or pressure, head between the Straits and the Gulf of Mexico. Although the differences in sea level corresponding to the pressure head could not be measured directly, tide-gauge measurements could show variations in the slope and hence, in the Florida Current itself. The recent development of precision altimetry from satellites capable of resolving the shape of the sea surface has renewed interest in the possibility of monitoring major ocean currents remotely. A brief review is given of the use of sea-level records to infer variations of the Florida Current (see also chapter 11).

Iselin (1940a) in his report on the variations of transport of the Gulf Stream noted that sea-level measurement by tide gauges "provides a continuous and inexpensive record of the variations in the cross-current

density gradient, if it is assumed that the average surface velocity varies with the total transport of the current." The relation between sea-level and ocean currents had been used to infer variations of ocean currents (and vice versa) much earlier by Sandström (1903). Montgomery (1938b) first applied the method to the Florida Current using data from tide-gauge stations at Key West and Miami, Florida, and at Charlestown, South Carolina, from the eastern coast of the United States and from St. Georges harbor in Bermuda on the seaward side of the Florida Current. Variations in relative differences (absolute differences in heights of tide gauges were not determined) were examined as indicators of the strength of the mean surface current. Sea level, although reflecting tidal variation primarily and to a lesser extent local atmospheric pressure and winds, contains significant contributions also from the cross-stream slope necessary to balance Coriolis forces acting at the surface and the downstream slopes associated
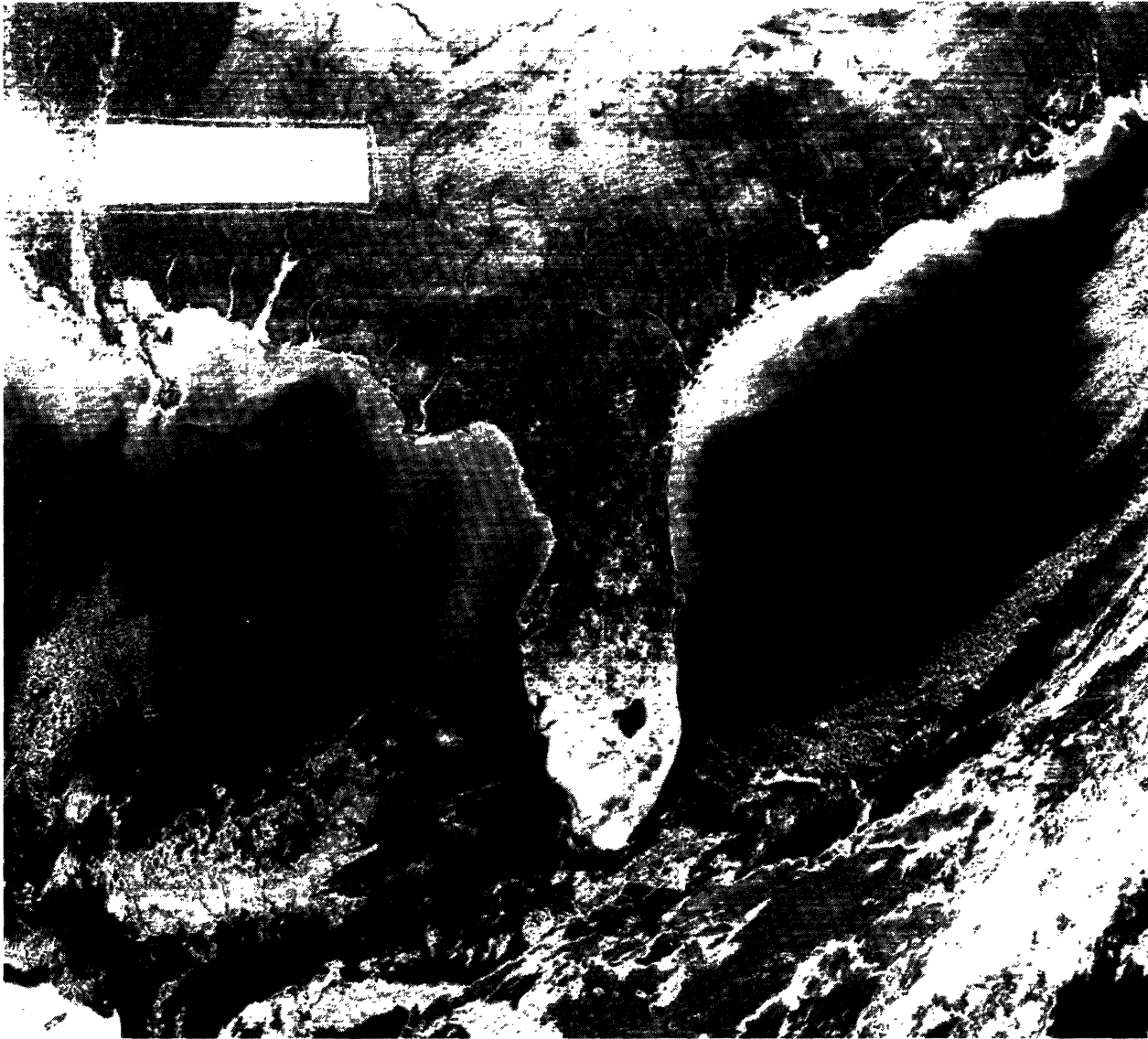
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Figure 4.3 The Florida Current between Miami and Cape Hatteras as seen in the infrared on February 26, 1975, from NOAA-4 satellite. A large eastward deflection occurs south of 32°N, possibly as a result of a topography feature. (Courtesy of R. Legeckis, NOAA-NESS.)

with accelerations or decelerations between stations. Montgomery (1938b) concluded from a 47-month study of mean differences of Bermuda minus Charleston that a seasonal cycle was present with the maximum difference and, hence, maximum surface current occurring in July and a minimum in October. The downstream difference of Key West minus Miami, based on 67 monthly values, showed a maximum hydraulic head in July with minima in November and February. As the gauges were not connected by geodetic leveling, the total hydraulic head was not known. Montgomery noted that the difference of 19 cm measured by leveling across the northern part of the Florida Peninsula would be adequate to accelerate the current off Miami to 193 cm s^{-1}, corresponding to maximum speeds observed. When leveling data were obtained, however, the drop in mean sea level from Key West to Miami was found to be only 4.9 cm, too small to account for the observed increase of speed between the two stations. Stommel (1953a) estimated that a difference of 20 cm is required to produce the observed acceleration to satisfy simple geostrophy and Bernoulli's equation. The lack of confirmation of the driving head by direct leveling forced Montgomery (1941) to conclude that the downstream differences between Key West and Miami could not be regarded as an indicator of the strength of the Florida Current. The cross-stream differences, however, clearly indicated a seasonal variation.

Schmitz (1969) reexamined Stommel's estimate using data from free-fall instruments obtained in the Florida Straits off Miami. He noted that the measured relative vorticity was considerably smaller than the value used by Stommel ($0.1f$ rather than $0.4f$, where f is the Coriolis parameter). Furthermore, the change of the Coriolis parameter between the Key West–Havana and the Miami–Bimini sections is approximately $0.1f$, offsetting the change in layer thickness necessary to conserve potential vorticity. Based on vorticity estimates, it apparently is not necessary to have a drop in head much larger than that found by land leveling. However, the observed maximum surface speeds in the Straits would indicate a considerably larger drop of 20 cm or more. It seems likely that the horizontal pressure gradient does not vanish with depth, so that the two-layer assumption of both Stommel (1953a) and Schmitz (1969) of a resting lower layer appears to be overly restrictive. Furthermore, the geodetic leveling may contain errors, and the actual drop in sea level could be larger than reported.

The disagreement between land-leveling and sea-level differences expected from the distribution of currents and density was examined by Sturges (1968). Using historical surface-current and wind observations, he calculated a surface topography for the western Atlantic near the Gulf Stream that represented a best fit

to the slopes estimated from the data. He concluded that the northward rise in sea level found by precise geodetic leveling along the east coast of the United States was inconsistent with his results. Sea level within the Gulf Stream must drop northward to maintain the northward flow. In a later paper, Sturges (1974) estimated the north–south slope from hydrographic-station data to be 0.8 cm deg^{-1} (centimeters per degree of latitude) upward to the north seaward of the Florida Current. From the estimated downstream increase in transport and the increase in magnitude of the Coriolis parameter, the cross-stream difference in level would require the inshore edge of the Gulf Stream to slope down 2.8 cm deg^{-1} relative to the seaward edge or a net downward slope of 2.0 cm deg^{-1} in the direction opposite to the land-survey results. Sturges concluded that the precise leveling surveys must contain systematic errors of undetermined nature that gave rise to the slight bias in meridional geodetic leveling.

4.3.2 Variability of the Florida Current

Speculation about the variability of the Florida Current was inspired not only by evidence from tide gauges but also from measurements of electrical potential using a telegraph cable from Key West to Havana, Cuba (Wertheim, 1954). The electrical potential induced by flow of sea water through the earth's magnetic field, shunted partially by the conducting sea floor, provides a signal that is correlated with the transport. The variations of nontidal flow appear to be exaggerated in the electrical potential (Schmitz and Richardson, 1968) because of shifts of the Florida Current relative to the bottom topography. The cause of these shifts was not determined. Maul et al. (1978) have speculated that meanders of the Loop Current in the Gulf of Mexico may affect the Florida Current. Sanford and Schmitz (1971) concluded that induced electrical potential was more closely correlated with the transport at the Miami section. The estimated error was found to be about 10% of the mean compared to a factor of two changes for the Key West section.

Supporting evidence for the seasonal variation of the Florida Current found by Montgomery (1938b) in the tide-gauge data came from other sources. Fuglister (1951) used monthly mean current speed and direction charts from an atlas published by the U.S. Navy Hydrographic Office (1946) to estimate seasonal variability in 10 regions following the Gulf Stream System from Trade Wind Latitudes to beyond the Grand Banks. He found that the maximum currents occurred in summer (July) in southern portions and in winter in northern portions, while the minimum tended to occur in fall (September to November) in all regions. The seasonal variability first seen on tide gauges was also confirmed by direct measurements. Niiler and Richardson

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(1973), using a 7-year study of volume-transport measurements at a cross-stream section at Miami, found that the annual variation accounted for 45% of the total variability, with transports ranging from an early winter low of 25.4×10^6 m³s⁻¹ to a summer high of 33.6×10^6 m³s⁻¹. The average transport was 29.5 m³s⁻¹. They also reported a strong short-period (2 weeks or less) modulation of the velocity and density structure that was as strong as the seasonal change. These rapid fluctuations have been studied intensively since (Brooks, 1975; Wunsch and Wimbush, 1977; Schott and Düing, 1976).

Wunsch, Hansen, and Zetler (1969) extended the analysis of sea-level record and sea-level differences over the remainder of the frequency range contained in the tide-gauge data. They examined simultaneous records from Key West and Havana (1953–1956), Key West and Miami (1957–1962), and Miami and Cat Cay (1938–1939). The power levels below tidal frequencies were low at all sites. About half the power was in seasonal variations with no detectable peaks between seasonal and tidal frequencies. Coherences between stations were small and, where detectable, showed zero phase across the Florida Current, consistent possibly with a Bernoulli effect or large-scale atmospheric forcing. Lack of downstream coherence was attributed to Doppler shifting because of different mean speeds between stations. An inverted-barometer response to atmospheric-pressure forcing could be detected at lower frequencies (periods 10–128 days), and a direct response at higher frequencies.

4.3.3 Eddy–Mean Flow Interaction

A downstream pressure gradient is called for in inertial models of westward intensification. In the simplest model of this type, the frictionless, homogeneous circulation on a β-plane described by Fofonoff (1954), the pressure and free surface drop along the western boundary as the flow intensifies. The lowest pressure is found at the boundary where the highest speeds are attained. Along each streamline, pressure is related to speed by the Bernoulli equation. In the model, the highest speeds and lowest pressures (and hence sea levels) are found at the boundary. In the real ocean, the sea surface within the Florida Current has to be matched to a coastal boundary region, implying that the pressure gradient is continued into the coastal region. This, in turn, implies an active dynamic regime inshore of the Florida Current. Several studies have described the fluctuations within and adjacent to the Florida Current in some detail.

Von Arx, Bumpus, and Richardson (1955) observed a succession of short, overlapping segments that they described as "shingles" extending from the Florida Straits past Cape Hatteras. These shingles were first noted during an attempt to follow the Florida Current

with an airborne infrared radiometer. The inshore edge was found not to be continuous in its thermal structure but made up of a series of fronts. They speculated that the cause might be tidal modulation of the Florida Current emerging from the Florida Straits, but admitted that no sound basis had been found for a relation between tides and the short-term fluctuations observed. These structures in the thermal field could be interpreted as instabilities of the Florida Current and evidence for exchange of energy between the mean flow and a time-dependent field.

Webster (1961a, 1961b, 1965) analyzed these and other surface-velocity measurements made during repeated crossings of the Florida Current at sections off Miami and Jacksonville, Florida, and off Onslow Bay and Cape Hatteras, North Carolina, to estimate Reynolds stresses associated with the nontidal velocity fluctuations present in the flow. One of the objectives of the study was to evaluate the magnitude of eddy–mean flow interactions within the Florida Current. The surface currents were estimated using a towed GEK (Von Arx, 1950), which responds to the current component perpendicular to the ship's track. In some cases, currents were inferred from the ship's set during crossings. The repeated crossings enabled Webster to compute means and fluctuations of the cross stream $(\bar{u}, u')$, the long-stream $(\bar{v}, v')$ components of surface flow, and the momentum-flux component $\rho\overline{u'v'}$ in several zones across the Florida Current. The Reynolds-stress component τ_{xy} corresponding to this eddy momentum flux is $-\rho\overline{u'v'}$. At nearly all sections, the velocity correlations were positively correlated ($\rho\overline{u'v'} > 0$), implying that momentum was being transported offshore and that the Florida Current was exerting a negative (southward) stress on the coastal region. As the northward flow $\bar{v}$ increases offshore, the momentum has to be transported into regions of increasing mean flow against the mean-velocity gradient. Thus, the slower-moving coastal waters appear to exert an accelerating stress on the swiftly moving Florida Current offshore, a result that is opposite to the intuitive expectation that the Florida Current might tend to be retarded by the coastal boundary and lose momentum to it. Webster calculated also the rate of work W done by the Reynolds stresses on the mean flow from the term

$$W = \bar{v}\frac{\partial \tau_{xy}}{\partial x} = \frac{\partial \bar{v}\tau_{xy}}{\partial x} - \tau_{xy}\frac{\partial \bar{v}}{\partial x}.$$

Integration from a straight coast $x = 0$ to the axis of the current $x = L$ (Webster integrated across the entire current) yields the total work per unit time within the coastal strip inshore of the current axis:

$$\int_0^L w\,dx = \bar{v}\tau_{xy}\big|_{x=L} - \int_0^L \tau_{xy}\frac{\partial \bar{v}}{\partial x}\,dx.$$

Assuming $\bar{v}$ to be zero at the coast, the total work done in the coastal strip is equal to the work done on the seaward boundary $(\bar{v}\tau_{xy})$ plus the eddy work on the mean flow within the strip. For a steady state to exist, the two terms must balance; otherwise, the mean flow in the strip would have to gain or lose energy at a rate equal to the difference between the two terms, assuming that other terms, such as work against pressure gradients, neglected above, remain small. Webster examined the eddy-mean flow term at each section and concluded that the net energy transfer was from eddy to mean flow for all sections. As a consequence of the eddy-mean flow interaction, the inshore strip is doing work on the Florida Current and therefore must contain an energy source to supply the offshore flux of momentum and energy. Schmitz and Niiler (1969) reexamined Webster's estimates and analyzed additional measurements made by free-fall instruments that confirmed the earlier conclusions about significant eddy-to-mean energy flux within the coastal region of cyclonic shear. They found, in addition, a region of negative velocity correlation in shallow depths close to shore, indicating a region of retarding stress and flow of momentum to the shore. This feature was not observed by Webster in Onslow Bay presumably because his sections did not approach close enough to the coast. Lee (1975) and Lee and Mayer (1977) describe recent measurements in this dissipative near-shore strip in the Florida Straits. Schmitz and Niiler (1969) found that the total energy flux integrated across the entire width of the current was not significantly different from zero within each section. They concluded that although a region of intense energy transfer from eddy to mean flow existed, it was offset by a wider region of mean-flow-to-eddy transfer over the rest of the current, resulting in a redistribution of energy that required no external energy source.

Brooks and Niiler (1977) carried out a comprehensive study of historical and new transport-profile data for a section across the Florida Current in the vicinity of Miami. Their estimates showed that statistically significant conversions of kinetic and potential energy between fluctuations and mean flow occurred in either direction in parts of the section, but the net conversion rates were too small to be dynamically important. Based on these rates, the decay time for the total perturbation energy was about 50 days, much longer than the residence time for the Florida Current in the Florida Straits. They concluded that pressure gradients must be present to balance the energy flow. The coupling between mean flow and fluctuations may, in fact, be rather weak compared to the major energy conversion between mean potential energy and mean kinetic energy, with the fluctuations playing a minor or negligible role. Such a model is also suggested by the distribution of surface velocity and kinetic energies of the mean and eddy flow tabulated for the Florida Current by Hager (1977) from ship-drift reports collected by the U.S. Hydrographic Office for the period 1900-1972. While these data are not of the same quality as direct measurements, they reveal clearly the spatial extent of the Florida Current and its region of intensification as it flows into the Florida Straits. The peak currents and kinetic energies appear to be underestimated by the dead reckoning used to compute ship drift because of the spatial averaging involved. Hager found that the eddy kinetic energy was comparable to the mean-flow kinetic energy in the Loop Current and in the Gulf Stream past Cape Hatteras. However, within the Florida Straits, the eddy kinetic energy ($1-2 \times 10^3$ cm^2 s^{-2}) was much smaller than the mean-flow kinetic energy ($>10^4$ cm^2 s^{-2}) and showed little similarity in its spatial distribution. These results suggest that the fluctuations are not essential to the intensification of the mean flow in the Florida Straits.

4.3.4 The Downstream Pressure Gradient

The downstream pressure gradient is important to the energetics of the Florida Current because it provides the simplest mechanism for converting potential to kinetic energy within the Florida Current (Webster, 1961a). The development of satellite radar altimetry with precision and resolution capable of detecting differences in surface elevation of less than a meter (Vonbun, Marsh, and Lerch, 1978) has created the opportunity to use sea-level slopes to infer behavior of current fluctuations in considerably greater detail than possible with surface observations only. The interpretation of surface slopes and internal pressure gradients related to these slopes will become increasingly important as altimetry measurements are accumulated (see figure 4.11).

The Florida Current increases in transport as it flows northward along the continental slope to Cape Hatteras (Iselin, 1936; Richardson, Schmitz, and Niiler, 1969; Knauss, 1969). Its momentum, energy content, and flux increase, implying the presence of strong energy sources within the Florida Current and perhaps the surrounding regions. As the increasing momentum and energy within the Florida Current is most likely produced by a downstream pressure gradient acting to accelerate the flow, the most probable source of energy for the inshore region is a continuation of this downstream gradient into the coastal region.

Godfrey (1973) has given a clear physical interpretation of the effects of a downstream (northward) pressure gradient based on an examination of a six-layer numerical model reported by Bryan and Cox (1968a,b). The longshore pressure gradient was well developed in the upper layers and weakened with depth along the western wall. The drop was equivalent to about 1 m at the 100-m level and had reversed sign at 1600 m. Be-

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cause a balancing geostrophic flow would have to be outward from the coast, a complete geostrophic balance is impossible. The gradient must be balanced partly by an outflow, causing upwelling along the boundary, and partly by downstream acceleration. The upwelling along the coastal boundary implies shoreward motion at depth. Godfrey used the model to interpret eddy formation in the East Australian Current, but it had been developed originally by Bryan and Cox with application to the Gulf Stream in mind.

Blanton (1971, 1975) presented evidence for a vigorous movement of shelf water into the Florida Current and intrusion of Gulf Stream Water from the Florida Current along the bottom onto the North Carolina Shelf off Onslow Bay in summer. A section taken on July 22, 1968, showed Gulf Stream Water covering the entire shelf, with shelf water forming an isolated lens in the upper layer at mid-shelf. A month earlier, Gulf Stream Water had shown only a slight intrusion at the shelf break (40-m depth). Many other factors may be present. The driving mechanism, whether dominated by pressure gradients originating in the Florida Current as described by Godfrey (1973) or by local winds, has not been clearly established. The occurrence of strong upwelling and exchange with the coastal region is apparent, however, and may be evidence of a current-induced pressure gradient on the shelf.

The mechanism by which the pressure gradient can supply momentum to the eddies and not to the mean flow remains obscure. Because the meanders described by Webster (1961b) move downstream in the direction opposite to the propagation of topographic Rossby waves, the mechanism of wave-momentum transport suggested, for example, by Thompson (1971, 1978) does not appear to be appropriate.

Pedlosky (1977) studied the radiation conditions for a linear two-layer ocean model to propagate waves away from a forcing region consisting of a sinusoidal moving zonal boundary. Eastward-moving meanders can radiate into the ocean interior only if their phase speed is less than the local interior velocity. If the local interior velocity is westward, the eastward-moving meanders cannot radiate in either baroclinic or baro-

tropic Rossby-wave modes. For nondivergent flows over a sloping north–south boundary, such as the continental shelf, these results seem to imply that topographic waves in the shallow coastal region cannot be coupled to northward-moving meanders. Other mechanisms may be possible. Webster (1961a) noted that "each of the meanders resembles a sort of skewed wave motion and consists of an intense offshore running current (time 1 to 4 days) followed by a broad diffuse flow onshore time 4–7 days then followed by another intense offshore current." The intense offshore jets shown in figure 4.4 may be similar to inertial jets formed along western boundaries. The sloping shelf provides a strong topographic β-effect in the coastal strip. Currents that flow offshore down a pressure gradient and across depth contours on the shelf so as to conserve potential vorticity would intensify into narrow jets with strong cyclonic relative vorticity that may be incorporated into the cyclonic inshore region of the Florida Current. Such jets could carry momentum and energy offshore. If the instantaneous downstream (northward) pressure gradient were concentrated across a narrow jet, the transfer of momentum into the Florida Current would be readily accomplished. The existence of intensifying jets detached from solid boundaries has not been established, however, so that this line of reasoning must be considered speculative. Most theoretical studies applicable to the Florida Current assume that the basic flow is nondivergent with zero downstream pressure gradient. It is possible that the neglect of the pressure gradient excludes relevant mechanisms of meander formation and may exaggerate the role of eddy–mean flow interactions in numerical models.

4.3.5 Stability and Atmospheric Forcing
Mechanisms for the conversion of kinetic and potential energy associated with the mean flow to perturbations have been examined in several studies of the stability of the Florida Current.

Orlanski (1969) developed a two-layer model for two cases of bottom topography in the lower layer resembling the continental slope under the Gulf Stream and
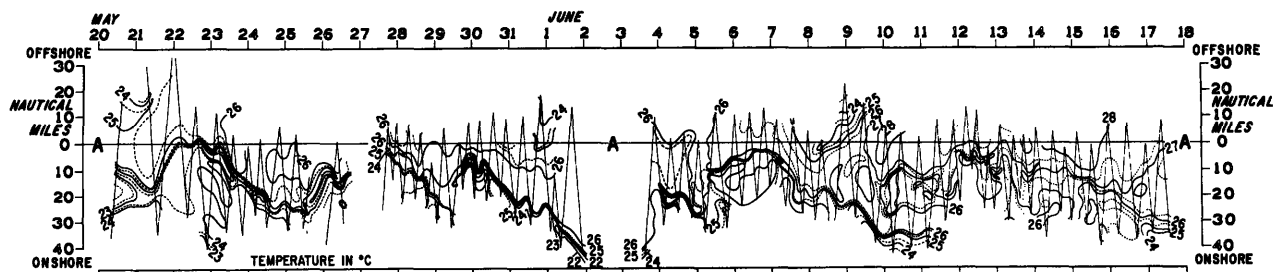

Figure 4.4 Space-time variation of temperature off Onslow Bay, showing movement of temperature fronts with 4–7-day time scale. Note that the offshore motion was discontinuous

or more rapid than the onshore motion of the front. (Webster, 1961b.)

the continental rise in the open ocean further downstream. The model has a constant Coriolis parameter with no downstream variation of the basic current, pressure fields, or topography. Orlanski found that a necessary condition for instability to occur is that the gradient of potential vorticity of the basic flow be of opposite sign in the two layers. As only cross-stream variation occurs, the stability depends critically on the slope of the interface between the two layers relative to the bottom slope. The change of thickness of the bottom layer across the current can determine the sign of the potential-vorticity gradient and hence the stability. The most unstable modes found by Orlanski are given in table 4.1. Orlanski and Cox (1973) reexamined the stability of the western boundary current in a three-dimensional numerical model. The model had better resolution in the vertical (15 levels) but was periodic along the coast, thus excluding a downstream pressure gradient. Nonlinear terms and a β-effect were included in the model. Instabilities developed as predicted by linear theory but with a growth rate about double that of the simpler two-layer model. The growth rate decreased by an order of magnitude as finite amplitude was attained.

Niiler and Mysak (1971) analyzed a barotropic, constant-f model in which the velocity distribution and bottom topography of the continental shelf were approximated by segments of constant potential vorticity and depth. Unstable barotropic waves were possible in the model because the potential vorticity was chosen to contain maxima in its distribution across the current. The arguments for these extrema are that the cyclonic shear in the inshore region raises the relative vorticity sufficiently to overcome the opposing effect of increasing depth of the shelf and slope. Thus if the slope is small enough, a maximum occurs in potential vorticity. A region of anticyclonic shear on the seaward side of the Florida Current over a slowly varying depth yields a minimum in the cross-stream distribution.

These extrema in the potential-vorticity distributions imply the existence of unstable barotropic modes. With no basic current, the solutions are shelf and topographic waves already discussed by Robinson (1964) and Rhines (1969a) (see chapters 10 and 11). With a basic northward current, the southward-traveling waves can be reversed and made unstable. The most unstable barotropic mode on the Blake Plateau has a period of about 10 days and a wavelength of 140 km and can reach finite amplitude in a few wavelengths downstream. Because these barotropic waves require bottom topography to induce regions of instability, their growth is not sustained in deep water. Here the unstable waves were found to have a period of 21 days and a wavelength of 195 km. The authors suggest that the unstable shelf modes can be triggered by narrow fast-moving frontal systems. These short-period waves increase in amplitude as they move northward to deep water, where they are no longer unstable because of the change in potential-vorticity structure of the basic deep flow but may persist as a smaller-scale structure on the longer and slower meanders that develop downstream.

Brooks (1978) has also pointed out the importance of wind stress and its curl as a forcing mechanism for shelf waves. He concludes that strong coupling can occur for periods that are less than or greater than the zero group-velocity period of barotropic shelf waves for the continental shelf off Cape Fear (i.e., 2.5–3.5 and >10 days, respectively). The model was used to interpret correlations between atmospheric-pressure variations and winds and sea-level variations from tide gauges at Beaufort and Wilmington, North Carolina. Recently, Brooks and Bane (1978) reported that deflections of the Florida Current are induced by a small topographic feature in the continental slope off Charlestown, South Carolina. Satellite observations of thermal patterns (figure 4.3, for example) show considerable difference in amplitude upstream and down-

Table 4.1 Characteristics of Perturbations Found for the Florida Current and Gulf Stream

Author	Wavelength (km)	Period (days)	Growth rate (days^{-1})	Phase speed (cm/s)	Type
Orlanski (1969)					
Shelf waves	220	10	1/5		Baroclinic
Deep ocean	365	37.4	1/7.23		Baroclinic
Orlanski and Cox (1973)	246		1/12.1		
Niiler and Mysak (1971)					
Shelf	140	10	1/13	14	Barotropic
Ocean	195	21	1/13	9	Barotropic
Brooks (1975)	190	12	0	South	
Schott and Düing (1976)	170	10–13	0	South (17 cm s^{-1})	From current measurements

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stream of the "Charlestown Bump" located near 32°N. Stumpf and Rao (1975) suggested possible topographic influences in studying a sequence of infrared images of meanders off Cape Roman and Cape Fear. They point out that a well-coordinated field experiment would be necessary to distinguish wind forcing from topographic influences or instability of the Florida Current.

Schott and Düing (1976) found southward-traveling waves in the Florida Straits based on velocity measurements from three moored buoys located close to the same isobath at 335 m near the "approximate location of the axis of Gulf Stream" according to nautical charts. Records were obtained for a duration of 65 days from a depth of about 300 m. The most likely wave parameters were fitted by least-square methods to 36 independent auto- and cross-spectra. A significant fit was found in the 10–13-day spectral band for a wavelength of 170 km traveling south at 17 cm s^{-1}. These are identified as stable continental-shelf waves probably generated by passage of atmospheric cold fronts. The parameters obtained agree well with a model by Brooks (1975) that included realistic topography and horizontal current shear to yield a southward-propagating wave of 12-day period and 190-km wavelength at maximum response to forcing by cold fronts. The characteristics of these wave models are summarized in table 4.1.

The observed coherence with meteorological events noted by Wunsch and Wimbush (1977) and Düing, Mooers, and Lee (1977) may be a consequence of the weak coupling between mean flow and the fluctuations. The perturbations apparently can receive a significant fraction of their energy from atmospheric forcing rather than from the mean flow and consequently show measurable correlation with wind events.

Several mechanisms for generation of meanders in the Florida Current have been identified: barotropic and baroclinic instability in the presence of topography; bottom features forcing deflections and downstream lee waves; and excitation of propagating waves by atmospheric forcing. Nonlinear mechanisms are yet to be explored, as are the effects of the downstream inhomogeneity of the Florida Current.

Richardson et al. (1969) found that the transport of the Florida Current increased relatively slowly (17%) through the Florida Straits from Miami to Jacksonville with a slight increase in surface speeds and a shift to westward of the current axis. A larger increase in transport (67%) was found from Jacksonville to Cape Fear with a slight decrease of maximum surface velocity and a broadening of the current. The effects on the instability modes of the downstream increase in transport are not known. Other intermittent perturbations to the Florida Current are the passage of rings and eddies seaward of the Florida Current. These apparently can be swept into the current (Richardson, Strong and Knauss, 1973). The consequences of such events

are not known. The deep western boundary current predicted by Stommel (1957b) and the first observed by Swallow and Worthington (1957, 1961) has been found in recent studies to contain strongly time-dependent components (Riser, Freeland, and Rossby, 1978). The effects on the Florida Current are not adequately known at present but may be profound.

4.3.6 The Deep Western Boundary Current

If an upwelling velocity of broad horizontal scale is assumed in the deep interior flow of an ocean, Stommel (1957b) showed that the conservation of mass and potential vorticity cannot be satisfied by geostrophic flow alone. A deep western boundary current is necessary to allow both constraints on the deep flow to be met. In the North Atlantic, Stommel concluded that a southward flow should be present along the continental slope. This prediction together with the development of the neutrally buoyant float for measuring current by Swallow (1955) led Swallow and Worthington (1957, 1961) to measure the deep flow off Cape Romain, South Carolina, near the northern end of the Blake Plateau, where the flow was expected to lie seaward of the strong surface current over the Blake Plateau. Southward flows of 9–18 cm s^{-1} measured over a period of a month led them to conclude that the deep western boundary current is a persistent feature of the circulation along the continental slope. The transport of the undercurrent was estimated to be 6.7 × 10^6 m^3 s^{-1}.

Subsequent measurements by a number of investigators (Volkmann, 1962; Barrett, 1965; Worthington and Kawai, 1972; Richardson and Knauss, 1971; Amos, Gordon, and Schneider, 1971; Richardson, 1977) reported transports ranging from 2 to 50 × 10^6 m^3 s^{-1} with an average of 16 × 10^6 m^3 s^{-1}. The flow is persistent, though apparently quite variable. Westward and southward deep currents along the continental rise north of the Gulf Stream have been reported by Webster (1969), Zimmerman (1971) and Luyten (1977). Recent measurements using SOFAR floats (Riser, Freeland, and Rossby, 1978) show flow south of the Blake-Bahama Outer Ridge along the Blake Escarpment. The southward flow may be simply a consequence of a deep westward flow onto the sharply rising topography. Steady slow geostrophic flows are constrained to follow contours of f/h, where h is depth, which are concentrated along the slope. Near Cape Hatteras, where the Gulf Stream crosses over the deep current (Richardson, 1977), the combined effect of vortex stretching within the northward-moving current and the deep flow crossing the bottom slope would be felt. Holland (1973) has examined the enhancement of transport in the western boundary current in a numerical model including baroclinicity and topography. The vortex stretching in the stratified upper layers must counteract changes of f only to conserve potential vorticity, whereas the deep

water can be subjected to a much larger stretching by crossing depth contours. Thus a relatively weak deep flow crossing depth contours can have a vertical velocity equal and opposite to a large meridional baroclinic flow. This type of flow was described briefly by Fofonoff (1962a) under a general class of thermohaline transports.

 Slow steady barotropic flow must be along contours of f/h or must cross contours of f/h at a rate that balances the combined divergence of baroclinic and Ekman flow. For deep flow onto the continental slope, the intensification of the current on the slope can be estimated from the potential-vorticity equation along each streamline:

$$\frac{f_0}{h_0} = \frac{f_0 + \beta y}{h_0 - s_x x - s_y y},$$

where f_0, h_0 are open-ocean values of Coriolis parameter and depth, and s_x, s_y the (constant) bottom slopes. For a slope width Δx, the deep streamlines are displaced equatorward by an amount Δy, where

$$\Delta y = \frac{\beta_x}{\beta_y} \Delta x = \frac{f_0 s_x / h_0}{(\beta + f_0 s_y / h_0)},$$

and β_x, β_y are the horizontal gradients of f/h_0. From the sketch in figure 4.5, it is seen that the narrowing or intensification over the slope is

$$U = U_0 \frac{h_0}{h} \frac{\sqrt{\Delta x^2 + \Delta y^2}}{\Delta x} = U_0 \frac{h_0}{h} \sqrt{1 + \left(\frac{f_0 s_x}{\beta h_0}\right)^2}$$

$$\sim \frac{U_0 f_0 s_x}{\beta h_0} \quad \text{(for } s_y = 0\text{)}.$$

For $h_0 = 2500$ m, $f_0 = 10^{-4}\,\text{s}^{-1}$, $\beta = 2 \times 10^{-13}$ cm s^{-1}, $s = 1/100$ (continental rise),
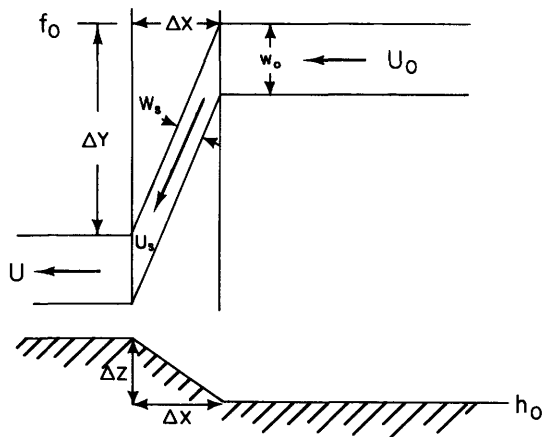
$$U = 20 U_0.$$

Figure 4.5 Displacement Δy of a current flowing over a bottom slope of width Δx on a β-plane. The velocity U_s on the slope is magnified by the ratio of widths w_0/w_s. Relative vorticity is assumed to be small.

For the continental slope (e.g., $s = 1/15$, $h = 1500$ m),

$$U = 200 U_0.$$

The intensification even over the gentle continental rise is sufficient to magnify flows U_0 that are below a measurable level in the interior to observable velocities on the rise and slope. Thus, it is very difficult to determine by direct measurement whether the flow over the continental rise is being forced by an upslope or downslope component.

The main thermocline deepens northward (Iselin, 1936) on the seaward side of the Florida Current, intensifying the apparent β-effect below the thermocline. The deep flow must move southward to conserve potential vorticity. Within the Gulf Stream itself, the thermocline rises sharply downstream. The rise is equivalent to $s_y < 0$ in the lower layer. Furthermore, the thermocline slopes sharply downward in the x-direction because of the shear across the thermocline. If the thermocline slopes are denoted by T_x, T_y, the lower-layer-displacement equation becomes

$$\Delta y = \frac{f_0 (T_x - s_x)/h_0}{\beta - f_0 (T_y - s_y)/h_0} \Delta x.$$

The displacement Δy is no longer necessarily southward along the western boundary. Northward deep flows are permitted by the vorticity equation if

$$T_x - s_x < 0 \quad \text{or} \quad T_y - s_y > \frac{\beta h_0}{f_0}.$$

These flows would likely be unstable because the potential-vorticity gradient would then be of opposite sign in the two layers.

According to simple potential-vorticity conservation, westward deep flow on reaching the continental rise should turn southward and continue to have a southward component as long as the main thermocline slopes downward to the north. The decreasing thickness of the deep-water layer has to be compensated by decreasing the Coriolis parameter. In the region of the accelerating western boundary current, the thermocline slope is reversed and the constraint on the deep flow is altered. The current may then turn northward if the thermocline slope is sufficiently large. The circulation diagram given by Worthington (1976) for the deep water (potential temperature $\theta < 4°C$) reproduced in figure 4.6 has southward flow along the continental slope with northward flow further to the east opposite to the deep flow expected based on the elementary potential-vorticity arguments given here. The present interpretation of the deep circulation in western North Atlantic and its interaction with the deep boundary current is not consistent with potential-vorticity conservation and needs further development.

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