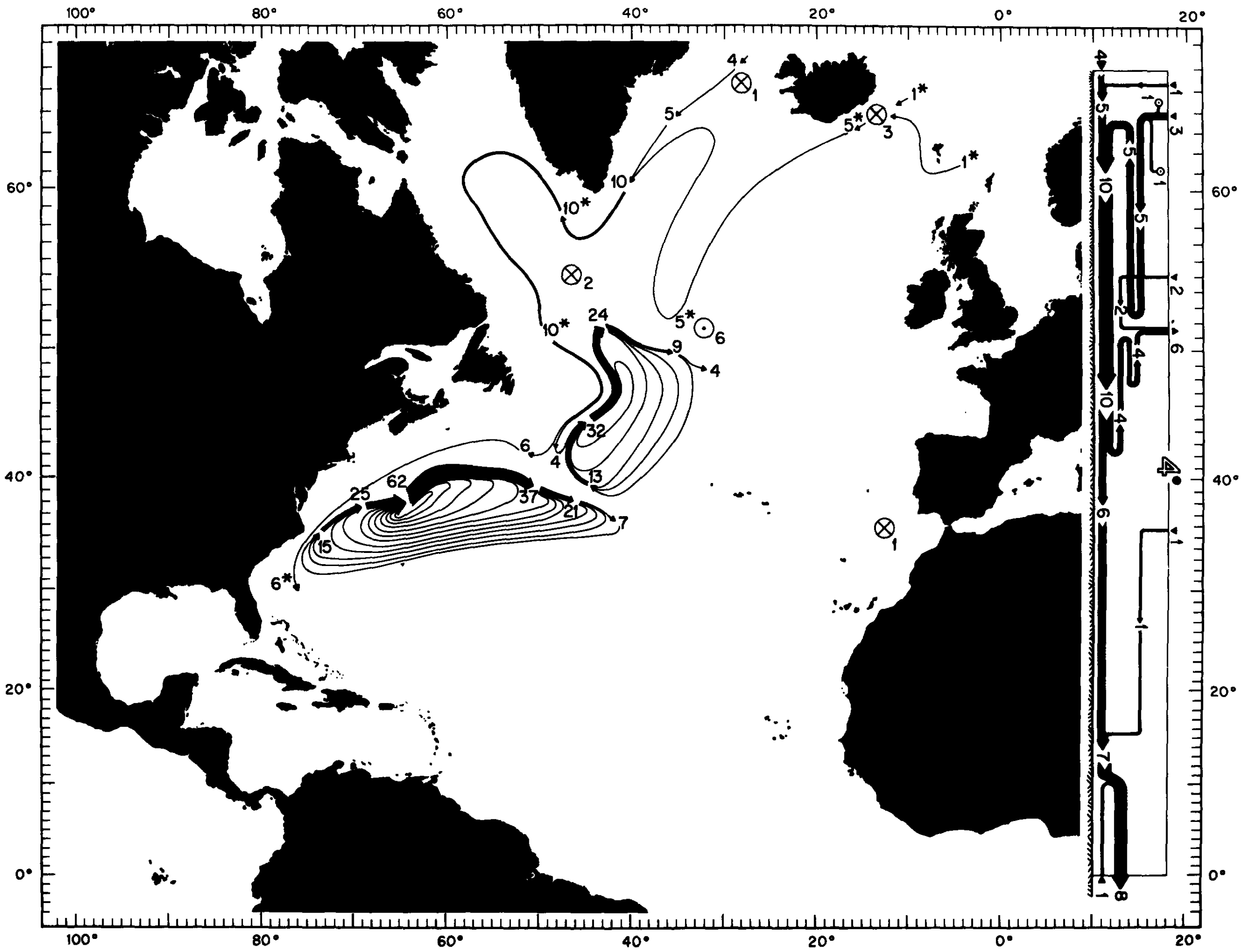


Figure 4.6 Circulation diagram for the deep ($\theta < 4°C$) circulation in the North Atlantic. Worthington estimates a flow of $62 \times 10^6\ m^3 s^{-1}$ in the recirculation gyre with $6 \times 10^6\ m^3 s^{-1}$ flowing southward inshore of the recirculation. (Worthington, 1976.)

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

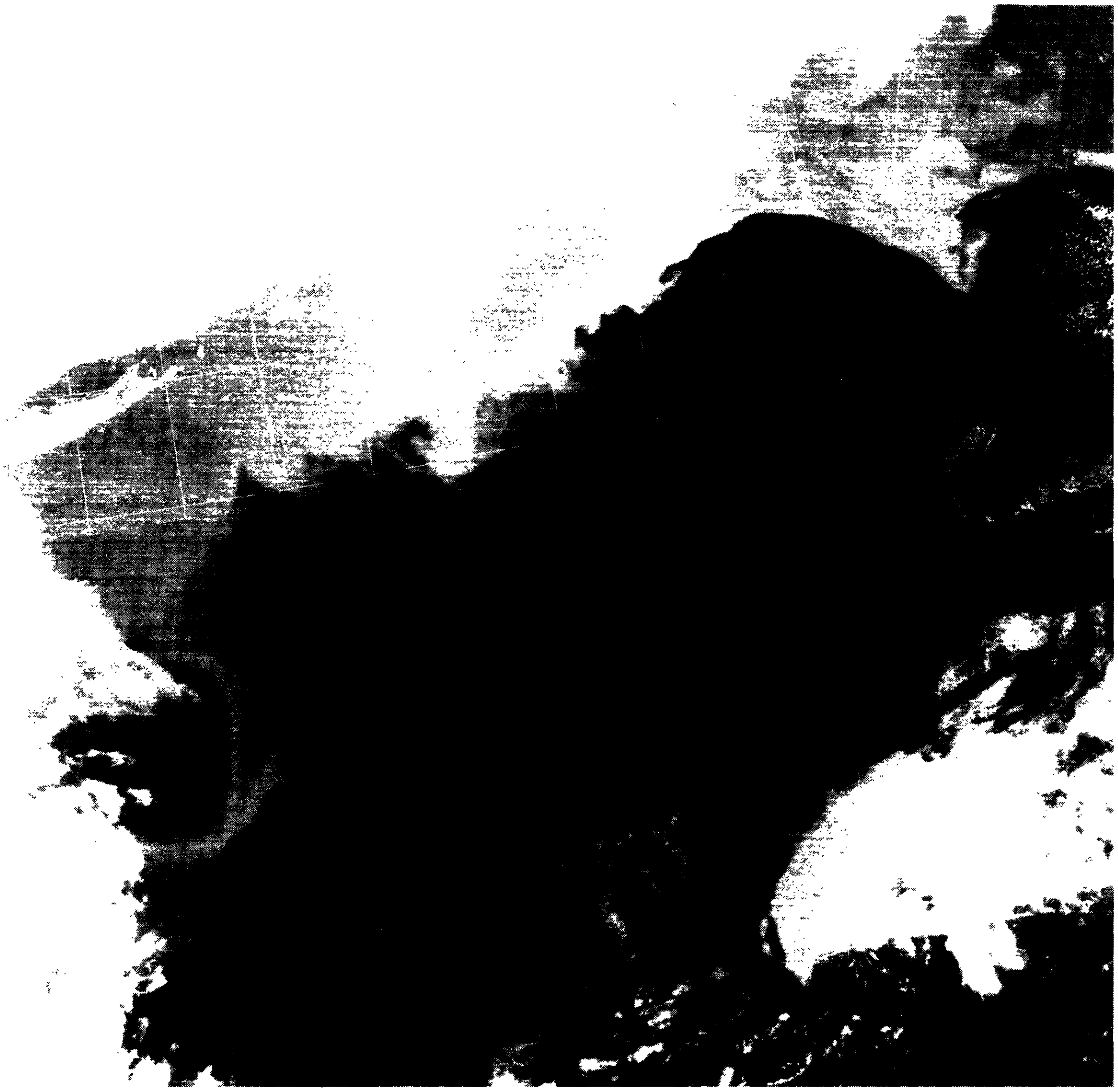


Figure 4.7 The Gulf Stream south of Cape Cod, showing well-developed meanders with several eddies formed in the slope water to the north (Courtesy of R. Legeckis, NOAA-NESS, from NOAA-4 satellite November 12, 1975.)

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 × 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

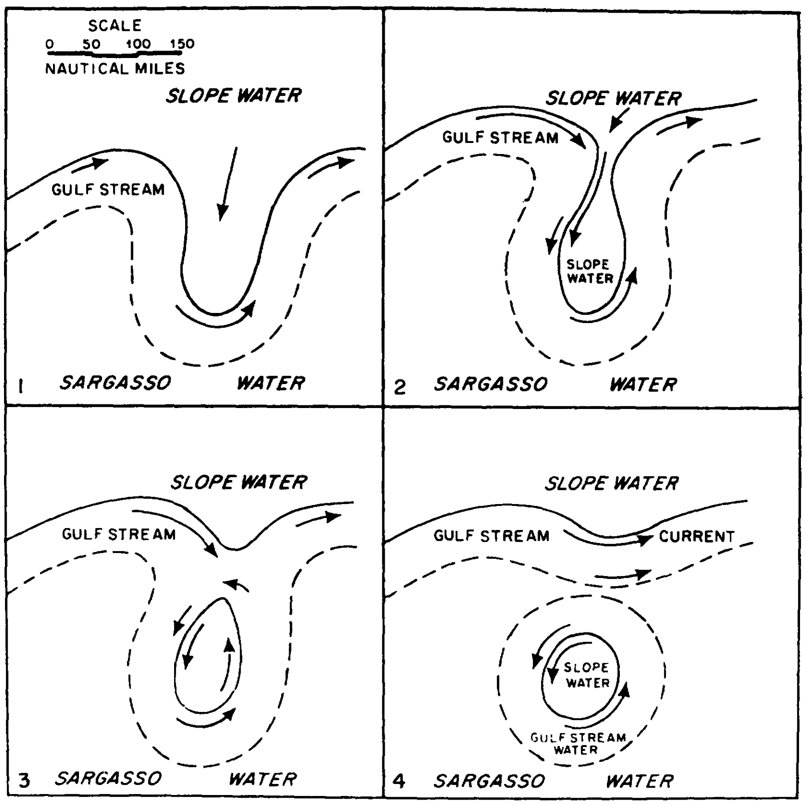


Figure 4.8 Diagram of Gulf Stream ring generation from meander formation to separation. (Parker, 1971.)

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

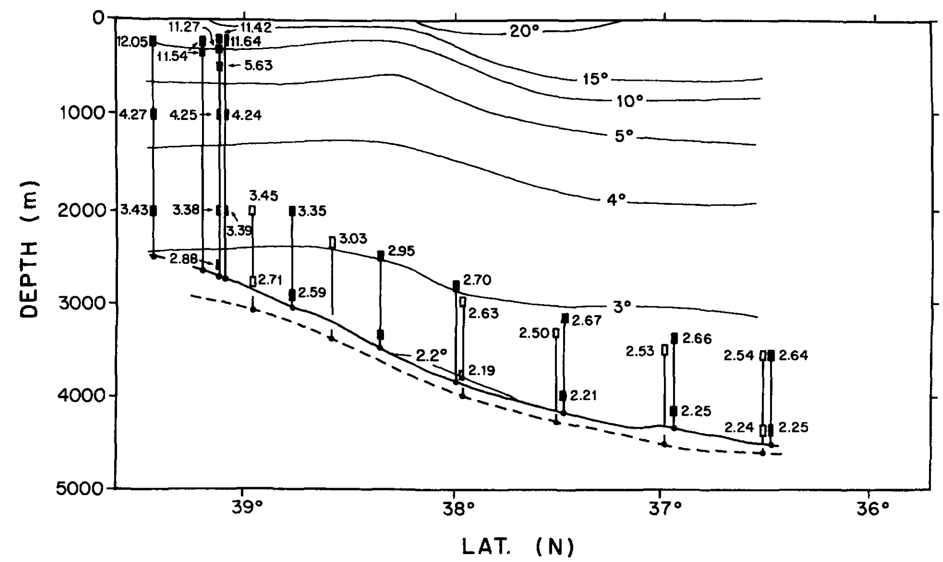


Figure 4.9A Distribution of current meters and temperature-pressure recorders on the Luyten (1977) Continental Rise array. The solid lines refer to 70° W, dashed to 69°30′ W. (Luyten, 1977.)

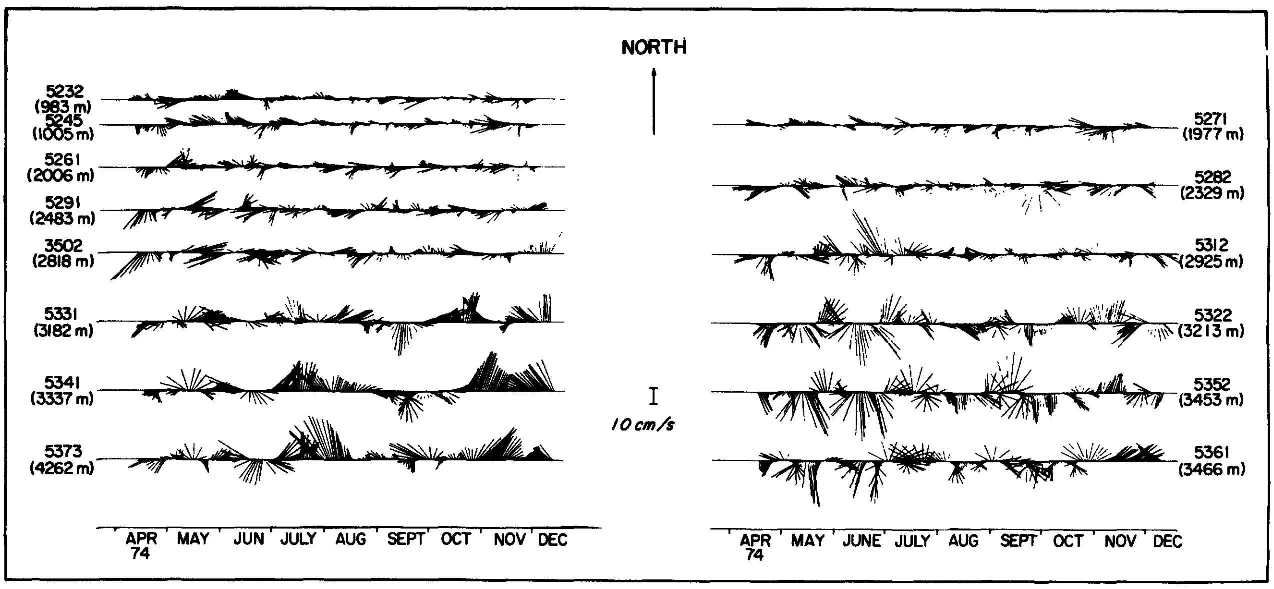
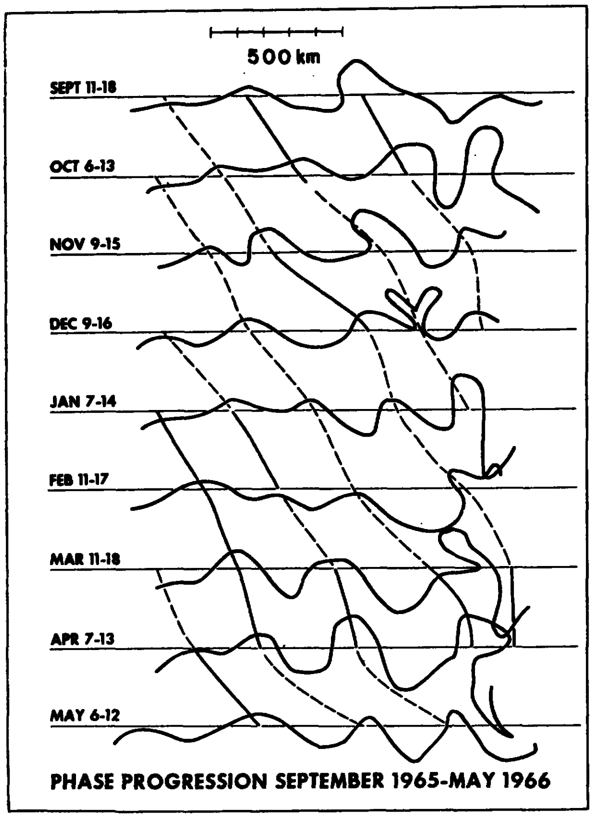


Figure 4.9B Time sequence of 1-day averaged current. The numbers identify mooring and instrument. Instrument depths in meters are shown in parentheses. (Luyten, 1977.)



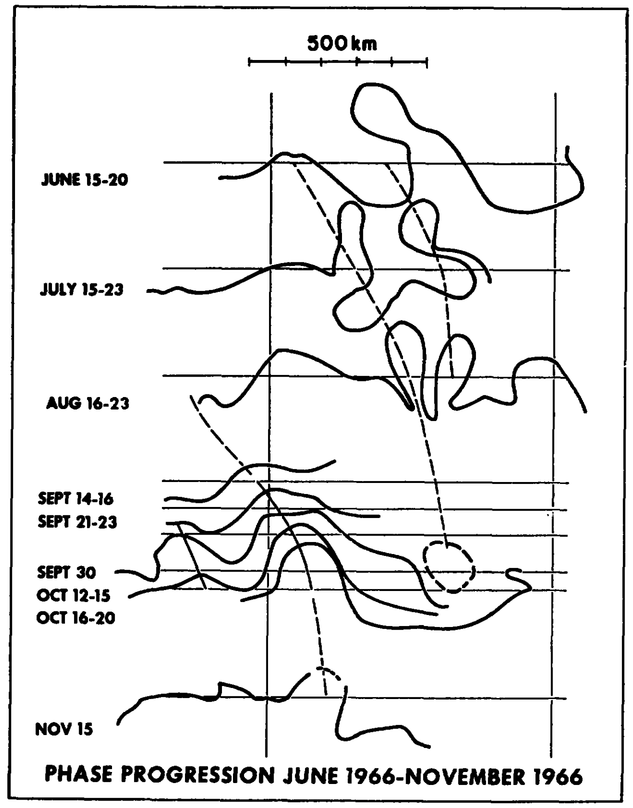


Figure 4.10 Inferred progression and evolution of meanders relative to the mean Gulf Stream path. The diagonal lines show phase propagation (solid where supported by other evidence). (Hansen, 1970.)

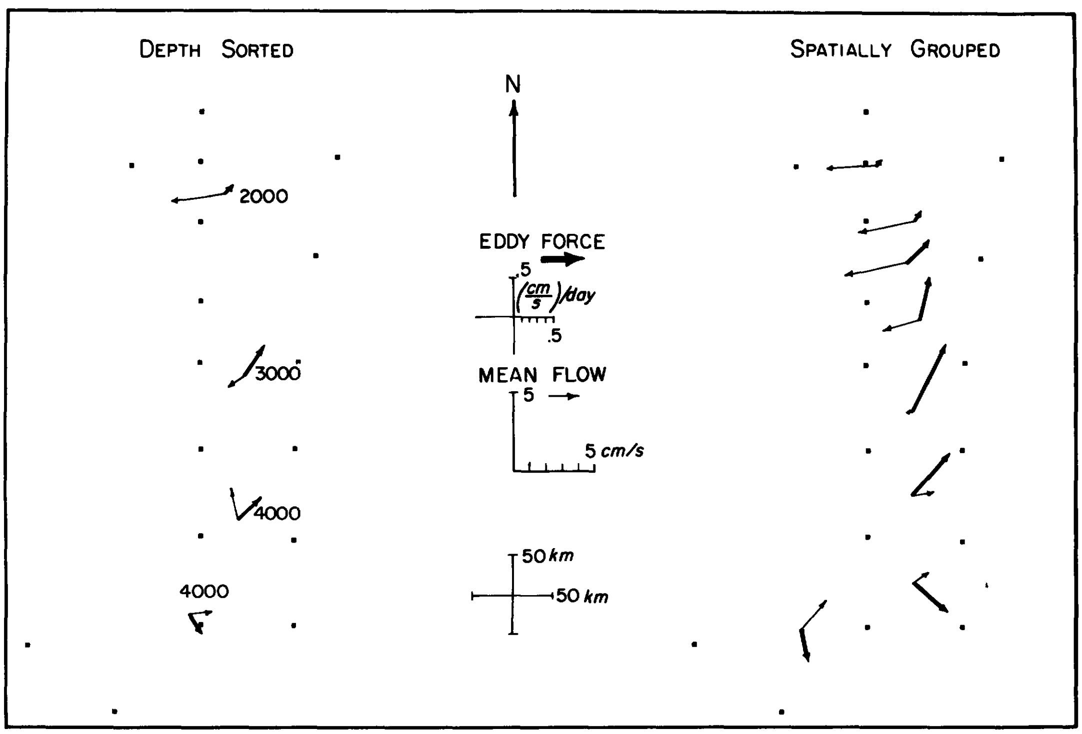


Figure 4.11 Vector acceleration of mean flow by eddy Reynolds stress gradients. The "eddy forces" tend to oppose the mean flow on the continental rise and to accelerate it under the Gulf Stream at the southern portion of the array. The eddy forces are estimated by grouping the data in depth intervals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

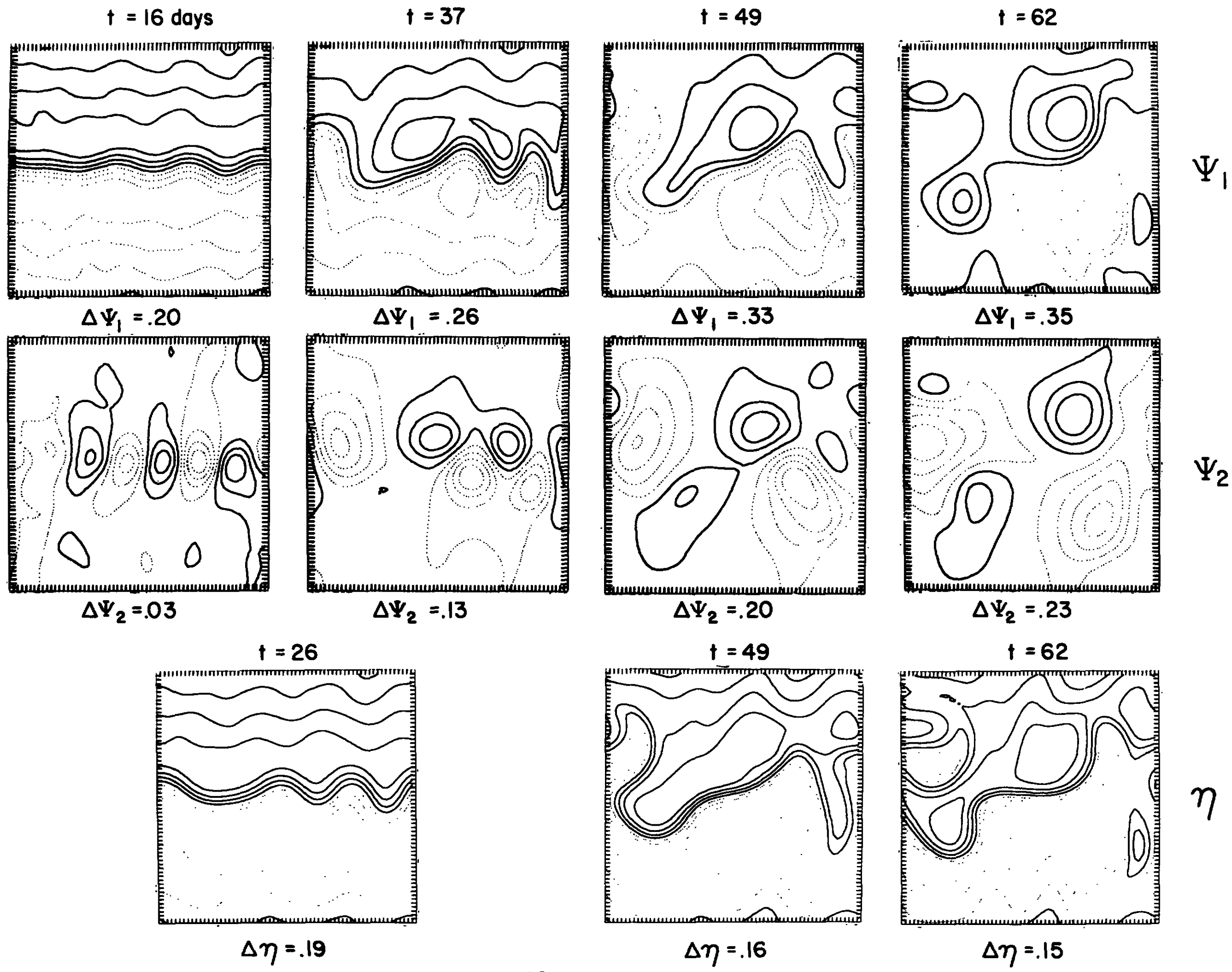


Figure 12

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

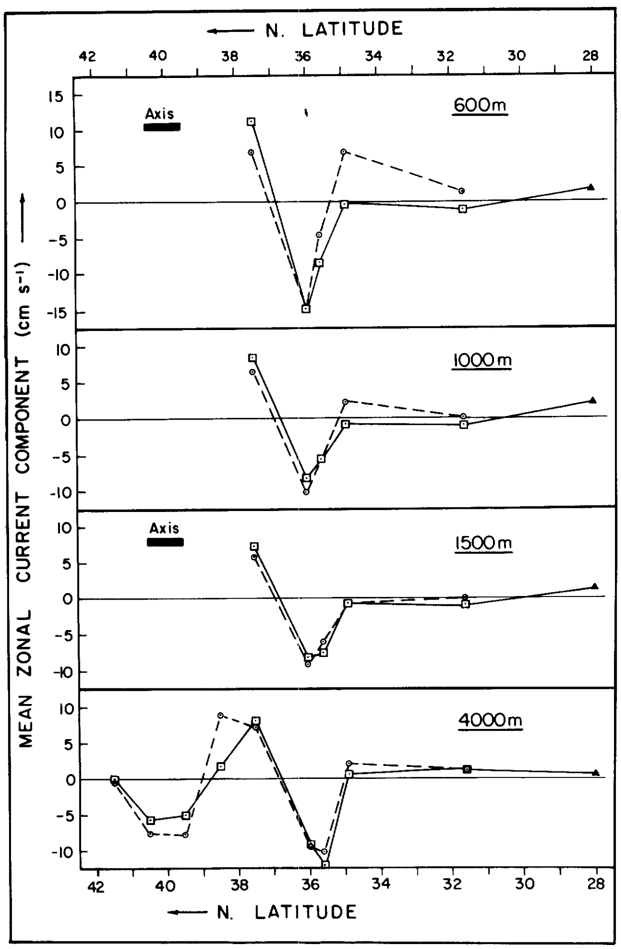


Figure 4.13 Meridional distribution of time-averaged zonal current at four depths along 55°W from the first POLYMODE array shown by circles ○ and from the combined first and second setting of the array shown by squares □. The approximate mean axis of the Gulf Stream is indicated. The westward mean flow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

The mass flux of the Gulf Stream has been estimated to be in the range of 100–150 $\times$ 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 $\times$ 10^6 $m^3 s^{-1}$. Thus the recirculating portion may be as much as 60–110 $\times$ 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 $\times$ 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 $|\boldsymbol{\tau}| \cdot |\mathbf{v}| \approx 2$ $erg\,cm^{-2} s^{-1}$. Over the entire basin (area $\approx 1 \times 10^{17}$ 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 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 80×10^6 m^3 s^{-1}, making a total transport of 120×10^6 m^3 s^{-1}. The total kinetic energy flux would be three times the interior value or 6×10^{17} ergs s^{-1} if the section deepened with no change of mean velocity. Only 2×10^{17} erg s^{-1} must be dissipated and 40×10^6 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 $1.7 \pm 0.2 \times 10^{21}$ ergs day^{-1}, or about 37 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

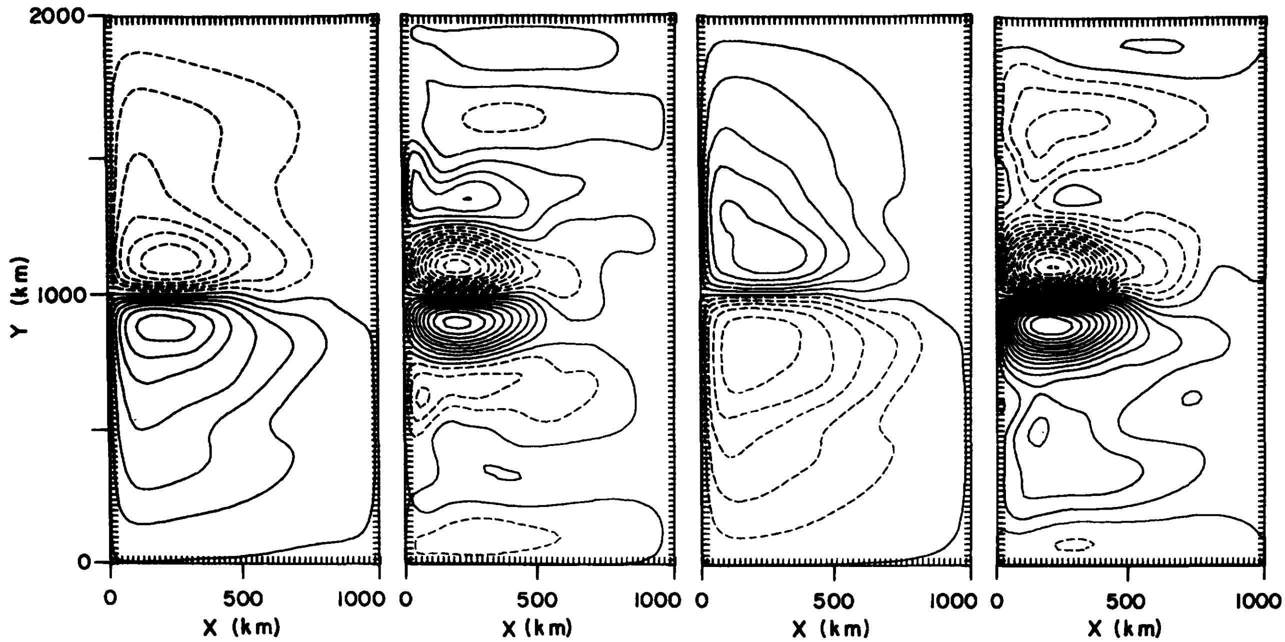


Figure 4.14 The mean fields for Holland experiment 3: (A) upper-layer streamfunction (contour interval CI = 5000 $m^2 s^{-1}$); (B) lower-layer streamfunction (CI = 1000 $m^2 s^{-1}$); (C) interface height (CI = 20 m); (D) total transport (CI = 50 × 10^6 $m^3 s^{-1}$). (Holland, 1978.)

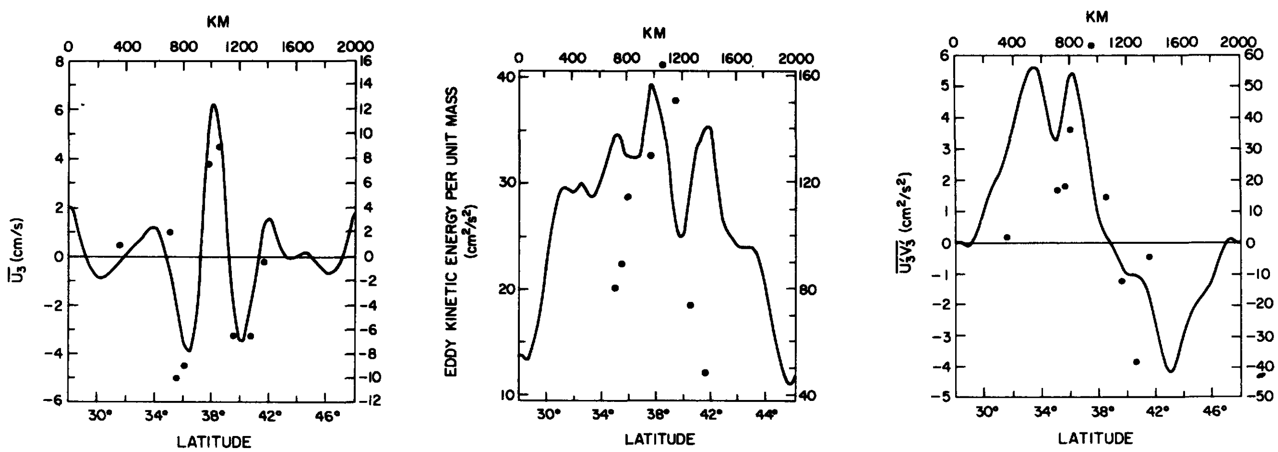


Figure 4.15 Comparison of meridional distributions of (A) zonally averaged mean flow $\bar{u}_3$, (B) eddy kinetic energy $\overline{u_3'^2 + v_3'^2}$, and (C) Reynolds stress $\overline{u_3'v_3'}$ in the lower layer (subscript 3) for experiment 3 compared with observed distribution (Schmitz, 1977) along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

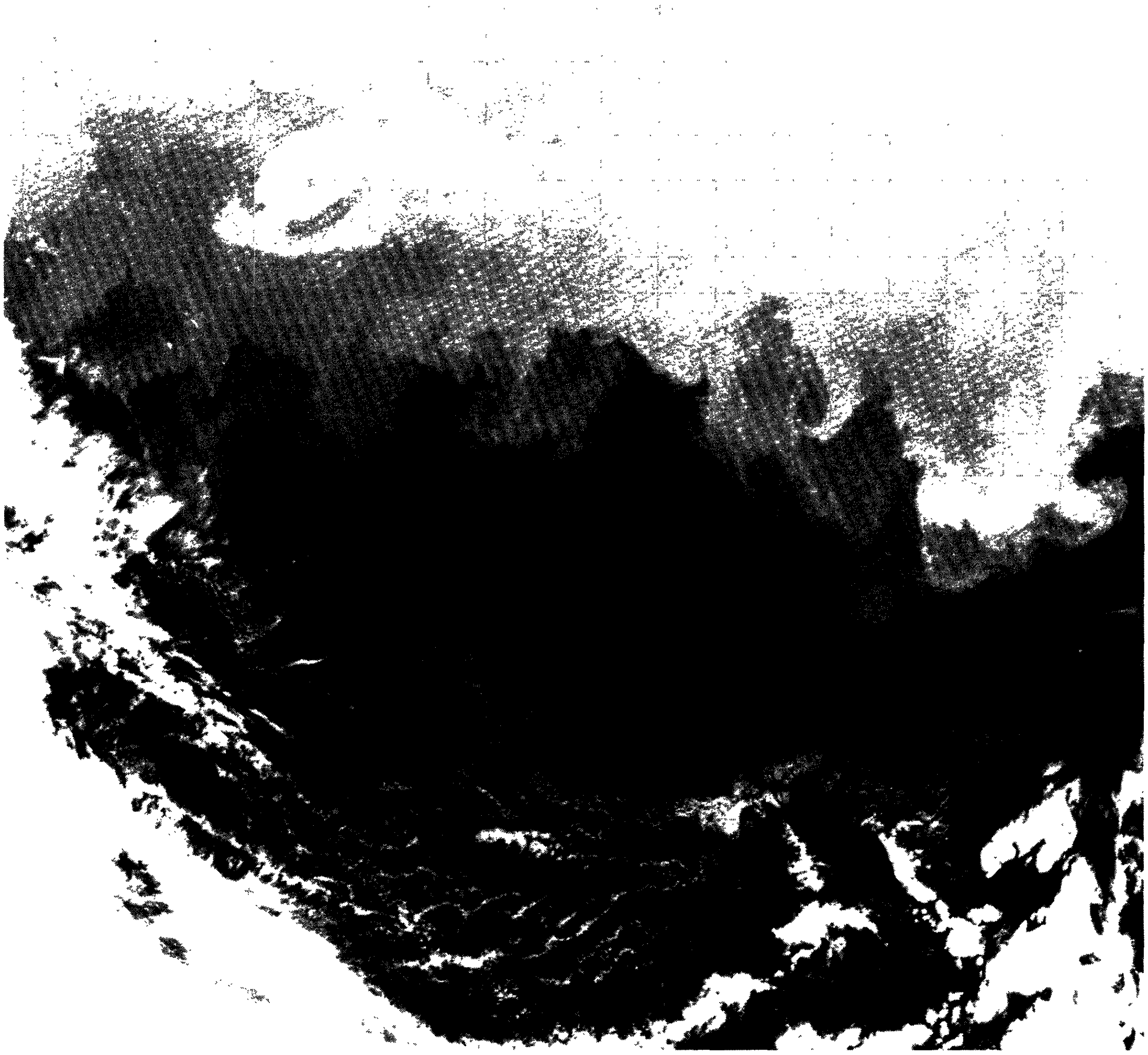


Figure 4.16 The Gulf Stream and North Atlantic Current as seen from an infrared image taken November 2, 1977. The figure shows the colder shelf and slope water responding to a complex meandering of the Gulf Stream. The cold ribbon of Labrador Water is seen flowing south along the eastern edge of the 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 $20 \times 10^6\ m^3\,s^{-1}$ joined by about $15 \times 10^6\ m^3\,s^{-1}$ originating in the slope water formed the branch of the North Atlantic Current to the north. A southward-flowing branch of about $30 \times 10^6\ m^3\,s^{-1}$ 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

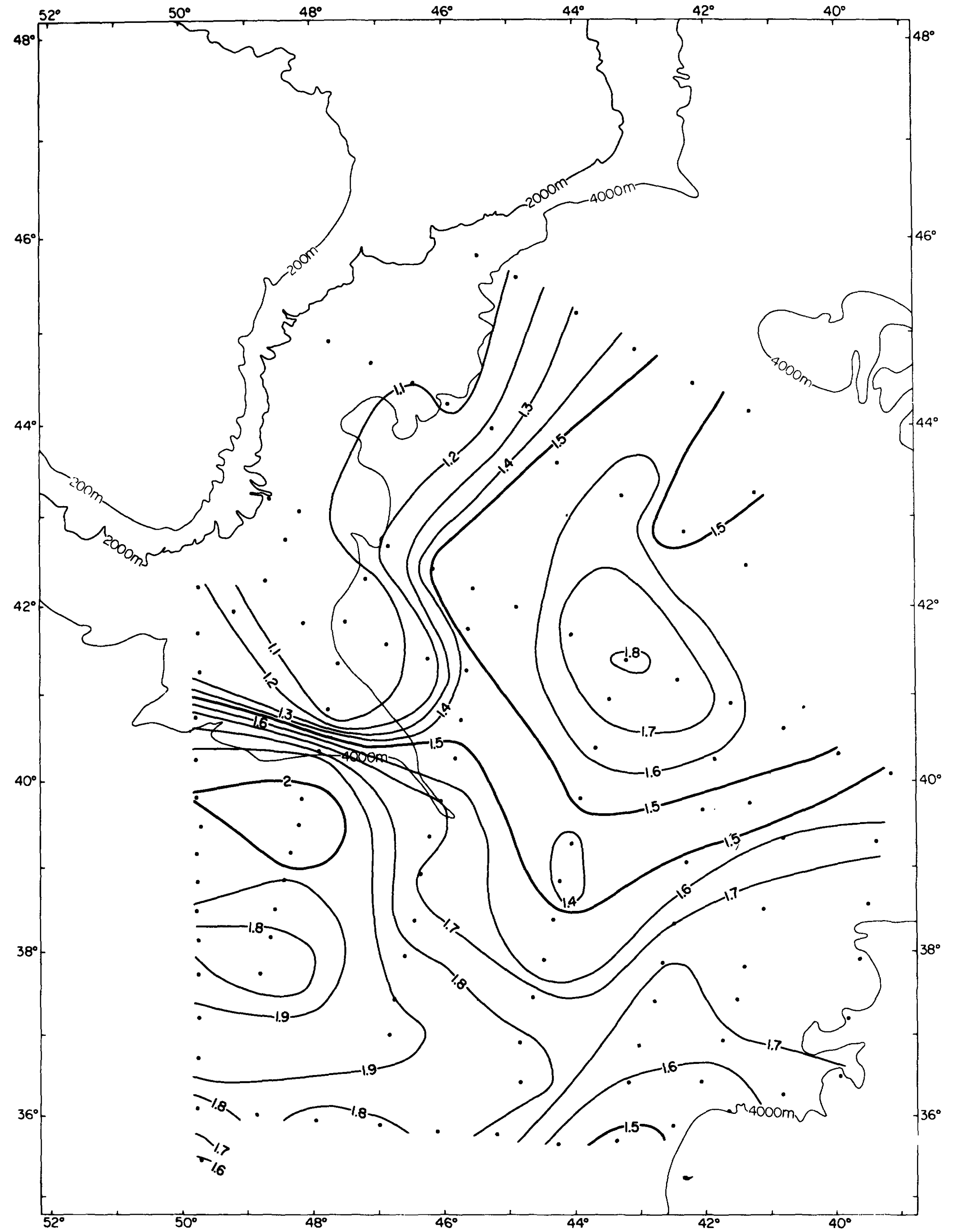


Figure 4.18 Dynamic topography from the North Atlantic Current showing the separation of the Gulf Stream into a northern branch flowing around a weak, anticyclonic eddy in the Labrador Basin. (Clarke et al., 1980.)

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.