4 The Gulf Stream System

N. P. Fofonoff

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

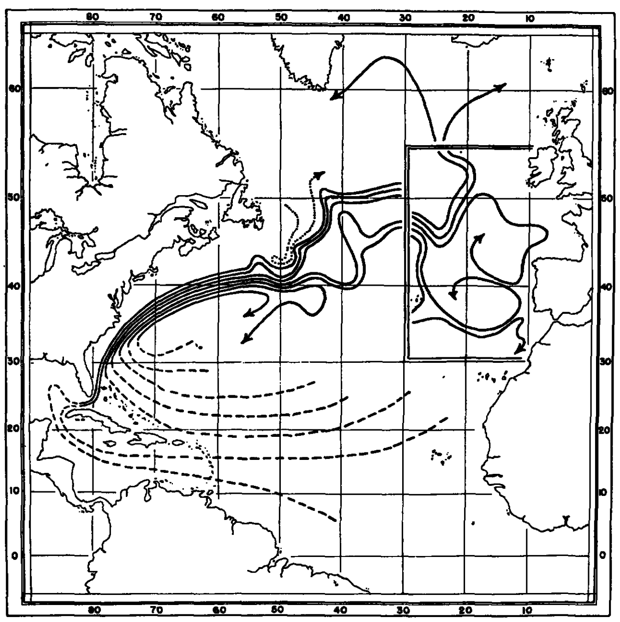

Figure 1

Figure 4.1 A schematic diagram showing the Gulf Stream System as described by Iselin (1936). Each streamline represents a transport interval of about 12×10^6 $m^3 s^{-1}$.

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.

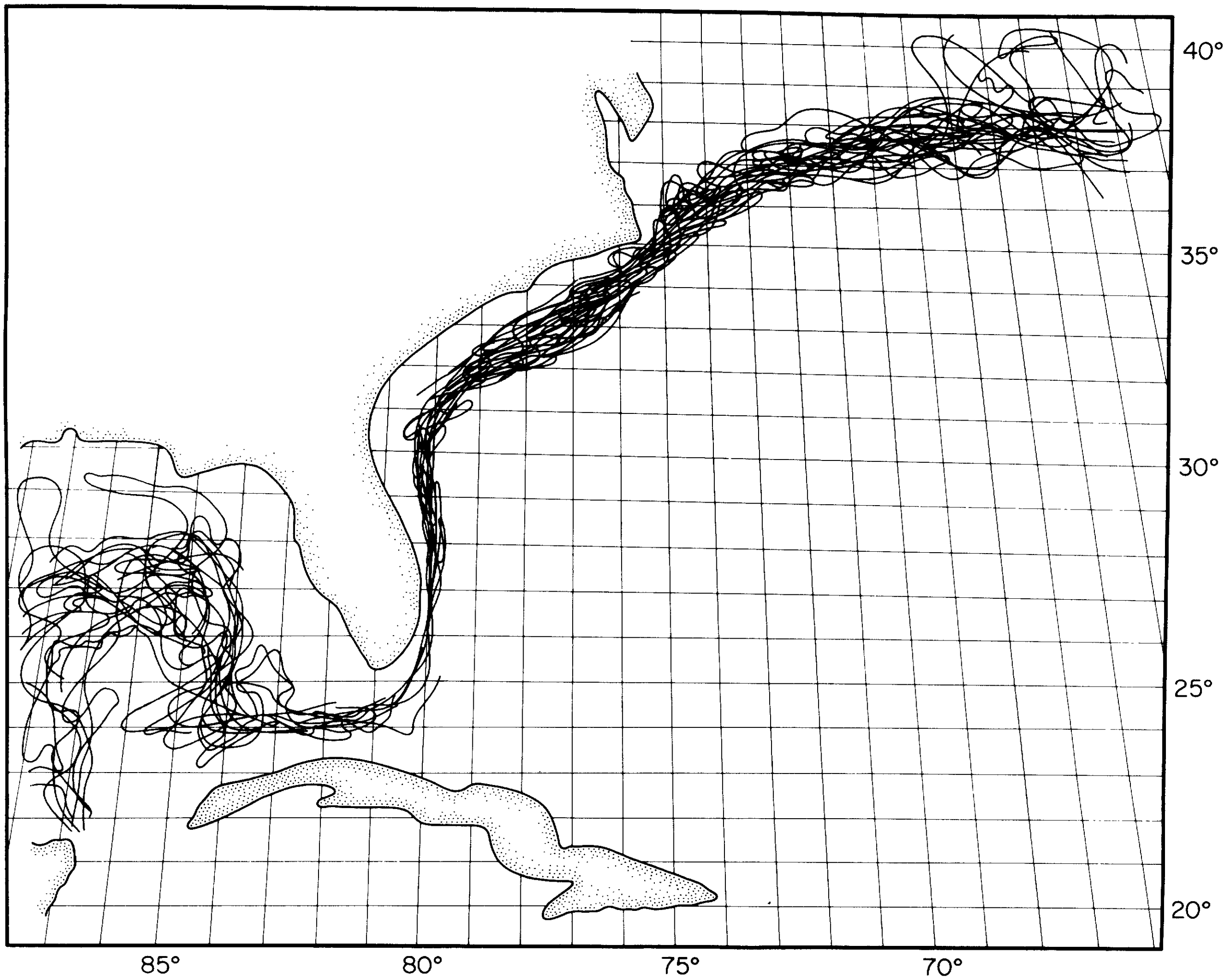

Figure 4.2 A composite of thermal fronts of the Gulf Stream System showing variability for a 9-month period between February and November 1976. [Courtesy of G. Maul from Geostationary Operational Environmental Satellite (GOES) infrared observations.]

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

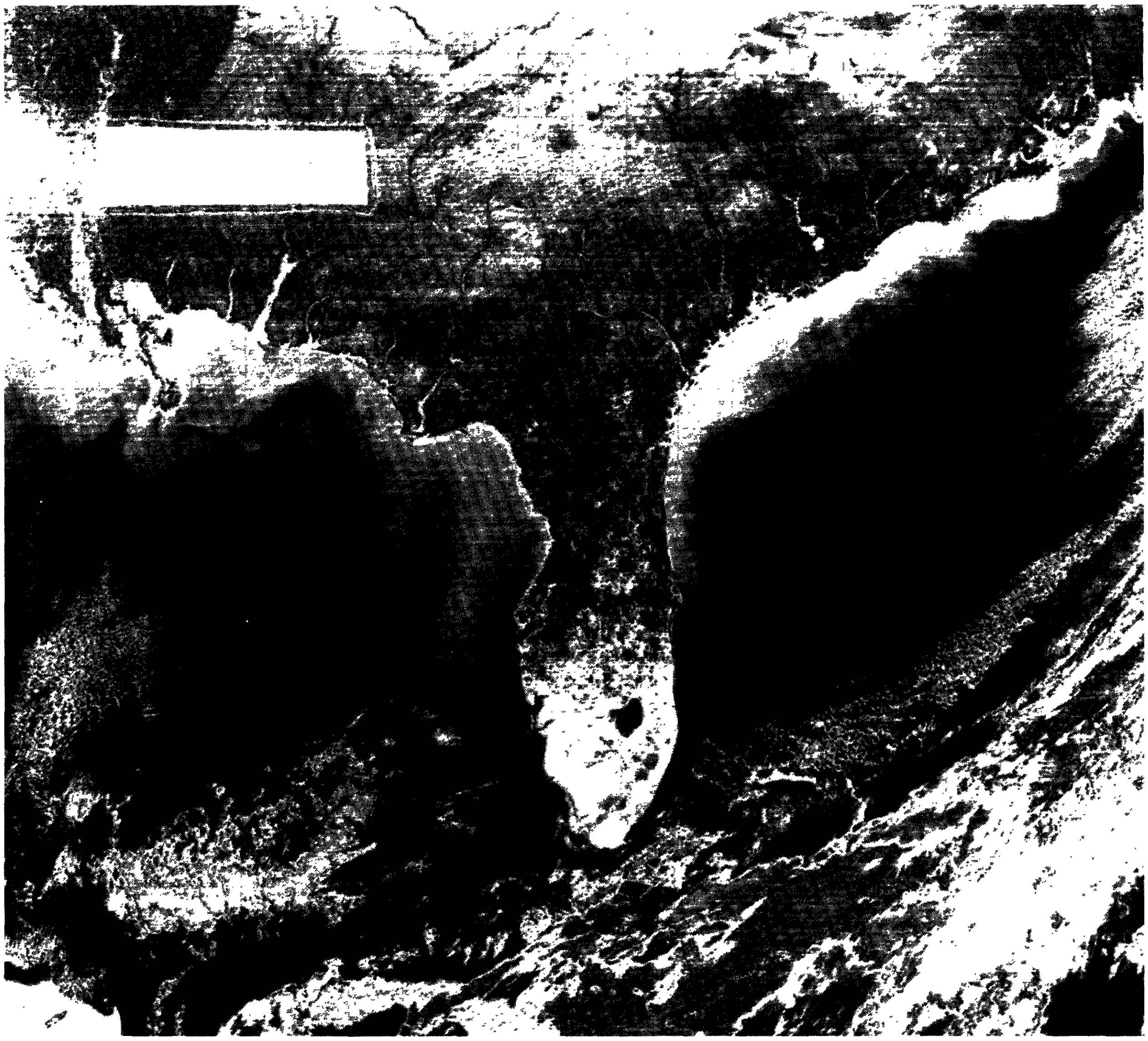

Figure 4.3 The Florida Current between Miami and Cape Hatteras as seen in the infrared on February 26, 1975, from NOAA-4 satellite. A large eastward deflection occurs south of 32°N, possibly as a result of a topography feature. (Courtesy of R. Legeckis, NOAA-NESS.)

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

(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 m^3 s^{-1} to a summer high of 33.6×10^6 m^3 s^{-1}. The average transport was 29.5 m^3 s^{-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 ($1\text{-}2 \times 10^3\ \mathrm{cm^2\,s^{-2}}$) was much smaller than the mean-flow kinetic energy ($>10^4\ \mathrm{cm^2\,s^{-2}}$) 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. Because 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 barotropic 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

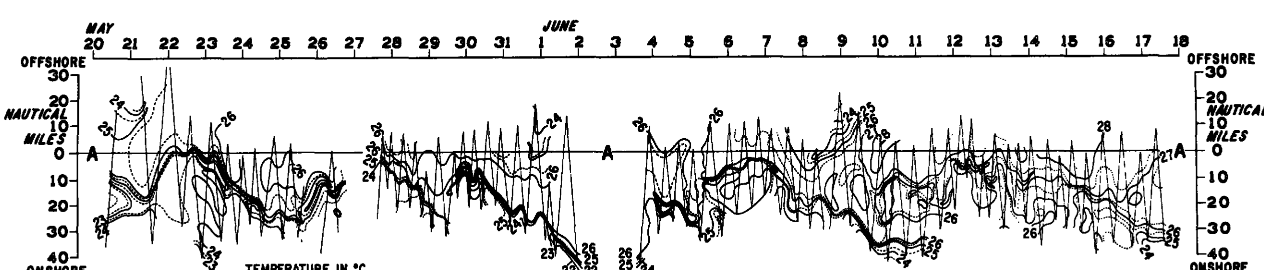

Figure 4.4 Space-time variation of temperature off Onslow Bay, showing movement of temperature fronts with 4–7-day time scale. Note that the offshore motion was discontinuous or more rapid than the onshore motion of the front. (Webster, 1961b.)

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

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

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

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

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

$$U = 20U_0.$$

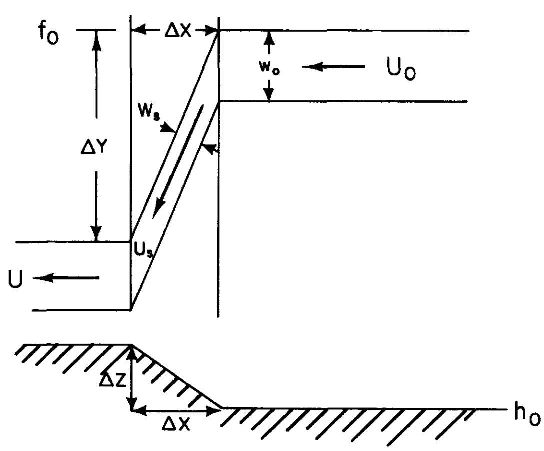

Figure 4.5 Displacement Δy of a current flowing over a bottom slope of width Δx on a β-plane. The velocity U_s on the slope is magnified by the ratio of widths w_0/w_s. Relative vorticity is assumed to be small.

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

$$U = 200U_0.$$

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

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

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.

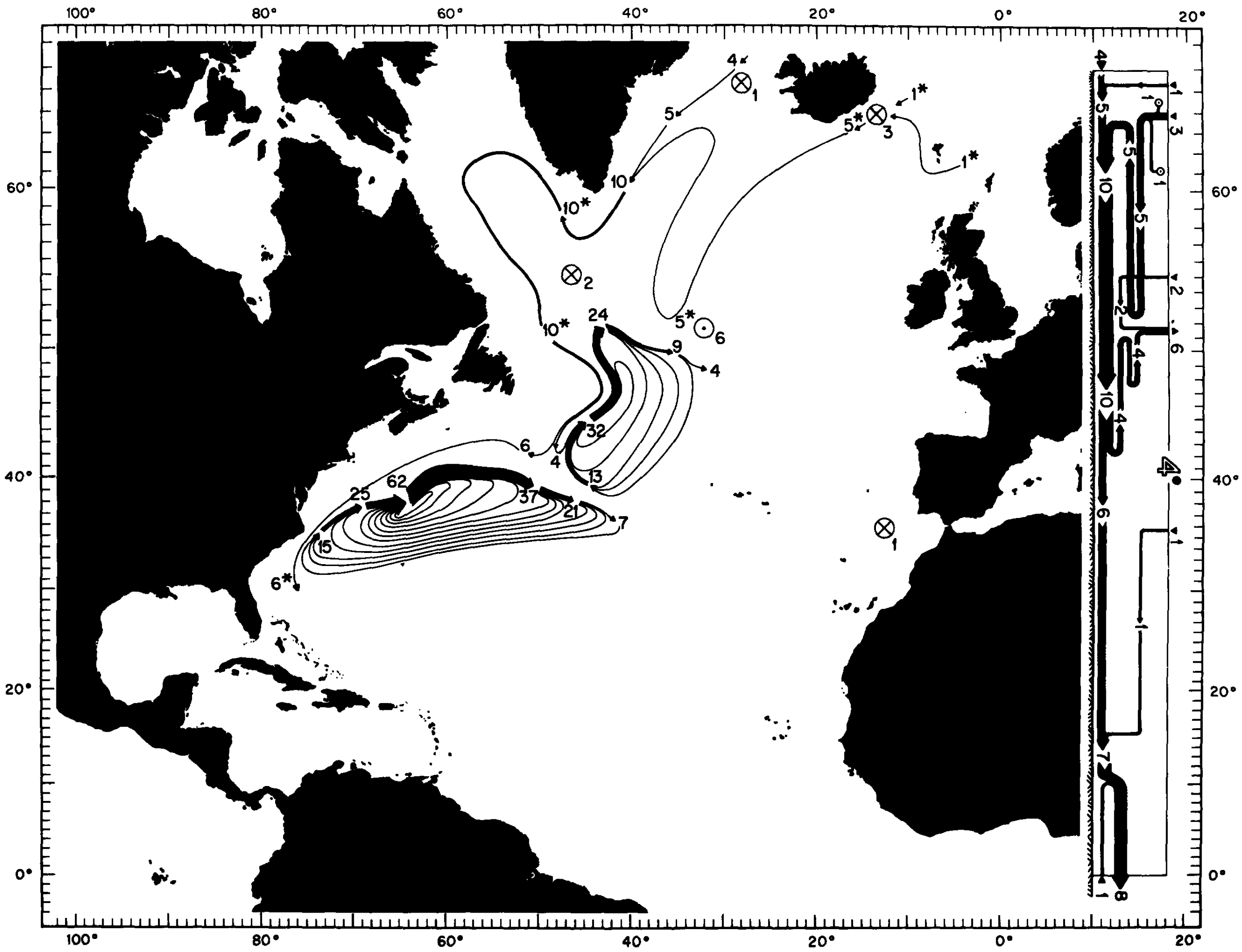

Figure 4.6 Circulation diagram for the deep ($\theta < 4°C$) circulation in the North Atlantic. Worthington estimates a flow of $62 \times 10^6\ m^3 s^{-1}$ in the recirculation gyre with $6 \times 10^6\ m^3 s^{-1}$ flowing southward inshore of the recirculation. (Worthington, 1976.)

4.4 The Gulf Stream

The portion of the Gulf Stream System from Cape Hatteras to the Southeast Newfoundland Rise is the *Gulf Stream* according to the nomenclature introduced by Iselin (1936). The Gulf Stream is separated from the continental shelf to the north by a band of westward-flowing Continental Slope Water (McLellan, 1957; Webster, 1969) and bounded to the south by the westward recirculation gyre described by Worthington (1976) and Schmitz (1978). Thus, the Gulf Stream is an eastward-flowing current flanked on either side by broader regions of westward flow. As the Gulf Stream is assumed not to be locally driven, enough energy and momentum must be carried by the flow into the region to maintain the eastward motion and the eddy and circulation fields in the surrounding areas. Fuglister (1963) noted that the Gulf Stream leaving Cape Hatteras flows approximately along a great circle while the continental slope turns northward. There is no pronounced curvature of the Gulf Stream on entering deep water, as might be expected from the increasing depth. The lack of a mean curvature at a point of rapid deepening of the ocean bottom is interpreted as an indication that the Florida Current is injected into the upper layer above and into the main thermocline and may only contact the bottom intermittently. Richardson (1977) found that the Gulf Stream did not extend to the bottom off Cape Hatteras for direct current measurements from six moorings over periods ranging from 5 to 55 days. Other measurements (Barrett, 1965; Richardson and Knauss, 1971) show northeast flow near the bottom under the axis of the Gulf Stream. After leaving Cape Hatteras, the Gulf Stream gradually develops meanders, clearly visible in the infrared image in figure 4.7. The meanders become progressively larger downstream (Hansen, 1970), but are especially marked after crossing the New England Seamounts (Fuglister, 1963; Warren, 1963). The most intense horizontal thermal and density structures are found between Cape Hatteras and the Seamounts. Strong horizontal density gradients are found throughout the entire depth. These

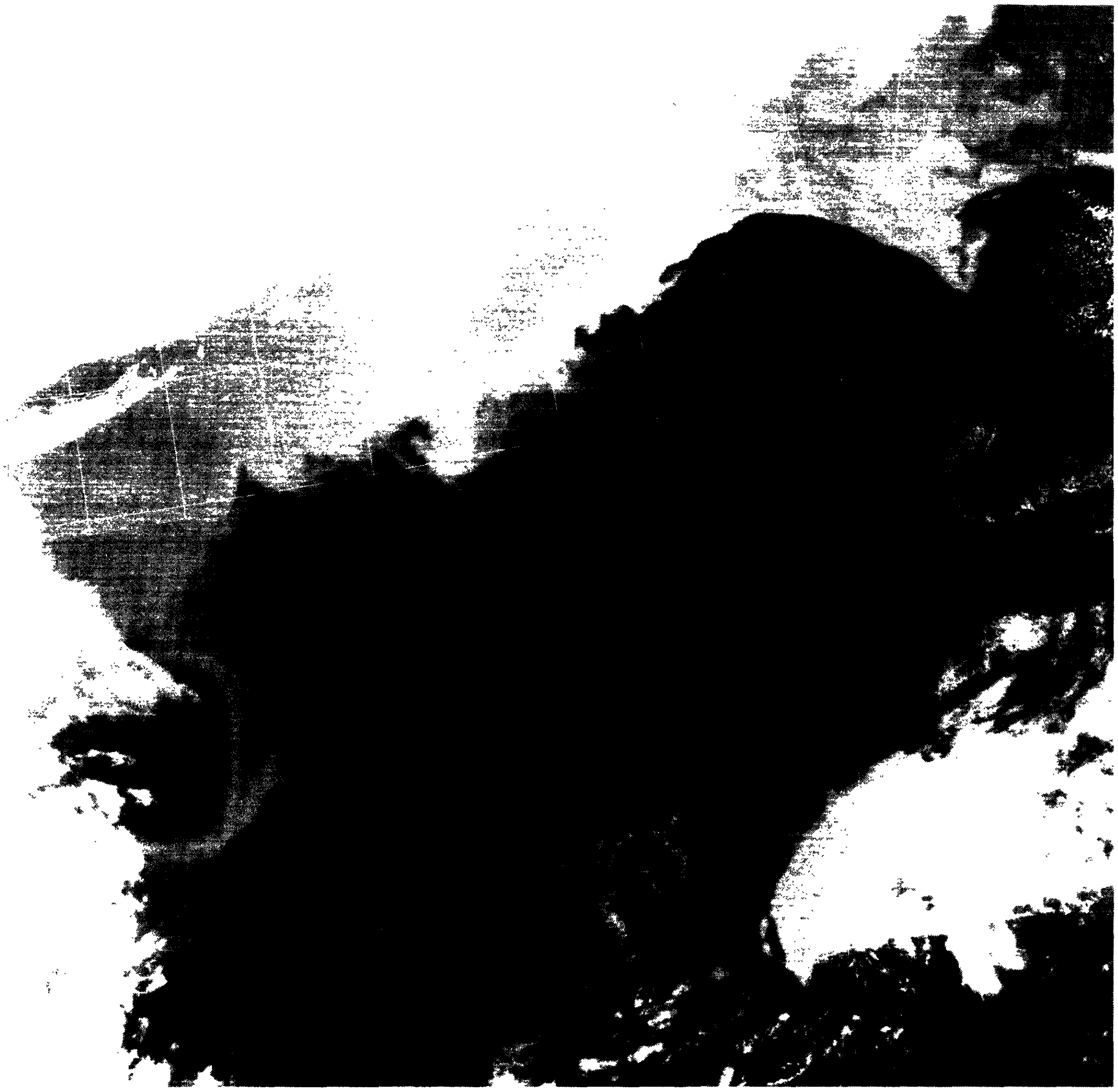

Figure 4.7 The Gulf Stream south of Cape Cod, showing well-developed meanders with several eddies formed in the slope water to the north (Courtesy of R. Legeckis, NOAA-NESS, from NOAA-4 satellite November 12, 1975.)

gradients begin to weaken in the deep water after crossing the Seamount chain (Fuglister, 1963). The meanders become sufficiently large to form detached cold eddies (rings) to the south as shown in figure 4.8 (Fuglister, 1972; Parker, 1971) and warm eddies to the north (Saunders, 1971) of the Gulf Stream at irregular intervals. The meander amplitudes probably do not continue to grow eastward because of the confining effects of the Grand Banks and Southeast Newfoundland Rise. These topographic features appear to lock the Gulf Stream into quasi-stationary spatial patterns similar to those described by Worthington (1962) and Mann (1967) that are more constrained than the meanders farther to the west.

4.4.1 Gulf Stream Separation Mechanisms

The mechanism of separation of the Gulf Stream from the continental slope at Cape Hatteras remains ambiguous. The early theories of mean ocean circulation developed by Stommel (1948) and Munk (1950) required an intensifying current along the western boundary only to the latitude of maximum wind-stress curl. Poleward of the maximum, the current weakened and returned into the ocean interior as a broad slow flow specified by the meridional scale of the wind-stress curl field. It was soon recognized that the lack of even qualitative agreement with the Gulf Stream could be attributed to the neglect of nonlinear terms in the western boundary. Munk, Groves, and Carrier (1950) showed by a perturbation analysis that the nonlinear terms acted to shift the point of maximum velocity downstream past the maximum in the stress curl.

The inertial models that were developed subsequently indicated that an intensifying current with westward flow from the interior (Charney, 1955b; Morgan, 1956) could be extended well past the latitude of maximum curl of the wind stress by inertial recirculation. In two-layer inertial models, the northward extent is limited by surfacing of the inshore isopycnal and ending of the potential to kinetic-energy conversion in the boundary current [Veronis (1973a) and chapter 5]. By increasing the size of the recirculation gyre, the boundary current can be extended to the latitude with zero wind-stress curl that separates the major ocean gyres (Munk, 1950). Leetmaa and Bunker (1978) recomputed the curl of wind stress from recent data and found that the zero stress-curl contour lies near Cape Hatteras on the average. Thus the separation may be a consequence of the larger-scale mean wind field rather than the local dynamics at Cape Hatteras. Moreover, Stommel, Niiler, and Anati (1978) point out that all of the transport in excess of about 30–36 $\times$ 10^6 $m^3\,s^{-1}$ required by the wind-stress-curl distribution can be attributed to recirculation without violating conservation of mass and heat. The possibility that the

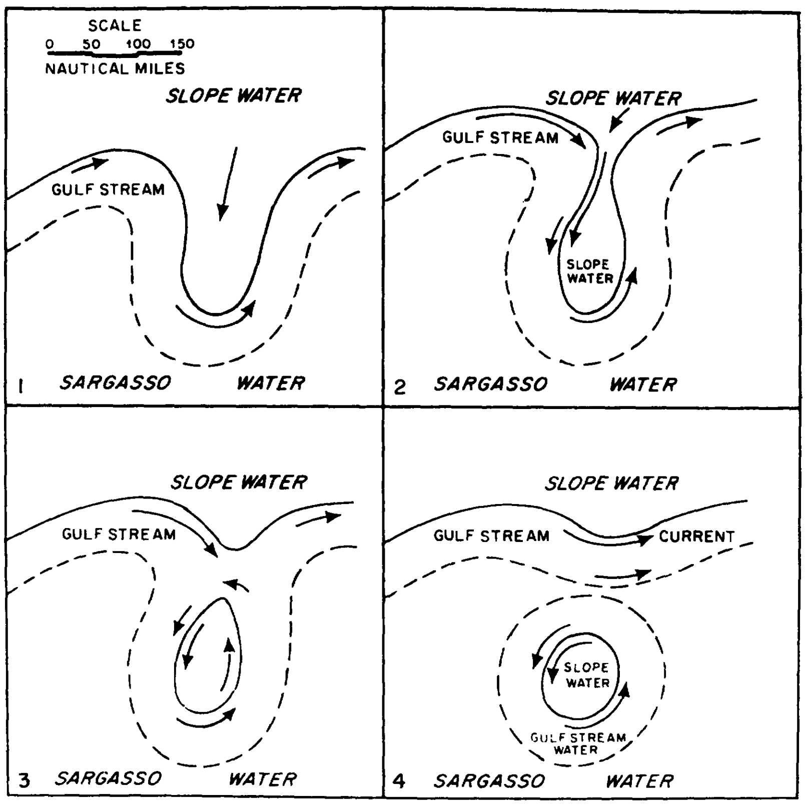

Figure 4.8 Diagram of Gulf Stream ring generation from meander formation to separation. (Parker, 1971.)

mean Gulf Stream path may be determined by the mean-wind pattern deserves further study. It is not obvious why the Gulf Stream should separate from the continental slope to allow formation of the slope-water region extending from Cape Hatteras to the Grand Banks, rather than flow along the slope to the Banks and beyond (for further discussion, see chapter 5).

4.4.2 Gulf Stream Trajectory Models

Different mechanisms for determining the shape of the Gulf Stream have been proposed in terms of path or trajectory models. These will be examined briefly.

The systematic measurements in a series of 11 meridional hydrographic sections through the Slope Water and Gulf Stream carried out in 1960 under Fuglister's guidance (Fuglister, 1963) provided the basis for a number of theoretical studies in the following years, including the development of the trajectory or path theories. The observations during the 3-month duration of GULF STREAM '60 showed rather small changes of the large meander found in the Stream path. Fuglister stated that no evidence was found in the data for shifts in the meanders by more than the Gulf Stream width. Moreover, neutrally buoyant floats tracked at 2000 m depth to provide reference velocities for the computation of geostrophic currents yielded currents extending to the ocean bottom flowing in the same direction as the surface Gulf Stream. These characteristics of the Gulf Stream prompted Warren (1963) to develop a model based on bulk or integrated properties of the Gulf Stream. By assuming that the Gulf Stream could be treated as a narrow current or jet and integrating between streamlines over a section across the current, the vorticity equation was converted to a form relating path curvature to vortex stretching by the changing depth along the path of the current and by changes in the Coriolis parameter resulting from a change of latitude. Given initial conditions of position and direction, as well as the bulk properties of volume transport, momentum transport, and volume transport per unit depth, the subsequent path is determined by the topography and change of latitude encountered enroute. The simple model applied to five observed paths exhibited remarkable agreement in shape in the region of longitude between 65 and 73°W. The path computation could not be continued eastward because of the obvious breakdown of the model in describing meanders over the New England Seamounts. Warren noted, as did Fuglister (1963), that the New England Seamount Arc underlies the region where large meanders develop.

The model possessed several attractive features. The separation from the continental shelf at Cape Hatteras occurred as a natural consequence of the topography and was not related to a wind-stress mechanism as suggested by earlier theories. Moreover, the meanders could develop as a consequence of the initial angle of injection relative to the topography and would not necessarily indicate an instability of the current. Subsequent elaboration of the theory by Niiler and Robinson (1967) brought to light several shortcomings of the approach. The narrow-jet trajectory theory assumed a steady state, whereas later observations revealed the Gulf Stream to exhibit strong time dependence in its meanders. Neither the simple model studied by Warren nor the more elaborate models developed later could be fitted simultaneously to the mean-path and meander data (Robinson, 1971). Robinson concluded that "vortex-line stretching will undoubtedly play some role in the vorticity balance" in a properly posed nonlinear time-dependent theory of meanders. Hansen (1970) obtained a series of measurements of Gulf Stream paths to describe the occurrence and progressive development of meanders in an effort to discriminate between the inertial-jet theories and dynamic-wave models with possible unstable modes that can extract energy from the basic flow. The paths were mapped over a period of a year by towing a temperature sensor along the 15°C isotherm at 200 m depth supplemented as necessary by bathythermograph observations. Hansen concluded that although no model then available could account for all of the major features of the Gulf Stream, the most likely models would have to include topographic influences that are clearly seen in some, but not all, observed paths as well as energy conversion processes such as baroclinic instability necessary to account for meander development at least, where topography is too weak to influence the path. Path models alone were not adequate to account for the meanders.

Time-dependent extensions of the path model have been given by Luyten and Robinson (1974) and Robinson, Luyten, and Flierl (1975). A consistent dynamic quasi-geostrophic model was developed in which the velocity field is resolved into a jet velocity, a velocity of the jet axis, and a transient adjustment velocity assumed small relative to the geostrophic velocities. The model was applied to Gulf Stream data collected during 1969 near 70°W (Robinson, Luyten, and Fuglister, 1974). Using parameters appropriate to the Gulf Stream for 70°W, Robinson et al. (1975) found that an inlet period of 31 days had a spatial wavelength of 560 km and a downstream growth (e-folding) scale of 200 km, in agreement with the observed large-scale meanders of the Gulf Stream. In the local vorticity balance, advection and transient terms dominated the topographic and β-effects. The model contains mechanisms analogous to ring or eddy formation. Because the path displacements are not constrained to be small, the path equations can, at least in the case of the purely baroclinic limit and no β-effect, yield solutions in which the meanders grow spatially and close upon themselves to form isolated eddies. For the thin jet models to be applicable, however, the transient flows must be small.

4.4.3 Deep Currents of the Gulf Stream

The vertical coherence of the thermal and density field within the Gulf Stream and the slow evolution of meanders noted by Fuglister (1963) made plausible the assumption that the Gulf Stream extended to, and interacted with, the ocean bottom. A vertically coherent Gulf Stream, however, was not substantiated by subsequent direct current measurements. These showed a vigorous velocity field in the deep water (Schmitz, Robinson, and Fuglister, 1970) that was not a simple downward extension of the near-surface flow.

Luyten (1977) designed a current-meter array consisting of 15 moorings deployed in two meridional lines about 50 km apart at 70°W to measure the deep flow on the continental rise and beneath the Gulf Stream. Current meters were placed 1000 m above the ocean bottom on each mooring with a second instrument 200 m above bottom on some moorings as shown in figure 4.9. Data for 240 days were obtained on 30 of the 32 current meters. The measurements revealed a remarkable and unexpected feature of the deep flow in that the downstream coherence scale was very small (less than 50 km) while the meridional or cross-stream scale was found to be several times as large ($\approx$150 km).

The currents contained strong meridional bursts of speeds reaching 40 cm s;$^{-1}$ and lasting for about a month. The highest intensity of flow occurred nearest the Gulf Stream, with four to six bursts in individual records. Hansen (1970) calculated (figure 4.10) an average eastward phase speed of 8 cm s^{-1} and a mean wavelength of 320 km for meanders of the 15°C isotherm at 200 m depth. This corresponds to a period of 46 days, agreeing approximately with the interval between events (four to six events in 240 days) in Luyten's array shown in figure 4.9. Because the measurements were not simultaneous, a correlation between the motion in the deep water and the upper levels has not been firmly established. The agreement of time scales is suggestive of a strong coupling between them.

The mean flow for depths shallower than 4000 m on the continental rise is westward with speeds 2–5 cm s^{-1} and apparently dominated by the bottom slope. Deeper than 4000 m, the mean flow vectors tend to have a small eastward component with an erratic or perhaps rapid spatial variation of the meridional component, reflecting the burst structure of the variable flow. The Gulf Stream, if it exists near the bottom above 70°W as suggested by Fuglister (1963) and Robinson, et al. (1974), is nearly completely masked by the strong deep meridional eddy field. The lack of coherence downstream is puzzling. It may be a consequence of having only two points of measurement along the Gulf Stream direction. The deep fluctuations may be locked in some manner by local topography, resulting in an inhomogeneous spatial pattern of coherence. Luyten found the interaction between eddies and mean flow to be small north of the Gulf Stream on the continental rise. The horizontal eddy-stress divergence vector was nearly perpendicular to the mean flow. Under the Gulf Stream itself, the mean flow appeared to be gaining energy from the eddies (figure 4.11).

4.4.4 Dynamical Gulf Stream Models

The instability and subsequent evolution of a thin zonal jet in the upper layer of a two-layer model described by Rhines (1977) has several points of similarity with the observations taken by Luyten (1977) and Schmitz (1978) of the velocity field near the Gulf Stream. Rhines investigated numerically the evolution of an eastward jet imbedded in a broad westward flow in the upper layer of a two-layer ocean model. The initial field was perturbed by broadband noise. The sequence of development is shown in figure 4.12. In less than 20 days, organized meanders become visible in the upper layer and elliptical eddies with predominantly north–south motion in the lower layer resembling the meridional motions observed by Luyten (1977). In the early stages, the pattern moves downstream with the motion in the deep layer leading the upper layer as required by baroclinic instability. After the meander exceeds unit steepness (about 42 days), the nonlinear eddy–eddy interactions become evident, causing distortion and stretching of vorticity contours by horizontal velocity shear. In the lower layer, eddies of like sign begin to coalesce, producing a north–south displacement and creating an abyssal flow in the same direction as the upper layer jet.

As the eddy field develops, the jet is no longer recognizable in the velocity or streamfunction field, but is still visible in the topography of the interfacial surface. At later stages (t = 62 days), the eddies in the upper and lower layer begin to lock together and become more barotropic in structure. Rhines suggests, as illustrated in figure 4.12, that the time evolution of the model resembles the downstream development of the Gulf Stream with the injection of the jet into the upper layers of the ocean corresponding to initial conditions in the model. Downstream migration of the eddies in the lower layer, however, was not observed by Luyten (1977). Instead, the phase propagation appeared to be southward, probably dominated by the bottom slope. The downstream motion of the surface meander has been documented by several authors, notably Hansen (1970), as seen in figure 4.10. Coupling of the surface meanders and eddies to the lower layer drives a mean flow below the main thermocline and the development of barotropic eddies. Schmitz (1977, 1978) has found an eastward deep flow and increased barotropic signature

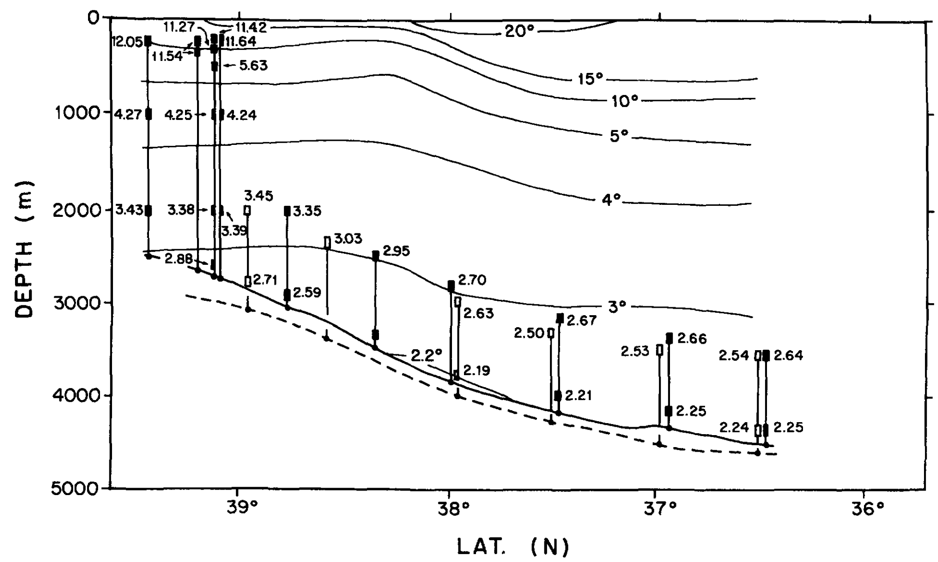

Figure 4.9A Distribution of current meters and temperature-pressure recorders on the Luyten (1977) Continental Rise array. The solid lines refer to 70°W, dashed to 69°30′W. (Luyten, 1977.)

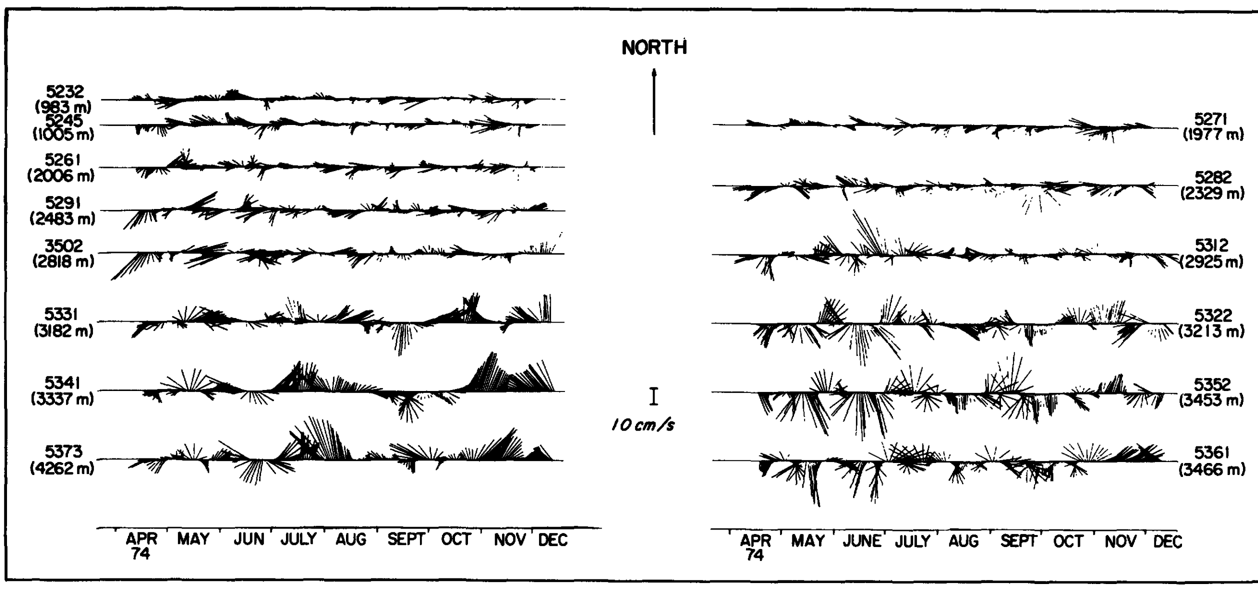

Figure 4.9B Time sequence of 1-day averaged current. The numbers identify mooring and instrument. Instrument depths in meters are shown in parentheses. (Luyten, 1977.)

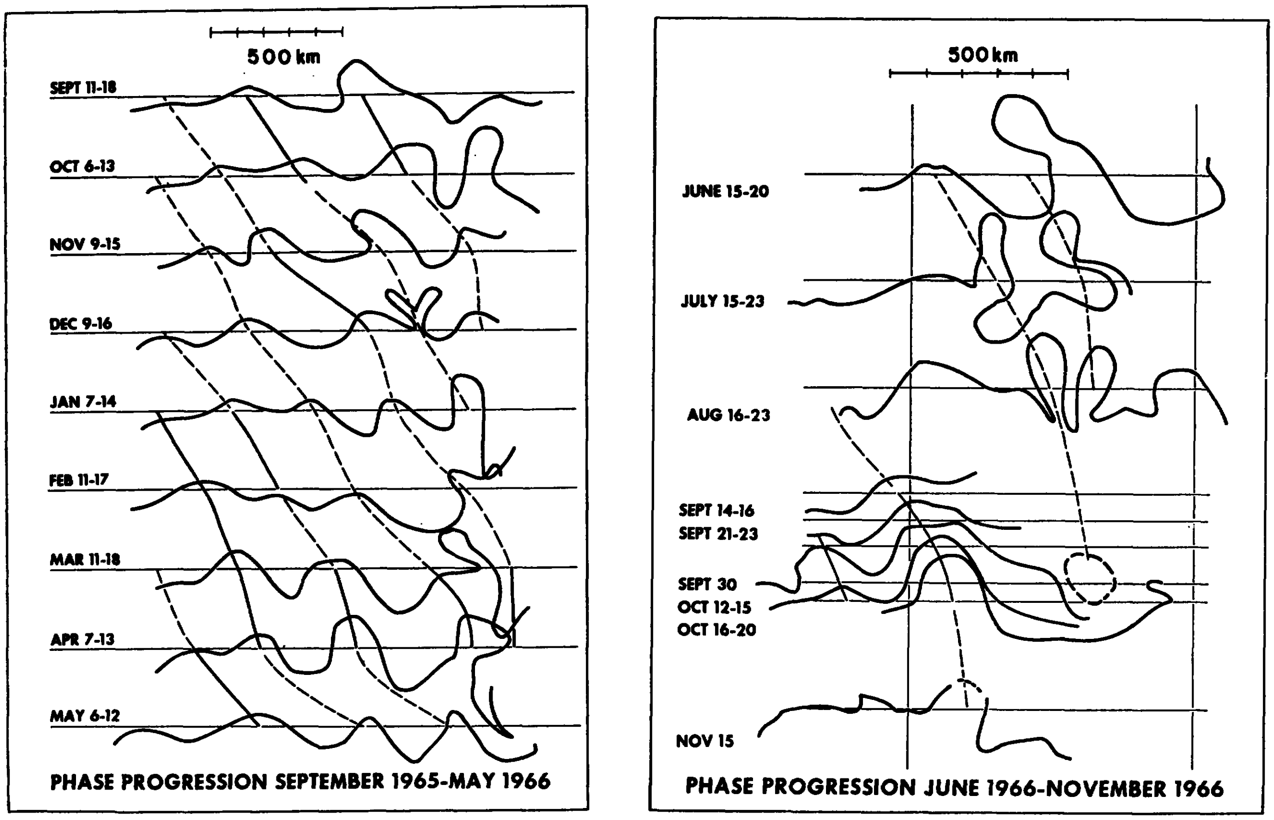

Figure 4.10 Inferred progression and evolution of meanders relative to the mean Gulf Stream path. The diagonal lines show phase propagation (solid where supported by other evidence). (Hansen, 1970.)

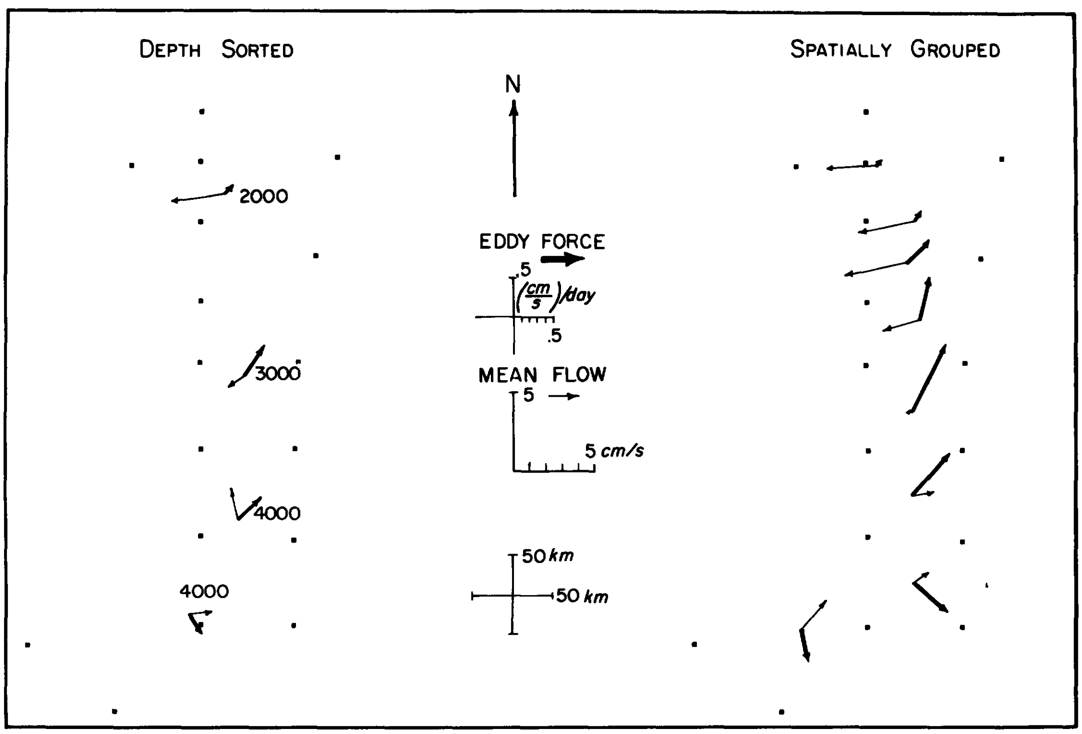

Figure 4.11 Vector acceleration of mean flow by eddy Reynolds stress gradients. The "eddy forces" tend to oppose the mean flow on the continental rise and to accelerate it under the Gulf Stream at the southern portion of the array. The eddy forces are estimated by grouping the data in depth intervals and from neighboring values in the array at a common depth for the spatial grouping. (Luyten, 1977.)

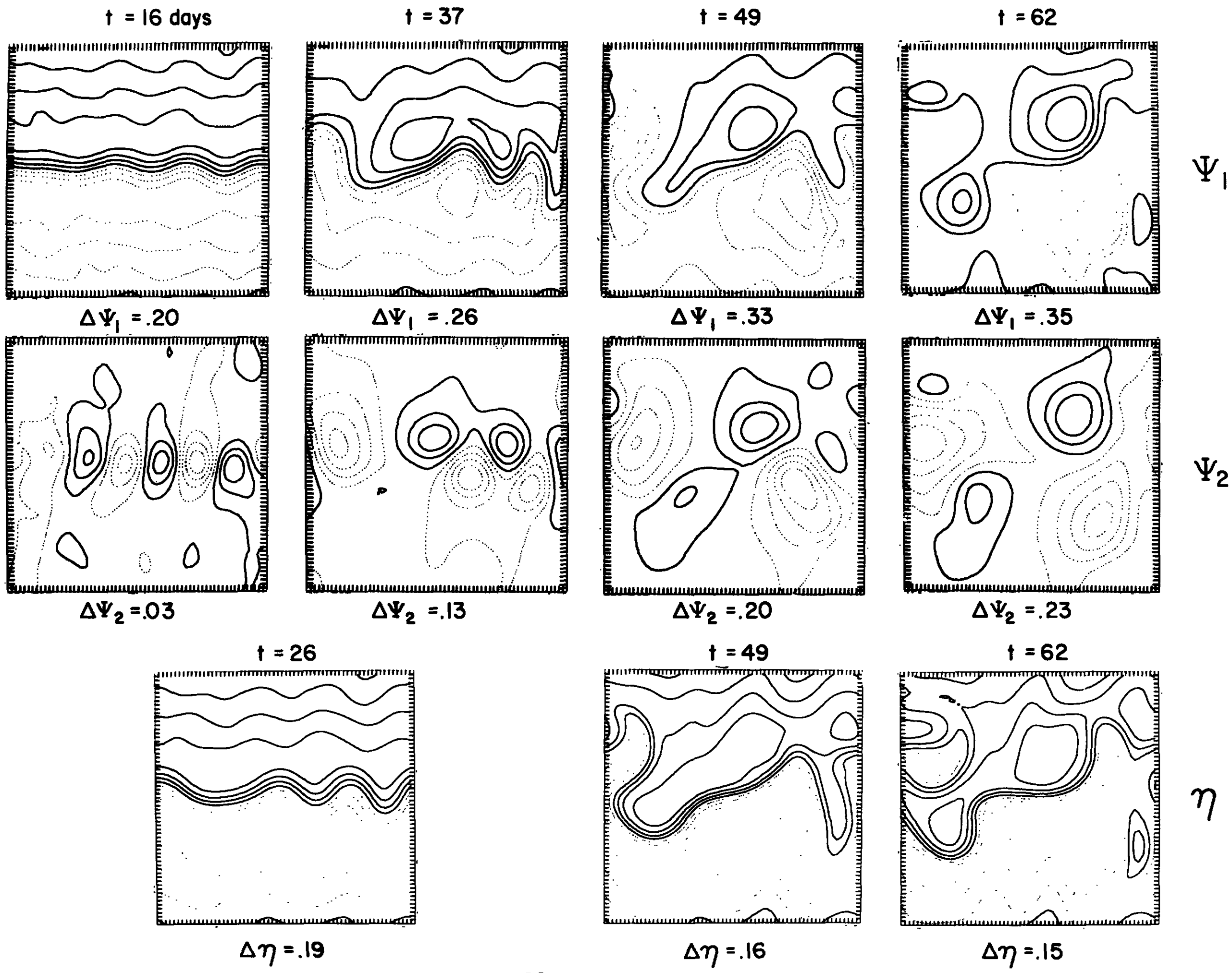

Figure 12

Figure 4.12 Evolution in time of a two-layer spatially periodic model. The figure shows upper-layer streamfunction ψ_1, lower-layer streamfunction ψ_2, and interface height η. The interface height retains the strong slope across the initial jet long after the jet has been obscured by barotropic flows in the velocity field. Contour intervals are indicated for relative comparison. Dimensions are 1250 km on each side with a speed of 50 cm s^{-1} averaged across the jet. (Rhines, 1977.)

to the velocity field (figure 4.13) in a long-term moored-array experiment along 55°W designed to examine the mesoscale eddy field in the vicinity of the Gulf Stream. [The array is part of a long-range program developed under the cooperative experiments MODE carried out in 1973 and in POLYMODE (1974-1979). The reader is referred to chapter 11 for a discussion of mesoscale eddies in the ocean.]

Although it is evolving and not stationary in time, Rhines's model is attractive and suggestive of processes that may be acting in the actual Gulf Stream. However, the model lacks an explicit recirculation mechanism and has no bottom topography. Both may affect the behavior significantly and may need to be incorporated as essential elements in a more complete Gulf Stream model. The simpler models must be explored and understood before the combined effects of several

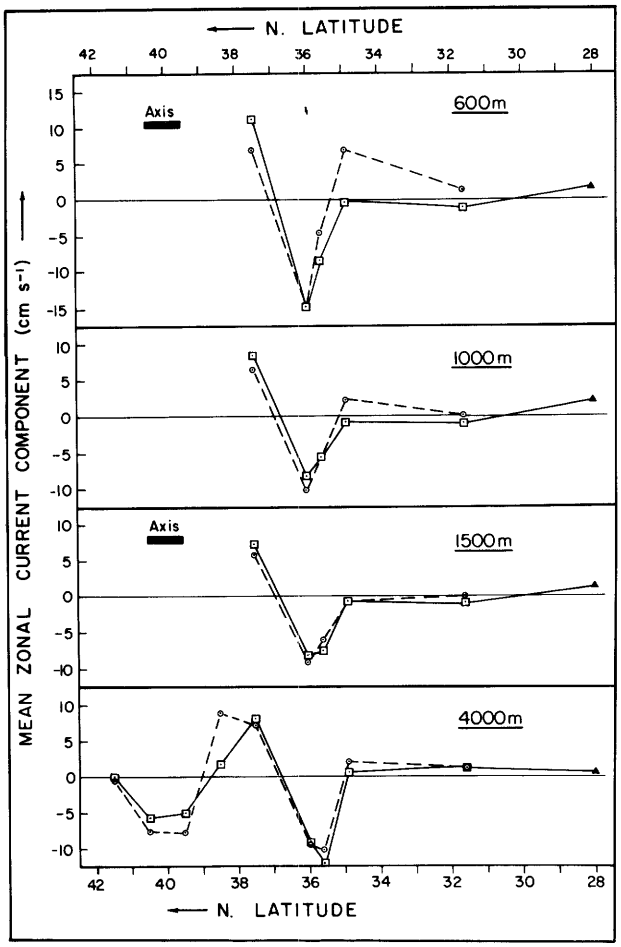

Figure 4.13 Meridional distribution of time-averaged zonal current at four depths along 55°W from the first POLYMODE array shown by circles ○ and from the combined first and second setting of the array shown by squares □. The approximate mean axis of the Gulf Stream is indicated. The westward mean flow shows little depth dependence. The mean flow is eastward under the Gulf Stream at 4000 m depth. (Schmitz, 1978.)

mechanisms can be interpreted in the more complex general circulation models.

The mass flux of the Gulf Stream has been estimated to be in the range of 100-150 × 10^6 m^{3-1} south of Cape Cod (Fuglister, 1963; Warren and Volkmann, 1968; Knauss, 1969). Stommel et al. (1978) concluded that most of the transport in the Gulf Stream is recirculated in the western North Atlantic. The net transport northeast, assumed to be wind driven in the ocean interior, is only 38 × 10^6 $m^3 s^{-1}$. Thus the recirculating portion may be as much as 60-110 × 10^6 $m^3 s^{-1}$. The high recirculation rate implies that a relatively small fraction of the energy is lost by the Gulf Stream in flowing between Cape Hatteras and the Grand Banks. Most of the kinetic energy is converted back to potential energy to form the recirculation to the south. Just as the conversion from potential to kinetic energy requires an accelerating pressure gradient along the western boundary, the conversion from kinetic to potential energy requires an opposing or decelerating pressure gradient. The details of the conversion process must be complicated because of the large-amplitude meandering and eddy formation that takes place between Cape Hatteras and the Grand Banks. Because it can be assumed that the recirculation is relatively stationary and does not change its size and intensity rapidly, its energy content is essentially constant. Energy is converted from potential to kinetic in flowing through the Gulf Stream and is converted back to potential energy on entering the westward gyre. As the recirculation transport is larger (perhaps twice) than the wind-driven transport, the kinetic-energy flux in the Gulf Stream probably contains only a minor contribution, depending on the velocity profile, attributable to direct forcing, and only this amount of energy needs to be dissipated to maintain a steady state. The magnitudes involved can be estimated very roughly from available data. The average wind-stress magnitude over the subtropical North Atlantic is about 0.5 dyn cm^{-1} (Leetmaa and Bunker, 1978). Assuming that the curl of the wind stress drives a transport of 40 × 10^6 $m^3 s^{-1}$ into the western boundary current through a section of average depth 500 m deep and 2000 km in length, the average inflow current would be 4 $cm\,s^{-1}$. Assuming that 4 $cm\,s^{-1}$ is also typical of interior geostrophic velocities and that wind stress-eddy forcing is small, the work done by wind stress would be $|\tau| \cdot |\mathbf{v}| \approx 2$ $erg\,cm^{-2} s^{-1}$. Over the entire basin (area ≈ 1 × 10^{17} cm^2), the rate of energy input by the wind is 2 × 10^{17} $ergs\,s^{-1}$; it is converted entirely to potential energy. The interior flux of kinetic energy by the mean flow is negligible. No net work is done by Ekman velocities in the frictional layer because the work by surface stresses must be dissipated within the Ekman layer for a steady state to exist. Only the surface geostrophic currents need be considered in estimating net wind work (Stern, 1975a).

If $\boldsymbol{\tau}$ is wind stress and

$$\mathbf{v}_g = \frac{1}{\rho f}(\mathbf{k} \times \nabla_H p)$$

the surface geostrophic current, the net rate of wind work on the ocean surface is

$$\mathbf{v}_g \cdot \boldsymbol{\tau} = \frac{1}{\rho f}(\mathbf{k} \times \nabla_H p) \cdot \boldsymbol{\tau}$$

$$= -\frac{1}{\rho f}(\mathbf{k} \times \boldsymbol{\tau}) \cdot \nabla_H p$$

$$= \frac{1}{\rho}\mathbf{V}_e \cdot \nabla_H p$$

where $\mathbf{V}_e = -1/f\,(\mathbf{k} \times \boldsymbol{\tau})$ is the Ekman mass transport, $\nabla_H p$ the horizontal pressure gradient, and ρ density of the surface layer. The net work by wind stress can be interpreted as the rate at which mass is transported up the pressure gradient in the Ekman layer. Because pressure gradients are produced hydrostatically, the "uphill" Ekman flow represents an increase of potential energy at a rate equal to the wind work. Thus, except for the portion dissipated in the Ekman layer, the wind work is converted entirely to potential energy in the ocean interior. If no interior dissipation is present, all of the input wind energy must be removed through the western boundary current and dissipated within the recirculation region. Assuming a width of the boundary current of 80 km and a depth of 500 m, the energy and mass fluxes require a mean speed of 100 $\mathrm{cm\,s^{-1}}$ in the western boundary current. These are joined near Cape Hatteras by recirculation fluxes of mass and energy, resulting in a deepening and intensification of the flow, but leaving the scale width of the Gulf Stream unchanged. Suppose the recirculating transport is $80 \times 10^6\ \mathrm{m^3\,s^{-1}}$, making a total transport of $120 \times 10^6\ \mathrm{m^3\,s^{-1}}$. The total kinetic energy flux would be three times the interior value or $6 \times 10^{17}\ \mathrm{ergs\,s^{-1}}$ if the section deepened with no change of mean velocity. Only $2 \times 10^{17}\ \mathrm{ergs\,s^{-1}}$ must be dissipated and $40 \times 10^6\ \mathrm{m^3\,s^{-1}}$ transport returned to the interior circulation to maintain a steady mean state. The kinetic-energy fluxes into the western boundary layer and out of the dissipation region are assumed to be negligible compared to conversion rates of potential energy to kinetic energy. The kinetic-energy flux is proportional to the cube of speed and is, therefore, very sensitive to the velocity profile in the intensified Gulf Stream. For a given mean flux, the kinetic-energy flux is least if the flow is spread throughout the entire water column. A more detailed calculation is required to improve the estimate. An essential point of the argument is that the conversion of kinetic energy to and from potential energy is an exchange between mean fields. These conversions are not usually examined in numerical models because their basin average is zero.

An approximate estimate of the rate of energy dissipation in the Gulf Stream can be obtained from the initial decay rate of Gulf Stream rings, as these are formed from segments of the Gulf Stream itself (Fuglister, 1972). Cheney and Richardson (1976) found decay rates in Gulf Stream rings of $1.7 \pm 0.2 \times 10^{21}\ \mathrm{ergs\,day^{-1}}$, or about 37 $\mathrm{ergs\,cm^{-2}\,s^{-1}}$ averaged over the ring area, from the observed decrease of available potential energy of the ring. If the same dissipation rate is assumed in the Gulf Stream, it would require over 7000 km of path or about 80 days to dissipate the kinetic energy transported in the boundary region from the ocean interior. The distance from Cape Hatteras to the Grand Banks is about 2300 km—too short to get rid of the kinetic energy by internal dissipation and radiation. Meandering will increase the Gulf Stream path significantly, perhaps by a factor of two or more. Some of the energy is removed into the rings and eddies.

The flux of mass and energy in the Gulf Stream is sufficient to form a cyclonic ring of 100-km radius per week or about 50 rings per year. Many fewer are believed to form. Fuglister (1972) and Lai and Richardson (1977) estimated that as many as 10 to 16 rings of either anticyclonic or cyclonic type may form annually north and south of the Gulf Stream. If so, the formation of rings, although a spectacular manifestation of the Gulf Stream decay mechanism, is not the major mechanism for dissipating the kinetic energy. Some, possibly a considerable fraction, of the kinetic energy is transferred to the lower layers to replace the energy loss in the recirculation gyre through instability, radiation, and dissipation from the gyre, as indicated in some numerical models or by other dispersive processes within the gyre. The bulk of the kinetic-energy flux, however, appears to be converted back into potential energy.

4.4.5 Numerical Gulf Stream Models

The complete energetics of the Gulf Stream are far from obvious. The development of numerical models with sufficient spatial resolution to permit significant eddy-mean flow and eddy-eddy interactions to take place has revealed energy conversion and dissipation modes to help interpret Gulf Stream behavior.

The first series of ocean-scale general circulation models to exhibit baroclinic instability in the western boundary current and recirculation gyre were described by Holland and Lin (1975a,b). They found that if lateral friction was taken sufficiently small or wind forcing sufficiently strong, a steady-state flow was not attained in the numerical integration. The flow remained time dependent but statistically stationary, in that means and variances approached constant values with time.

Thus, the time-varying eddy flow appeared to be an essential part of the momentum-transfer mechanism in the model. Their model was chosen with a "single-gyre" wind-stress distribution so that the western boundary current was constrained by both western and northern boundaries and did not exhibit strong instabilities. The next step was taken by Semtner and Mintz (1977), who developed a five-level primitive equation model with shelf topography and surface heat exchange to simulate the Gulf Stream and mesoscale eddies in the western North Atlantic. Their model contained two novel features not included in the Holland and Lin model: a biharmonic friction to prevent a "violet catastrophe" resulting from a transfer of mean square vorticity (enstrophy) to high wavenumbers (enstrophy cascade), and a bottom frictional Ekman layer to allow dissipation at the ocean bottom. These mechanisms had been introduced and explored earlier by Bretherton and Karweit (1975) and Owens and Bretherton (1978) in the study of open-ocean mesoscale eddy models. The primitive equation model showed that the dominant instability occurred within the simulated Gulf Stream over the continental rise. Over the flat abyssal plain, energy was transferred from the eddies to the mean flow.

Reduction of the effective lateral friction using the biharmonic dissipation allowed the eastward jet (Gulf Stream) to develop intense meanders, some of which formed ringlike eddies that separated from the jet and drifted westward in the recirculation gyres. Strong deep gyres developed in the vicinity of the meandering eastward jet as a consequence of downward flux of momentum associated with the meanders. The energetics of the primitive equation model were studied in detail by Robinson et al. (1977) to evaluate the types and rates of energy transfers in several regions of the basin. The primitive equation models are expensive to run and Semtner and Holland (1978) concluded after comparison that most of the behavior of the western boundary current and the free eastward jet used to simulate the Gulf Stream is contained in the simpler quasi-geostrophic two-layer model. Holland (1978) carried out a series of experiments using this latter model to explore the effects of horizontal diffusion and bottom friction on energy flow to the eddy field. For low lateral diffusion, most of the energy input by wind stress was transferred via upper-layer eddies to eddies in the lower layer, to be dissipated by bottom friction. For high lateral viscosity, the energy is dissipated by diffusion in the mean flow, with much smaller fractions being transferred to eddies in the upper and lower layers. An ocean basin with a "double-gyre" wind forcing produced a free mid-ocean jet with behavior recognizably closer to that of the Gulf Stream. Momentum and energy were transferred to the lower layer through the coupling of eddies across the interface, to be dissipated by friction at the bottom. A strong recirculating gyre is developed in the lower layer as seen in the streamfunction and interface topography and total transports shown in figure 4.14. The results are sufficiently promising to attempt a comparison (figure 4.15) with current measurements along 55°W by Schmitz (1977). Good agreement was obtained with the mean zonal currents in deep water. However, the north-south variations of model eddy kinetic energy $\overline{u'^2 + v'^2}$ and Reynolds-stress term $\overline{u' v'}$ were considerably broader than measured by Schmitz. Holland attributed the broader distribution of eddy variables to the absence of bottom topography in the model. A similar comparison for the Semtner-Mintz model was given by Robinson et al. (1977). Further comparisons with measurements are necessary to define the limits of applicability and to locate the parameter ranges for best fit of the numerical models to the Gulf Stream. The initial results are encouraging.

4.5 The North Atlantic Current

The Gulf Stream undergoes a radical change on reaching the Southeast Newfoundland Rise, a ridge running toward the Mid-Atlantic Ridge from the Tail of the Grand Banks. According to Iselin (1936), the Gulf Stream divides, with the major branch flowing north across the Southeast Newfoundland Rise parallel and opposite to the cold Labrador Current running south along the eastern face of the Grand Banks. Another part of the Gulf Stream continues eastward to divide further into northward and southward branches before crossing the Mid-Atlantic Ridge. The southern branch recirculates into the Sargasso Sea. The northern branch turns eastward after reaching a latitude of about 50°N to become the North Atlantic Current. Details of the separation into branches were not available to Iselin. Even today little is known of the time-varying features of this critical-breakdown region of the Gulf Stream. The sharp thermal contrast between the Labrador Current and the warm waters originating from the Gulf Stream is seen vividly in the infrared images reproduced in figures 4.16 and 4.17. The ribbon of cold (white) Labrador Water marks the eastern wall of the Grand Banks. A narrow strip of cold water (figure 4.17) is entrained along the edge of the northward-turning current, heightening the contrast.

Worthington (1962, 1976) proposed a fundamentally different interpretation of the hydrographic data, in which the Southeast Newfoundland Rise acts to separate the circulation into two independent anticyclonic gyres (figure 4.6). A trough of low pressure near to, and parallel with, the Southeast Newfoundland Rise, assumed to be continuous, marks the boundaries between the gyres. Worthington reached this conclusion

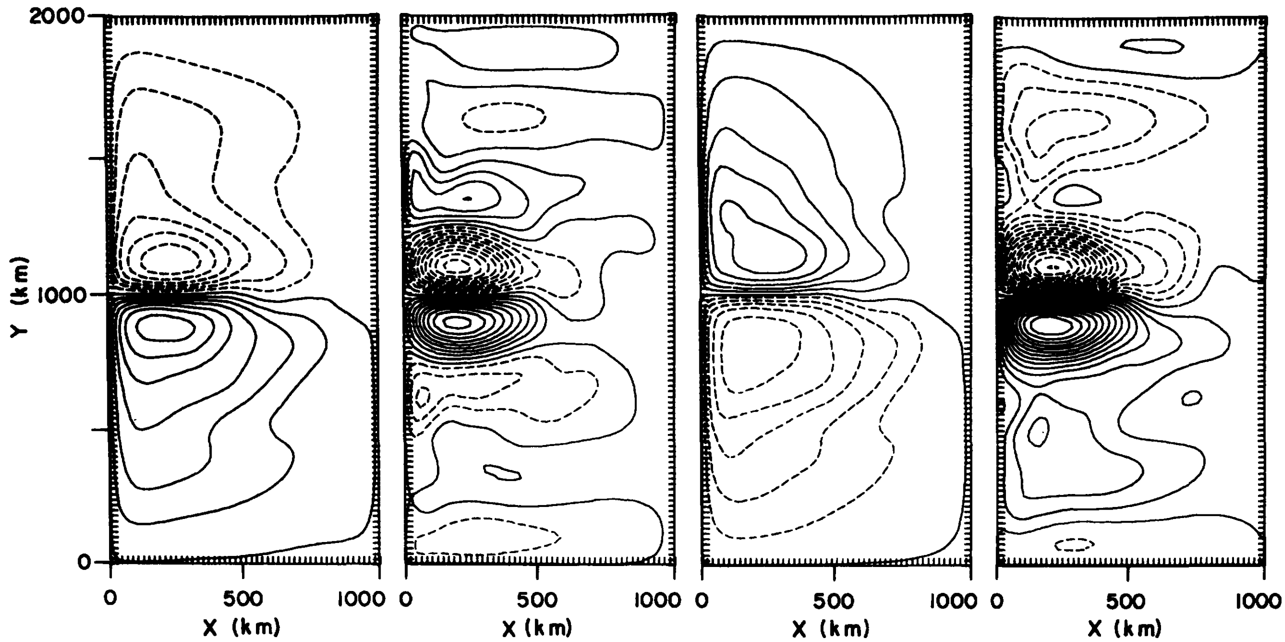

Figure 4.14 The mean fields for Holland experiment 3: (A) upper-layer streamfunction (contour interval CI = 5000 $m^2 s^{-1}$); (B) lower-layer streamfunction (CI = 1000 $m^2 s^{-1}$); (C) interface height (CI = 20 m); (D) total transport (CI = 50 × 10^6 $m^3 s^{-1}$). (Holland, 1978.)

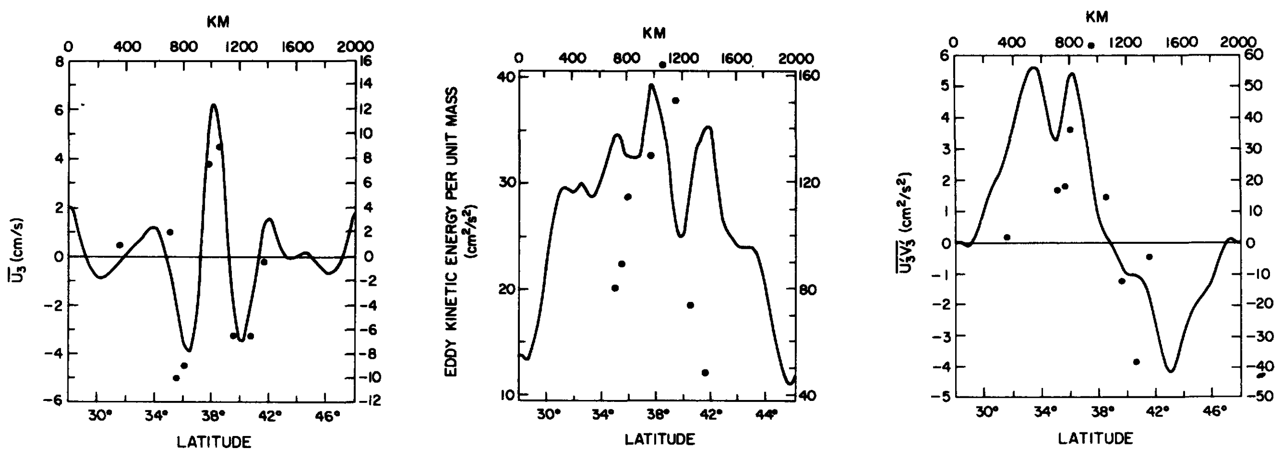

Figure 4.15 Comparison of meridional distributions of (A) zonally averaged mean flow $\bar{u}_3$, (B) eddy kinetic energy $\overline{u_3'^2 + v_3'^2}$, and (C) Reynolds stress $\overline{u_3'v_3'}$ in the lower layer (subscript 3) for experiment 3 compared with observed distribution (Schmitz, 1977) along 55°W. (Holland, 1978.)

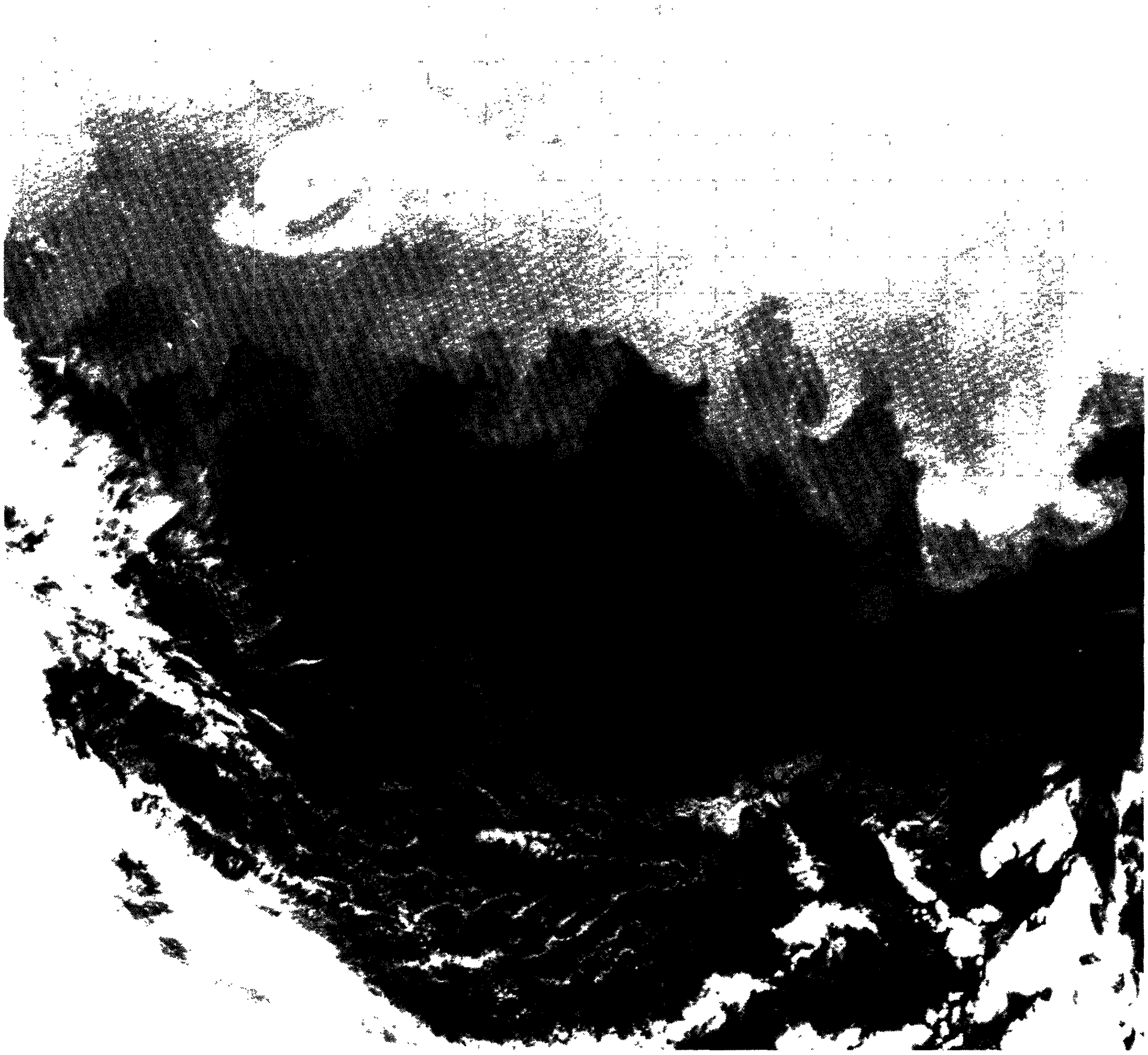

Figure 4.16 The Gulf Stream and North Atlantic Current as seen from an infrared image taken November 2, 1977. The figure shows the colder shelf and slope water responding to a complex meandering of the Gulf Stream. The cold ribbon of Labrador Water is seen flowing south along the eastern edge of the Grand Banks. (Courtesy of R. Legeckis.)

Figure 4.17 A detailed infrared image of the circulation in the vicinity of the Grand Banks showing the southward flowing cold Labrador Water. (Courtesy of P. La Violette, NORDA.)

from the observation that the northern gyre transported water about 1 ml l^{-1} richer in dissolved oxygen than the water of the same temperature and salinity type carried by the Gulf Stream. The source of water of the same density with higher oxygen content is available to the west but it is fresher and colder and would have to be entrained in large amounts to produce the observed concentration of oxygen in the North Atlantic Current. Moreover, a large flow into the Northern Labrador Basin from the Gulf Stream would require a compensating return flow of equal magnitude passing east and south of the Gulf Stream. Worthington argued that the close proximity of the saline Mediterranean Water anomaly is evidence against such a strong return flow from the Northern gyre. Mann (1967), using more recent data obtained on cruises of C.S.S. *Baffin* during April–May 1963 and June–July 1964, disagreed with Worthington's interpretation and proposed a splitting of the current analogous to Iselin's (1936) scheme. Analysis of the hydrographic stations indicated that a branch transporting about 20×10^6 m^3 s^{-1} joined by about 15×10^6 m^3 s^{-1} originating in the slope water formed the branch of the North Atlantic Current to the north. A southward-flowing branch of about 30×10^6 m^3 s^{-1} carried the remainder of the Gulf Stream back into the Sargasso. A weak anticyclonic eddy centered over the Labrador Basin appeared to be a persistent feature of the circulation. Mann suggested that mixing between waters in the eddy and the currents to the west could supply the additional dissolved oxygen observed in the northern basin. A similar interpretation has been constructed using data from subsequent cruises to the area in 1972 (Clarke, Hill, Reiniger, and Warren, 1980). These authors estimate that the oxygen gradient of 2 ml l^{-1} per 100 km across the northward current is sufficient to enrich the waters north of the Southeast Newfoundland Rise by horizontal diffusion along isopycnals given an eddy diffusion coefficient of 10^7 cm^2 s^{-1} acting along the 700-km length of the northward current. They point out also that the separate gyres postulated by Worthington would require significant departures from geostrophic flow. The dynamic topography relative to 2000 db shown in figure 4.18 is consistent with the branching hypothesis described by Mann (1967).

Although the weight of evidence available today seems to favor the general interpretation given by Mann, the definitive answer is not yet in. Evidence from the hydrographic cruises and infrared surface thermal structure indicate strong time dependence in the region. The flow across the Southeast Newfoundland Rise may be intermittent, so that the dynamic topography may show a varying degree of coupling across the Southeast Newfoundland Rise. The interpretation of mean flow across the Southeast Newfoundland Rise can be modified considerably by time dependence. It seems less likely that the interpretation of water-type characteristics will be altered significantly as these are inherently conservative and less affected by time variations than the spatial distributions.

4.6 Summary and Conclusions

The growing literature describing the Gulf Stream System and its dynamics is impressive in its diversity and detail. The author recognizes that the treatment of many of the topics included in this review is superficial and may not reflect accurately all of the accomplishments and directions of current research. The number of papers and their detail and complexity necessitated rather brutal simplification to reduce the length of the review and the time needed for its preparation. Many topics could not be discussed at all. The general circulation of the North Atlantic leading into the Gulf Stream System is bypassed. The reader is referred to Stommel's (1965) monograph on the Gulf Stream, in which much of the classical oceanographic material is summarized, and to Worthington (1976) for his detailed examination of the water masses and their sources and sinks in the North Atlantic (see also chapter 2). The Loop Current in the Gulf of Mexico has been excluded from the Gulf Stream System. Yet the dynamics of the Loop Current may have significant downstream effects. The exclusion seems arbitrary.

The deep western boundary current remains a mystery. How does it coexist with the Gulf Stream and the recirculation gyre? The topic of warm- and cold-core rings has been omitted. Both theory and observations of rings are being pursued vigorously at present and a substantial body of literature has accumulated (Lai and Richardson, 1977; Flierl, 1977, 1979a). Their formation by the Gulf Stream is of obvious importance for removing mass, momentum, and energy from the Gulf Stream and for exchanging Continental Slope Water and Sargasso Water across the Gulf Stream, even though their overall contribution to the kinetic-energy flow from the Gulf Stream may prove to be relatively small compared with the energy loss by other processes.

The transport of heat, salt, and other quantities by the Gulf Stream is omitted. The literature on transport of heat by the Gulf Stream is surprisingly meagre. Heat exchange for the North Atlantic has been estimated by Bunker and Worthington (1976). Newton (1961) concluded that rings represented the principal mechanism for transporting heat between the Sargasso and the Continental Slope Water. Vonder Haar and Oort (1973) estimate that 47% of poleward transport of heat in the northern hemisphere at latitudes 30–35°N is carried by ocean currents such as the Gulf Stream. It is expected

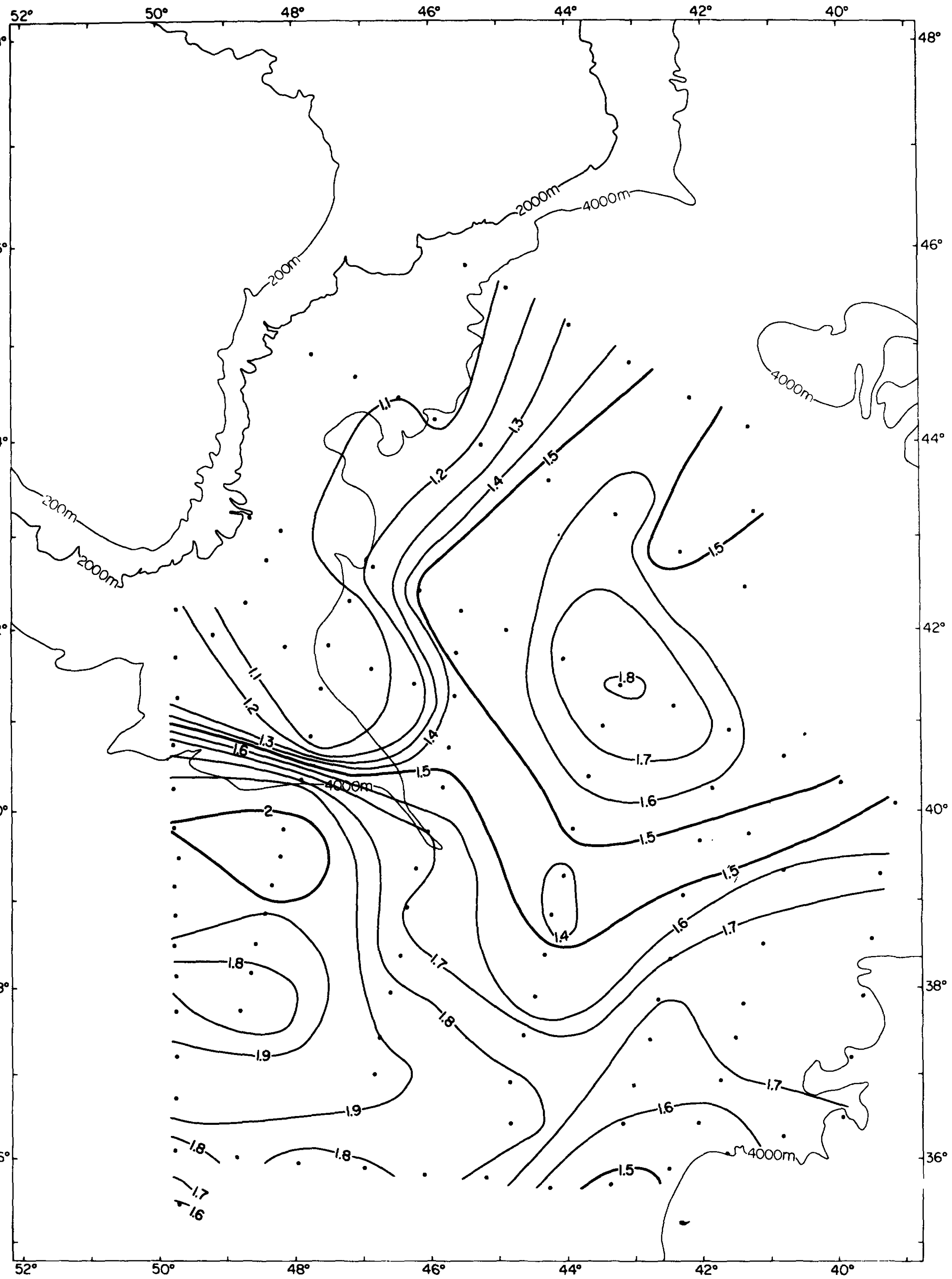

Figure 4.18 Dynamic topography from the North Atlantic Current showing the separation of the Gulf Stream into a northern branch flowing around a weak, anticyclonic eddy in the Labrador Basin. (Clarke et al., 1980.)

that interest in heat transport by the Gulf Stream will grow because of the need to develop more comprehensive and realistic models of world climate.

The Continental Slope Water to the north of the Gulf Stream is not adequately explained. Is the separation of the Gulf Stream from the continental slope at Cape Hatteras a consequence of a local dynamic process related to local topography, as suggested by many of the inertial models (Greenspan, 1963; Pedlosky, 1965a; Veronis, 1973a)? Is the separation a consequence of the large-scale wind-stress pattern? Leetmaa and Bunker (1978) show that the mean curl of the wind stress reverses sign near Cape Hatteras and is zero over a path that is surprisingly like the mean Gulf Stream across the Western North Atlantic. Is the path simply determined by the mean wind field? It is conceivable that the presence of the Gulf Stream with its strong lateral thermal contrast may significantly affect the wind-stress gradients in its vicinity. The position of the line of zero curl of the wind stress may be a consequence, as well as a cause, of the observed Gulf Stream location. Another possibility is an upstream influence of the Grand Banks jutting southward into the path of the Gulf Stream and possibly forcing it away from the continental slope as far back as Cape Hatteras.

The Florida Current emerges as the part of the Gulf Stream System that is best documented, analyzed, and understood. The Gulf Stream itself is likely to be more complex, but is as yet poorly measured by comparison. The systematic program of moored current and temperature measurements developed by Schmitz (1976, 1977, 1978) are slowly building a foundation of time-series data that will enable the next interpretive steps to be taken. Because of the complexities of the Gulf Stream System, understanding of the behavior in terms of dynamics will rely heavily on numerical modeling and analysis. It is essential that the observational and numerical studies of the Gulf Stream System proceed cooperatively.