

Gulf Stream Meanders Over Steep Topography

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Observations of the Gulf Stream thermocline as it meanders over the continental slope off Cape Hatteras are presented. Daily vertical sections of temperature and geostrophic velocity to 1500 m are calculated from an array of inverted echo sounders aligned across the Gulf Stream. These accompany time series of deep currents from a simultaneous array of bottom current meters. The vertical sections are computed using a previously derived technique which assumes the Gulf Stream variability is primarily first baroclinic mode. The sections reveal that the Gulf Stream front systematically shoals when it meanders offshore, an effect not observed farther downstream. This is accompanied by a compression of the main thermocline that in turn increases the core velocity of the jet but has little effect on overall transport. To explain this shoaling, the Gulf Stream is treated as a two-layer flow; the lower layer vorticity balance suggests that the shoaling is caused by deep water columns preserving their layer thickness in the presence of the steep topography.

1. INTRODUCTION

The Gulf Stream is one of the most thoroughly studied currents in the world ocean, yet many uncertainties remain regarding its complex dynamical nature. In the mean, the Gulf Stream flows from Cape Hatteras to the northeast, obtaining a maximum transport of roughly 150 Sv near 60°W [Hogg, 1990]. Further to the east its transport diminishes owing to recirculations both to the north and south. Much of the reason for the complexity of the Gulf Stream is that it is both baroclinically and barotropically unstable [e.g., Johns, 1988; Hall, 1986a]. This causes perturbations to grow within the current both in time and space, and as a result the Gulf Stream is characterized by large meanders which often become quite convoluted.

The dispersion characteristics of Gulf Stream meanders have been investigated using satellite imagery [e.g., Halliwell and Mooers, 1983] and in situ measurements [e.g., Watts and Johns, 1982]. Meanders exist at periods from as short as a few days to as long as several months and propagate primarily downstream with phase speeds decreasing with increasing wavelength. The structure and kinematics of the meanders appear to be quite complex. For instance, Bower and Rossby [1989] have shown using isopycnal floats that water parcels undergo significant vertical excursions during large meanders and that they are often expelled from the current. Even more complicated are the underlying dynamics governing the meanders.

Near Cape Hatteras, where the Gulf Stream leaves the continental shelf and flows into deeper water, meanders in the current are quite small. The standard deviation of lateral displacements there is less than 10 km [Pickart and Watts, 1990a]. Downstream, however, meanders grow quite rapidly, and beyond 70°W the lateral standard deviation is over 50 km [Halliwell and Mooers, 1983]. Thus one would expect these different geographical regions to represent different

dynamical regimes as well, one region being characterized by small sinusoidal meanders, the other being dominated by large convoluted disturbances that periodically form eddies and rings. In addition to the small meander envelope near Cape Hatteras, this region is also unique in that the Gulf Stream encounters steep topography (other than isolated seamounts further downstream). As the current flows off the shelf into deeper water, the configuration of the topography is such that for a short distance the Gulf Stream flows along the steep continental slope (Figure 1). This is in sharp contrast to the downstream portion (and vast majority) of the current, which flows along the weak continental rise and abyssal plain.

In this paper we present results from a moored array of inverted echo sounders (IESs) and bottom current meters in the Gulf Stream near Cape Hatteras, where the Gulf Stream flows over topography that is steeper than its own frontal slope. Results are contrasted to previous observations made just 100 km downstream, but where the bottom slope is smaller than the frontal slope. Important differences in the structure of the observed meanders are found at the two locations which can be attributed to topographic influence. In section 2 we describe the moored array and explain how the inverted echo sounder data are converted into thermocline temperatures and geostrophic velocities. The thermocline data are then used to create daily temperature and geostrophic velocity sections of the Gulf Stream which are described in section 3. In section 4 we use the moored data to present a two-layer description of the Gulf Stream which reveals the unique character of the meanders at Cape Hatteras. The effect of topography is investigated within the context of the lower layer potential vorticity balance. Finally, implications of strong topography on a continuously stratified Gulf Stream are considered in section 5.

2. THE MOORED DATA

The moored array is shown in Figure 2, consisting of a central line of five IESs and five bottom current meters, as

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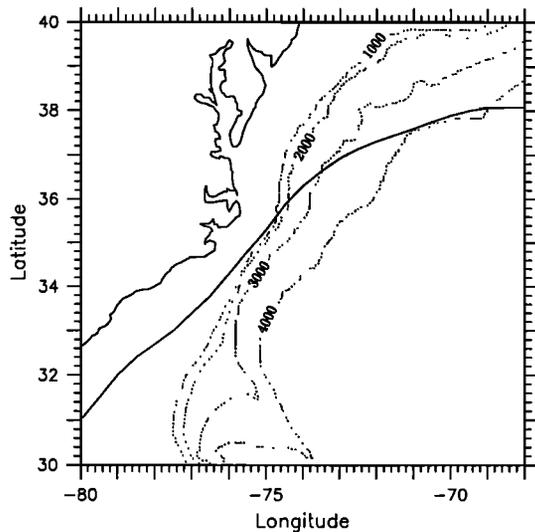


Fig. 1. Mean north wall of the Gulf Stream [from Olson *et al.*, 1983; P. Cornillon, personal communication, 1992] in relation to the topography.

well as two upstream and downstream IESs. This array was in place for nearly 3 years, from October 1987 to August 1990, as part of the Synoptic Ocean Prediction (SYNOP) experiment. Although we present results only from the first 8-month deployment, preliminary analysis on the full data set indicates that the results are representative of the entire measurement period. In particular, the mean Gulf Stream path and mean deep currents, as well as the associated variances, are nearly identical for each of the three deployments. For example, the 3-year standard deviation of Gulf Stream lateral displacements differs by only 1.7 km from that of the first deployment.

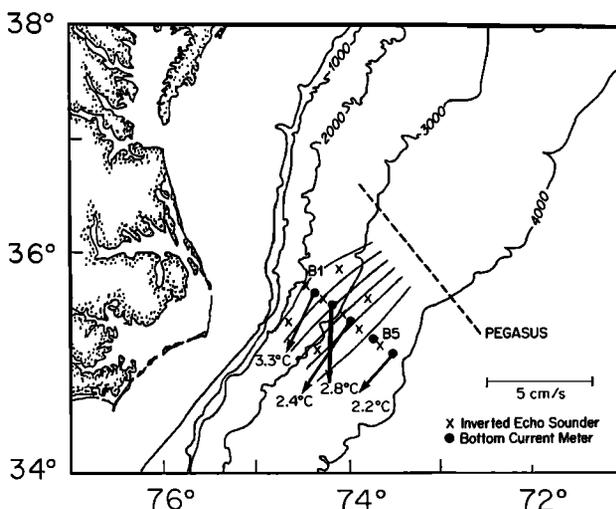


Fig. 2. The SYNOP array at Cape Hatteras, consisting of nine IESs (crosses) and five bottom current meters (dots). The central line of five IESs are labeled B1 to B5 (onshore to offshore). The mean topography of the Gulf Stream's 12°C isotherm surface from the IESs is shown by the thin lines (200–700 m by 100-m increments). The mean bottom currents are indicated by the thick arrows, with the mean temperatures alongside. The dashed line is the location of the repeat Pegasus sections [Halkin and Rossby, 1985] discussed in the text.

While the central line of instruments spans the Gulf Stream, the deep mean current vectors in Figure 2 all point to the southwest at 3–5 cm s^{-1} (one of the instruments failed to release). This is because the deep western boundary current (DWBC) crosses under the Gulf Stream at precisely this location. Pickart and Watts [1990a] analyzed the first deployment of bottom current meter data and showed that at periods longer than 10 days the observed variability was due mostly to bottom-trapped topographic Rossby waves, but also to fluctuations of the DWBC. By contrast, the IES data are not sensitive to the deep currents but primarily measure variations in the main thermocline. This was tested quantitatively by using conductivity-temperature-depth (CTD) data occupied near the central line. Specifically, for a Gulf Stream meander we computed the percent fluctuation in acoustic travel time due to the bottom 1000 m and found it to be less than 3%. Thus we interpret the IES data as information exclusively on the upper level Gulf Stream.

Pickart and Watts [1990b] have demonstrated that the IES can accurately measure the local vertical profile of temperature and dynamic height through the main thermocline, provided the dominant variability can be characterized by the first baroclinic mode. (Sea surface height changes due to the Gulf Stream and tides are of the order of 1 m or less, so the variation in travel time due to these barotropic fluctuations is negligible.) Pickart and Watts [1990b] tested their methodology on previously collected IES data at this same location in the Gulf Stream and found that the IES-derived temperatures and geostrophic velocities agreed well with independent moored measurements. We have applied this methodology to our central line of IESs to produce daily vertical sections of Gulf Stream temperature and geostrophic velocity to 1500 m. Briefly, the IES round trip acoustic travel time is converted into an amplitude of the local first baroclinic mode which is then used to perturb a local basic state temperature and dynamic height profile, resulting in a time-varying vertical profile. The profile is made absolute by using a reference travel time obtained from matching several of the instantaneous temperature profiles to in situ expendable bathythermographs (XBTs). The reader is referred to Pickart and Watts [1990b] for details. This procedure is done at each of the IES sites along the central line, resulting in a set of daily vertical sections. Then the dynamic height fields are cross-stream differenced to obtain sections of geostrophic velocity.

Traditionally, IESs have been used in the Gulf Stream to obtain the height of a specified isotherm in the main thermocline using an empirical relationship between the acoustic travel time and isotherm depth [see Watts and Johns, 1982]. Such an isotherm depth can of course be extracted from our vertical temperature profile as well. When this is done, the resulting time series of isotherm displacements at the individual IES sites are virtually indistinguishable from the analogous time series obtained through the traditional empirical technique.

3. TEMPERATURE AND GEOSTROPHIC VELOCITY SECTIONS

While the first baroclinic mode accounts for most of the Gulf Stream variability within the main thermocline [Pickart and Watts, 1990b; Hall, 1986b; Rossby, 1987], it obviously does not apply near the surface or close to the bottom (where the first-baroclinic-mode amplitude goes to zero, thus imply-

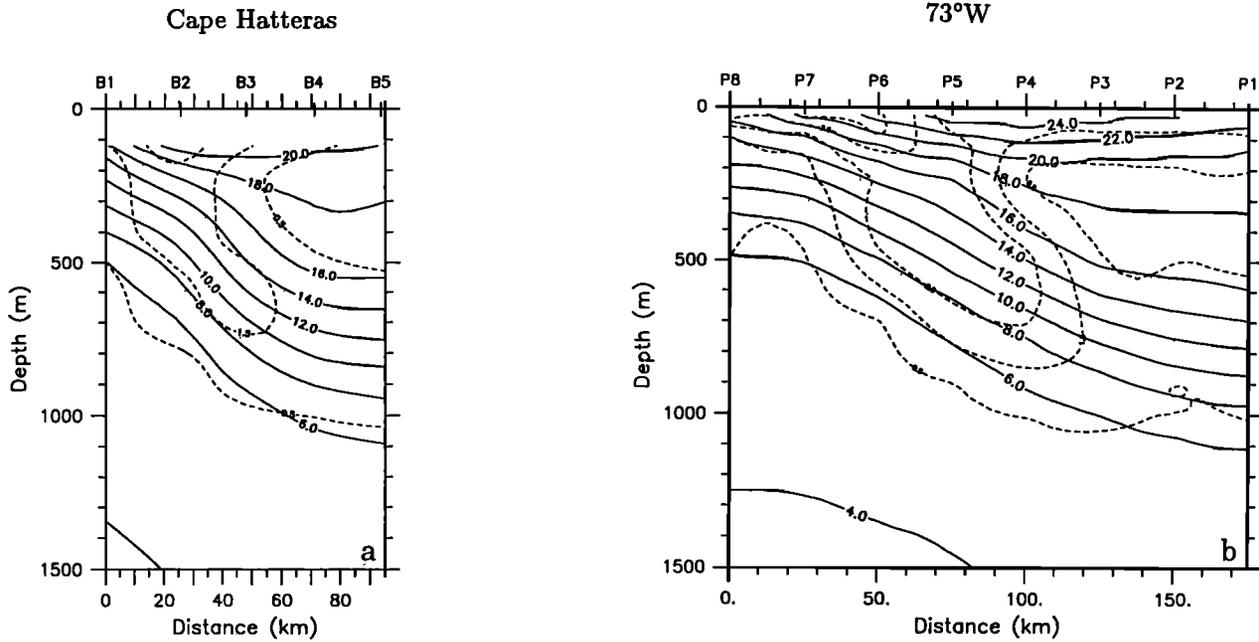


Fig. 3. (a) Mean vertical section of temperature to 1500 m at Cape Hatteras, calculated from the central line of IESs using the technique described in the text. The standard deviation is overlaid (dashed lines). (b) Mean Pegasus temperature section and standard deviation at 73°W (100 km downstream of Cape Hatteras).

ing zero variance). It is evident from the IES-derived sections that the first-mode description fails at depths shallower than 100 m. We chose 1500 m (which spans the entire thermocline) as the deep cutoff for our sections, although the velocities are actually referenced to 2000 m, which is the average level of no motion determined in this part of the Gulf Stream by *Pickart and Watts [1990b]*.

3.1. Downstream Broadening

Figure 3a shows the mean IES-derived temperature section at the central instrument line for the 8-month deployment period. This is to be compared with an analogous mean section of Gulf Stream temperature (Figure 3b) computed from 3 years of Pegasus data at 73°W, which is 100 km downstream of our array (Figure 2; *Halkin et al. [1985]*). The Eulerian or geographical mean Gulf Stream broadens by ~85% over this short downstream distance (the cross-stream axes in the two figures are on the same scale). The standard deviation of lateral displacements of the Gulf Stream was ±7.5 km near Cape Hatteras and ±20.8 km at the Pegasus line. Note that this difference is not a result of the different averaging lengths, since the 8-month standard deviation at Cape Hatteras is nearly identical to the 3-year standard deviation. While part of the difference might be due to the different sampling periods (the Pegasus experiment occurred in 1980–1983, and SYNOP in 1987–1990), historical data indicate that there are comparable discrepancies from year to year [*Watts, 1983*]. The two sections in Figure 3 are very similar in character; the highest standard deviations occur at the surface north wall, with a tongue extending into the main thermocline centered on 12°C associated with the meandering of the current. When the mean sections are computed in stream coordinates, i.e., aligning each individual section at the north wall (which removes the meandering [see *Halkin and Rossby, 1985*]), the discrepancy between the two locations is reduced. However, the “instantaneous” Gulf

Stream at the Pegasus line is still ~40% wider than at Cape Hatteras (not shown).

Figures 4a and 4b show the geographical mean Gulf Stream velocity sections at the two alongstream locations. Remember that the velocities at Cape Hatteras are geostrophic (referenced to 2000 m), while the Pegasus velocities are measured directly and include ageostrophic contributions. For proper comparison we have referenced the Pegasus velocities to 2000 m. Note in Figure 4 that the measured mean core speed of the Gulf Stream (at 100-m depth) has decreased downstream from a value of roughly 140 cm s⁻¹ to 110 cm s⁻¹.

3.2. Energy Implications

The striking change in width of the observed mean Gulf Stream from Cape Hatteras to the Pegasus line (both in geographical and stream coordinates) has important implications regarding the energetics of the current. It is worthwhile to compute the change in mean kinetic and available potential energy of the Gulf Stream over this distance and to compare the result with previous work. Our calculation will be done both in geographical coordinates, which includes the meandering, and in stream coordinates, which measures changes in the synoptic Gulf Stream. We consider a simple two-layer (rigid lid) representation of the Gulf Stream with the 12°C isotherm as the interface (Figure 5a). The height η is the interface displacement relative to the level of zero available potential energy. Note that in this framework, both the available potential energy (APE) and kinetic energy (KE) are obtained from the cross-stream profile of η (Figure 5a),

$$\frac{\text{APE}}{\text{unit area}} = \frac{g}{2} (\rho_2 - \rho_1) \eta^2, \tag{1}$$

$$\frac{\text{KE}}{\text{unit volume}} = \frac{\rho_1 g'^2}{2f^2} \eta_y^2, \tag{2}$$

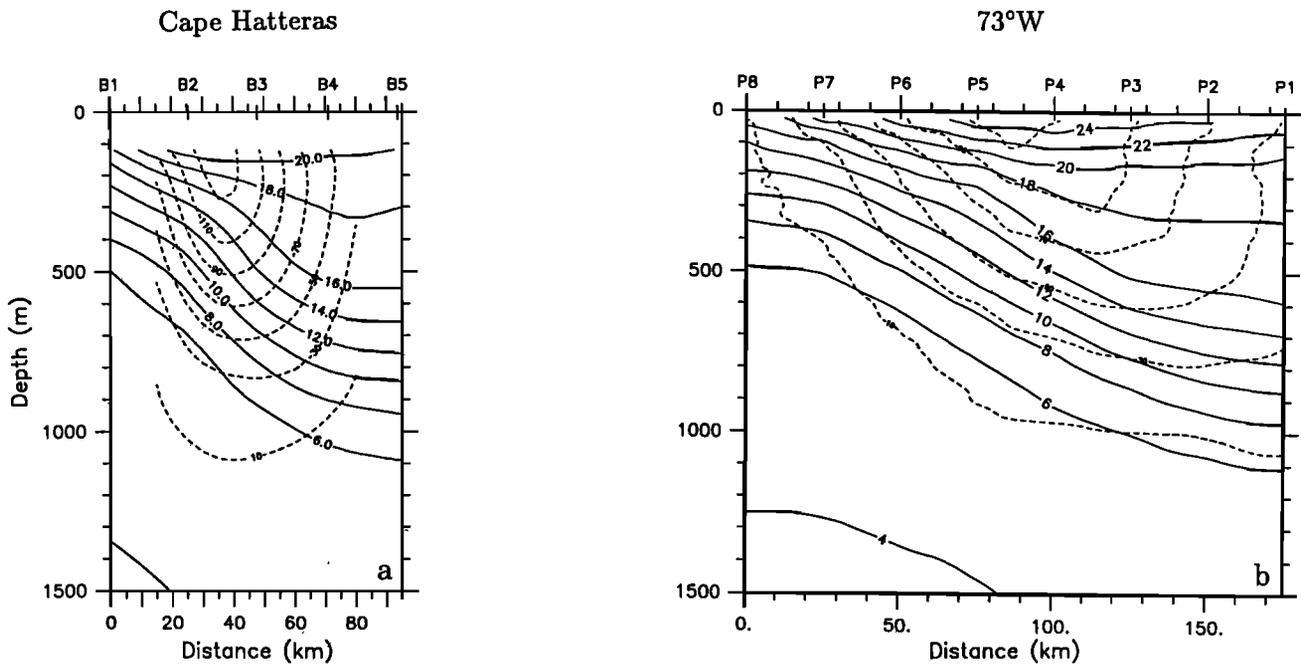


Fig. 4. (a) Mean geostrophic velocity at Cape Hatteras (dashed lines) referenced to 2000 m, overlaid on the mean temperature section of Figure 3a. The velocities were computed using the IES-derived dynamic height sections. (b) Mean Pegasus velocity section at 73°W overlaid on the mean Pegasus temperature section.

where x and y are downstream and cross-stream distance, f is the Coriolis parameter, and $g' = g(\rho_2 - \rho_1)/\rho_2$ is the reduced gravity.

We take $\eta = A \tanh (y/w)$, where A (amplitude) and w

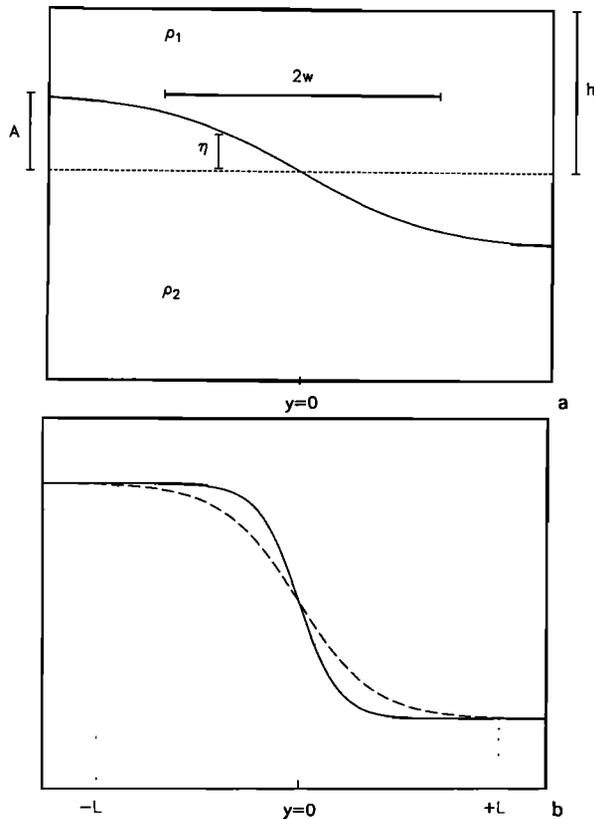


Fig. 5. (a) Schematic of the two-layer Gulf Stream with interface $\eta = A \tanh (y/w)$. The dashed line is the level of zero available potential energy; h is the height of the rigid lid. (b) Schematic showing the cross-stream limits of integration for comparing an upstream and downstream Gulf Stream profile.

(current width) are determined by fitting η to the observed mean 12°C profile. Equation (1) is integrated across stream, and (2) is integrated vertically and across stream. The resulting expressions for available potential and kinetic energy integrated across the Gulf Stream are

$$\int APE = A^2 g(\rho_2 - \rho_1) L \left[1 - \frac{w}{L} \tanh (L/w) \right], \quad (3)$$

$$\int KE = \frac{A^2 \rho_1 g'^2 h}{f^2 w} [\tanh (L/w) - 1/3 \tanh^3 (L/w)], \quad (4)$$

where the integration is from $-L$ to $+L$, and h is the undisturbed upper layer thickness (Figure 5a).

To compare $\int APE$ at two locations in the Gulf Stream, the integration should be carried out to where the two interface profiles meet (Figure 5b); note then that L is greater than w (which is an e -folding distance). This choice for L is appropriate for $\int KE$ as well (which is insensitive to L for $L > w$). Figure 6 shows the percent decrease of $\int APE$ and $\int KE$ for a range of Gulf Stream widths. In our case, for the geographical mean, the percent decrease of available potential energy from Cape Hatteras to Pegasus is 21%, and the percent decrease of kinetic energy is 46%. For the stream coordinate mean, the loss of available potential energy is 15% and the kinetic energy loss is 30% (Figure 6).

These losses, which are quite substantial over such a small downstream distance (100 km), can be compared to Rossby's [1987] energy conversion rates calculated using the Pegasus time series data. The aforementioned decrease in mean energy (ignoring pressure work) will result in an increase in eddy energy,

$$\frac{dE'}{dt} = -\bar{U} \frac{d\bar{E}}{dx}, \quad (5)$$

where dE'/dt is the rate of conversion of total energy ($\int APE + \int KE$) to the eddies, $d\bar{E}/dx$ is the downstream gradient of

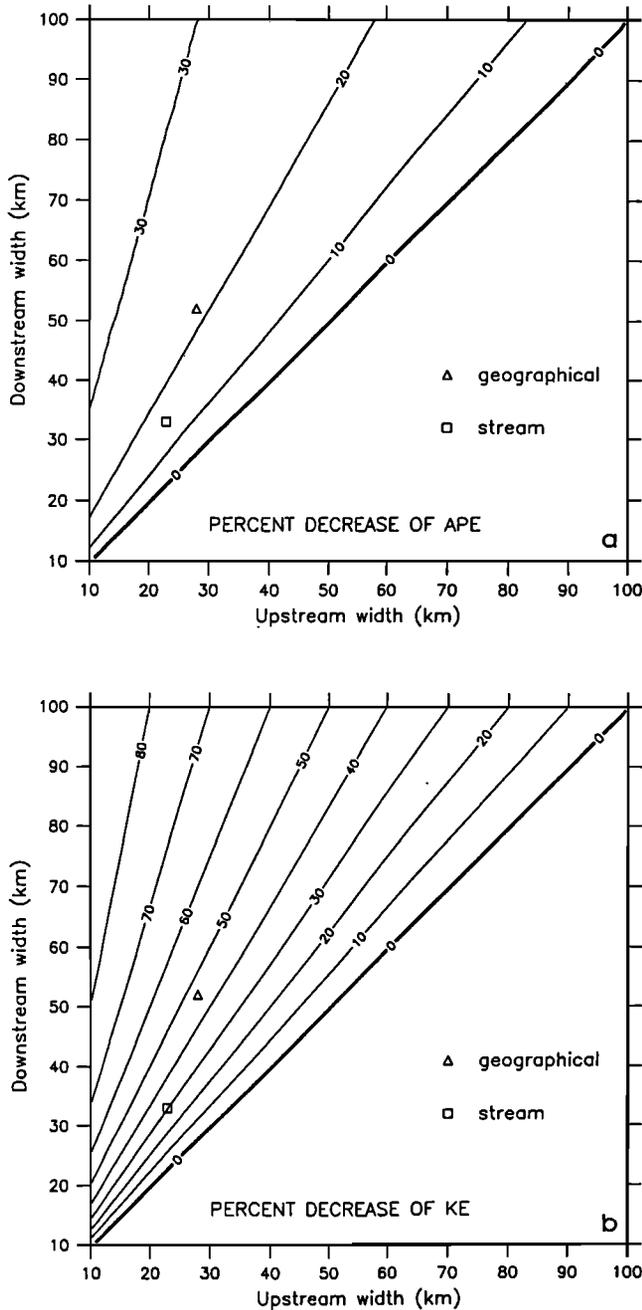


Fig. 6. (a) Percent decrease of f APE as a function of along-stream change in current width w . The value for Cape Hatteras to 73°W is indicated (for both geographical and stream coordinates). (b) Percent decrease of f KE.

mean total energy calculated above, and \bar{U} is the mean downstream advection. Using the values listed in Table 1a for the various parameters and scales, we obtain a geographical eddy conversion rate virtually identical to that reported by Rossby [1987]. However, our stream conversion rate is roughly half an order of magnitude larger than Rossby's [1987] (Table 1b). This latter difference is not surprising; at the Pegasus line the geographical mean Gulf Stream is 60% wider than the stream coordinate mean, and this is reflected in the difference between Rossby's [1987] geographical and stream conversion rates (Table 1b). Our two conversion rates are more similar because we invoke a finite difference using upstream information from Cape Hatteras, where both

TABLE 1a. Values of Parameters and Scales Used in Computing the Energy Conversion for the Two-Layer Representation, With Interface $\eta = A \tanh(y/w)$

Parameter	Value
A	250 m
H	500 m
f	$8.3 \times 10^{-5} \text{ s}^{-1}$
ρl	10^3 kg m^{-3}
$\Delta\rho$	1.5 kg m^{-3}
g'	0.015 m s^{-2}
\bar{U} (stream)*	55 cm s^{-1}
\bar{U} (geographical)*	40 cm s^{-1}
W (Cape Hatteras, stream)	23 km
W (Cape Hatteras, geographical)	28 km
W (Pegasus, stream)	33 km
W (Pegasus, geographical)	52 km

*Average between Cape Hatteras and Pegasus, where at each location the cross-stream averaged two-layer velocity was computed using the appropriate time-mean interface profile.

the mean and "instantaneous" Gulf Stream are substantially narrower than at the Pegasus line. The favorable comparison between Rossby's [1987] conversion rates and ours implies that the substantial energy losses in Figure 6 using the two-layer treatment are in fact realistic.

4. TWO-LAYER DESCRIPTION

Before considering the daily temperature and geostrophic velocity sections, it is informative to treat the Gulf Stream as a simple two-layer system with the 12°C isotherm as the bounding interface (as was done for the energy analysis). The in situ interface depth at each IES site can be easily extracted from the vertical temperature profiles (recall that this is equivalent to obtaining the interface depth empirically using the relationship of Watts and Johns [1982]). Important features of the Gulf Stream's meandering are effectively revealed in this two-layer framework. The dominant variability of the interface consists of simple lateral translations, i.e., the meandering of the current (see also Cronin and Watts [1989]). To reveal structural variations of the main thermocline (aside from meandering), Manning and Watts [1989] computed the temperature variance from a set of XBT sections across the Gulf Stream that were first aligned in a stream coordinate sense (equivalent to lining up the core of the current). Manning and Watts [1989] found that the remaining variability was dominated by what they termed a "transport" mode, i.e., a shoaling of the inshoremost part of the front associated with a deepening of the offshoremost part (and vice versa). This caused a change in overall transport of the Gulf Stream, but the central portion of the

TABLE 1b. Mean to Eddy Energy Conversion Rates

	Conversion Rate, $\text{erg cm}^{-1} \text{ s}^{-1}$	
	Geographical Coordinates	Stream Coordinates
This study	1.31×10^{10}	9.82×10^9
Rossby [1987]*	1.33×10^{10}	1.48×10^9

*For comparison, it was necessary to multiply Rossby's [1987] values (from his Table 1, sum of barotropic and baroclinic) by the basic state density (1.03 g cm^{-3}) and the cross-sectional area (from his Figure 4).

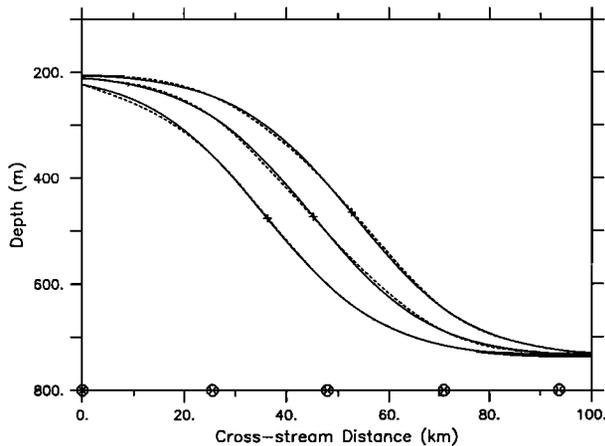


Fig. 7. Comparison of the ideal Gulf Stream interface (solid lines) to the cubic spline fit (dashed lines) as the Gulf Stream meanders. The spline fit was obtained using the values of interface height at the five IES locations (shown by the circled crosses). The pluses indicate the average frontal depth (explained in text).

main thermocline remained unchanged. The XBT sections used by *Manning and Watts* [1989] came from a variety of alongstream locations in the Gulf Stream, all downstream of Cape Hatteras.

We computed a daily cross-stream profile of the 12°C isotherm interface by fitting a natural cubic spline to the five IES interface depths. To make sure that the spline fit did not introduce any significant artificial structure we did the following test. First, we fit a hyperbolic tangent to the instantaneous Gulf Stream 12°C profile to obtain an ideal Gulf Stream front. Then we translated this ideal front across a line of five (artificial) IESs spaced the same as in the array, and then computed the corresponding interface time series at each site. These contrived time series are identical in character to those obtained by the actual instruments. Next we applied the cubic spline fit to each day of synthetic data, comparing the fit with the true Gulf Stream (i.e., the hyperbolic tangent) to assess the spline's accuracy. Figure 7 shows the favorable comparison between the ideal Gulf Stream and spline fit as the front translates ± 10 km (recall that the observed standard deviation at Cape Hatteras was ± 7.5 km). When the front approaches either end of the array the spline fit becomes significantly inaccurate near the edge. Unfortunately, when the Gulf Stream meandered significantly onshore, our array was insufficient to sample the entire current, so these periods are omitted from part of our analysis.

4.1. Frontal Shoaling and Deepening

The simplest conception of a meandering Gulf Stream is that of a rigid frontal structure shifting back and forth. If this characterization applies for the meanders at Cape Hatteras, then our IES data would indicate so, based on the results of the above test. However, our data reveal a systematic shoaling and deepening of the entire Gulf Stream front as one follows the meandering current in a stream coordinate sense (which is in contrast to remaining at a fixed location and observing the interface rise and fall, which does happen even for a rigid Gulf Stream). This effect is not observed further downstream, and we argue below that it is a result of the

steep topography underlying the Gulf Stream at Cape Hatteras.

Figure 8 shows three "snapshots" of the observed Gulf Stream interface during an offshore meander. This is to be contrasted with Figure 7, which shows the translation of the ideal rigid front. One immediately sees that the three observed profiles will not collapse to a single profile when adjusted laterally, as does the ideal front. To quantify this effect, we computed the average interface depth of the front for each day (in a stream coordinate sense). In particular, we averaged over the width of the current where the frontal slope is $\geq 55\%$ of its maximum value (which thus excludes the flat shoulders on the outer edges of the front). In Figures 7 and 8 this average interface depth is indicated by a cross on the respective profiles. For the rigid Gulf Stream this depth remains the same as the current translates. In the observed Gulf Stream the depth changes by over 100 m as the Gulf Stream progresses offshore.

These vertical excursions of the Gulf Stream front occur systematically during the entire deployment: the front shoals as the current meanders offshore, then deepens again when it meanders onshore (Figure 9a). The average frontal depths determined analogously from the spline fit to the idealized rigid Gulf Stream show no such evidence of shoaling, which signifies that the observed effect is real. Interestingly this shoaling and deepening of the front does not occur further downstream in the Gulf Stream. The Pegasus data show no comparable trend. There are of course far fewer data points at the Pegasus line, so we did the same calculation at 70°W using 2 years of IES data; again, there is no significant cross-stream trend (Figure 9b).

4.2. Potential Vorticity and the Role of Topography

We now address the cause of the systematic shoaling and deepening of the Gulf Stream front revealed by the IES array, which seems to be unique to the region near Cape Hatteras. Figure 10 shows a schematic of the front overlying the bottom topography. As was mentioned previously, at this location the bottom slope is greater than the Gulf Stream

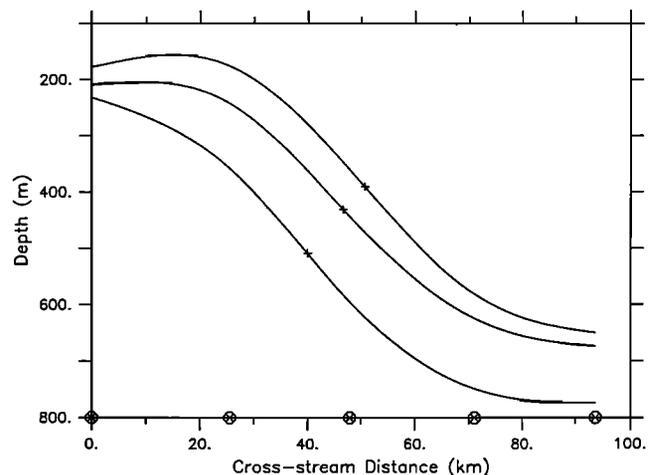


Fig. 8. Observed Gulf Stream interface at three times during an offshore meander. The onshoremost profile is year day 59; the middle profile, year day 63; and the offshore profile, year day 70 (year day 0 corresponds to October 14, 1987). The pluses indicate the average frontal depth.

frontal slope. Using moored data at 68°W, Hall [1986b] determined that deep cross-stream velocities under the Gulf Stream were often opposed the translation of the thermocline temperature front. That is to say, when the Gulf Stream front meandered offshore, the deep velocity had an onshore component (and vice versa). Johns and Watts [1986] found this to be true the majority of the time near Cape Hatteras. If in Figure 10 the deep column of water moves upslope in response to the front translating offshore, this allows for the column to maintain a constant thickness, or in terms of potential vorticity keep f/H constant. Since the bottom slope is so steep, however, the column can not move very far upslope without being compressed by the topography. We surmise that the inertia of the column causes it to move farther upslope and that to avoid being compressed, the column pushes up on the interface. The opposite would be true for onshore frontal translations, and this would result in the shoaling and deepening observed by the IESs. Note that this effect requires the bottom slope to be steeper than the frontal slope, as it is in the region near Cape Hatteras. Further downstream, where the bottom is nearly flat, the

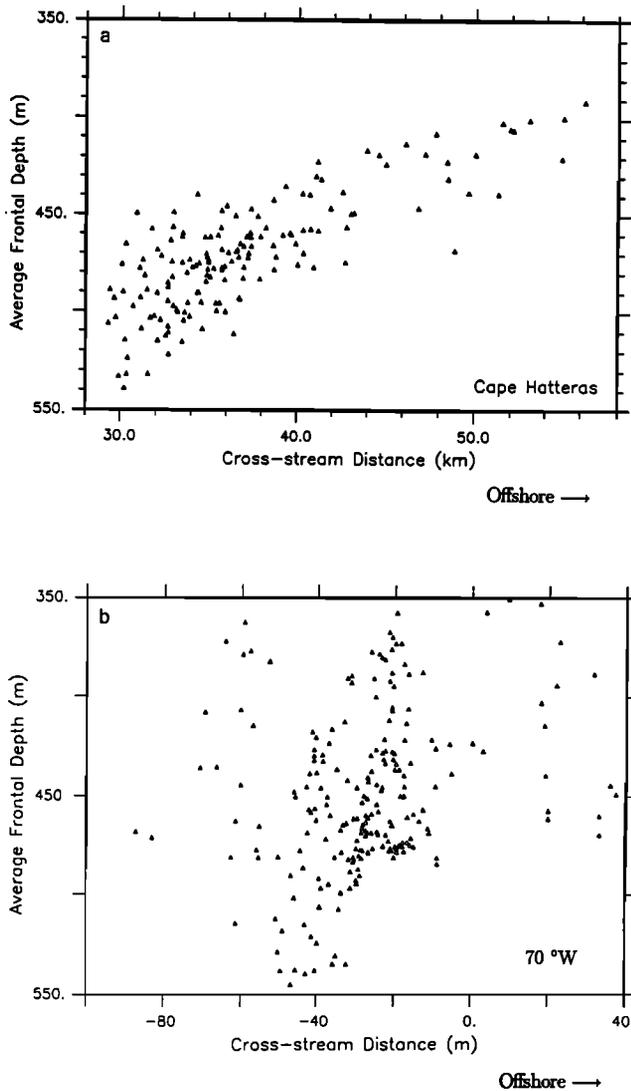


Fig. 9. Scatter plot of average Gulf Stream frontal depth versus cross-stream position of the Gulf Stream (a) at Cape Hatteras and (b) at 70°W.

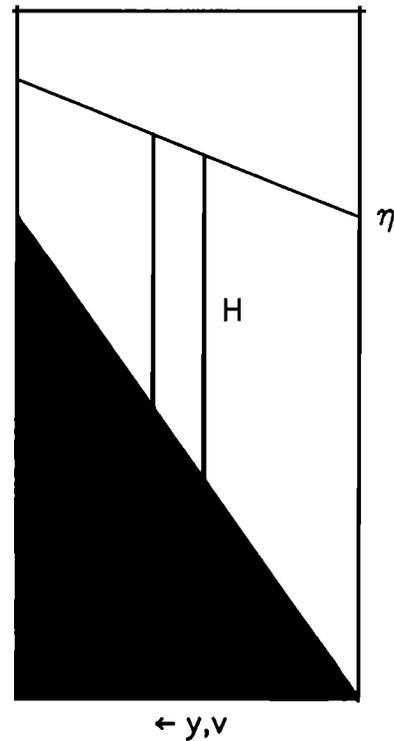


Fig. 10. Schematic showing the Gulf Stream interface at Cape Hatteras above the steep bottom topography. A deep water column bounded by the interface and bottom (of thickness H) is indicated.

deep water columns would experience stretching and compression as described by Hall [1986a].

The idea that the thermocline is vertically displaced in response to deep fluid columns preserving their layer thickness can be checked for consistency using the deep current meter data in conjunction with the IES data. In the two-layer formulation we can calculate the relevant terms in the lower layer potential vorticity equation. The IESs provide the depth of the interface, the bottom topography is known, and hence the layer thickness is determined. The bottom current meters measure the advection. The quasi-geostrophic potential vorticity equation (on an f plane) for the lower layer is

$$\frac{\partial}{\partial t} \left(\zeta - \frac{f}{D} \eta \right) = -v \frac{\partial}{\partial y} \left(\zeta - \frac{f}{D} H \right) - u \frac{\partial}{\partial x} \left(\zeta - \frac{f}{D} H \right), \tag{6}$$

where x and u are distance and velocity in the downstream direction, i.e., along the mean axis of the Gulf Stream (32°T, which is also alongslope), and y and v are cross-stream (upslope) distance and velocity; ζ is the relative vorticity, η is the interface depth, H is the layer thickness (interface minus bottom depth; see Figure 10), f is the Coriolis parameter ($= 8.4 \times 10^{-5} \text{ s}^{-1}$), and D is the average layer thickness ($= 2500 \text{ m}$). When expressed in natural coordinates $\zeta = -\partial V/\partial n + V/R$, where V is the speed, n is the instantaneous normal to the Gulf Stream (directed to the left of the flow), and R is the radius of curvature. Note that since there are IESs at three alongstream locations, we can measure R (Figure 2).

Equation (6) can be simplified on the basis of the following observations:

1. As was mentioned previously, the Gulf Stream path meanders very little at Cape Hatteras, and R is typically 500 km, which is an order of magnitude larger than the cross-stream length scale. Thus all terms involving the curvature vorticity can be neglected. Correspondingly, $-\partial V/\partial n \approx -\partial u/\partial y$.

2. Because x is oriented alongslope, $\partial H/\partial x = \partial \eta/\partial x$. Also, u and v are comparable in magnitude according to the bottom current meter records. Therefore downstream advection of thickness vorticity can be neglected relative to the cross-stream advection if $\partial \eta/\partial x \ll \partial H/\partial y$. The latter can be calculated directly using our data, and its average value is 6.8×10^{-3} . The former can be estimated by comparing the average η across the Gulf Stream at Cape Hatteras to that at the Pegasus line. We find that η shoals by ~ 15 m over this 100-km distance, so that $\partial \eta/\partial x = 1.5 \times 10^{-4}$. Thus downstream advection of thickness vorticity can be neglected.

Equation (6) is thus simplified to

$$\frac{\partial}{\partial t}(-u_y) + \frac{\partial}{\partial t} \left(\frac{-f}{D} \eta \right) \sim -v \frac{\partial}{\partial y}(-u_y) - v \frac{\partial}{\partial y} \left(\frac{-f}{D} H \right) - u \frac{\partial}{\partial x}(-u_y). \quad (7)$$

All except the last term on the right-hand side of equation (7) can be measured directly using the IES and current meter data at the central instrument line. If the deep water columns are preserving their thickness, then the following balance will apply,

$$\frac{\partial}{\partial t} \left(\frac{-f}{D} \eta \right) \sim -v \frac{\partial}{\partial y} \left(\frac{-f}{D} H \right). \quad (8)$$

Since we are interested in the dynamics of Gulf Stream fluctuations (meanders) in the deep layer, we must be careful, however, when testing this balance using the data. Recall that our array is located precisely where the equatorward DWBC crosses under the Gulf Stream (Figure 2). In addition, bottom-trapped topographic Rossby waves are superposed on the deep flow here (having been generated farther offshore). Despite these additional deep currents, however, the Gulf Stream does influence the bottom flow, and on occasion the deep transport here reverses to the northeast [Pickart and Watts, 1990a]. To investigate the Gulf Stream contribution, it is necessary to restrict ourselves to a particular range of time scales. The dominant topographic wave signal observed during the mooring deployment had a period of ~ 40 days [Pickart and Watts, 1990a]. Because the waves are so energetic it is difficult to discern anything about the deep Gulf Stream at these periods. However, the topographic waves have a high-frequency cutoff of roughly 10 days in this region [Johns and Watts, 1986], so fluctuations at periods shorter than 10 days will be devoid of the wave signal. Indeed, Johns and Watts [1986] observed that Gulf Stream meanders near Cape Hatteras were coherent to the bottom on these short time scales.

For this reason we consider motions only in the 4- to 8-day band. It is probable that no significant DWBC variability exists at such short time scales as well. We evaluated equation (8) at the center of the Gulf Stream (location B3 in Figure 2) using simple finite differencing for the derivatives,

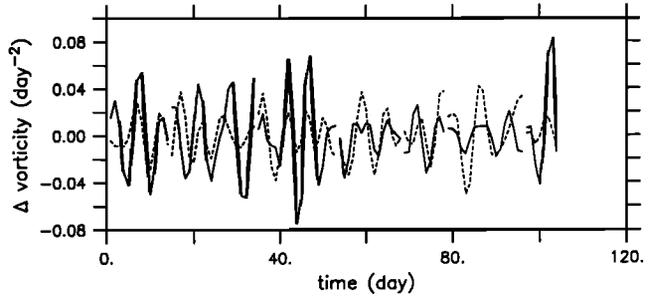


Fig. 11. Comparison of time rate of change of thickness vorticity at IES site B3 (solid line) with cross-stream advection of thickness vorticity (dashed line) in the 4- to 8-day band. Events are indicated by breaks in the curves. At a period of 6 days the two time series have a coherence of 0.47, which is significant at the 99% confidence level ($=0.39$), and a phase of -5.6° .

and then band-passed the two terms. Recall that for a deep water column to preserve its thickness during a meander, it must move oppositely to the translation of the interface (as visualized in Figure 10). Since we have time series of interface height (η) and deep cross-stream velocity (v), we can check to see if this is the case. It turns out that in the 4- to 8-day band it is true roughly half the time; i.e., during some events the deep cross-stream flow is oppositely directed to the translation of the front, and during other events the flow moves with the front. This is much like Hall [1986b] observed further downstream. In evaluating (8), therefore, we distinguished between these two regimes.

Figure 11 shows the comparison of the two band-passed vorticity terms in equation (8) during times when the bottom flow opposes the translation of the Gulf Stream front. One sees that the cross-stream advection of layer thickness does tend to compensate local changes in interface height. The agreement is by no means perfect, but at a period of 6 days the two time series are significantly coherent at the 99% confidence level, and in phase. Part of the disagreement in Figure 11 is probably due to computational error (e.g., comparing finite differences).

This result is in sharp contrast to the identical comparison for events when the deep flow is in the same direction as the translation of the front (not shown). In that case there is no tendency for the two terms in equation (8) to balance each other, and the time series are not significantly coherent at the 6-day period. Figure 12 shows the scatter plot of Gulf Stream frontal depths versus cross-stream position (analogous to Figure 9a) except that the time series was band-passed and divided into "opposite" and "same" events. One sees that the frontal shoaling and deepening occurs only during the opposite events. Thus Figures 11 and 12 together imply that the tendency for deep water columns to preserve their thickness causes the Gulf Stream interface to be displaced vertically during meanders. This happens when the deep columns move oppositely to the translation of the temperature front, which on short time scales (<10 days) occurs roughly half the time. Inspection of the non-band-passed frontal depths (Figure 9a) indicates that it happens the majority of the time at longer time scales (though some of the observed scatter may also be due to same events).

One might expect that when the remaining measurable terms in equation (7) are included on the right-hand side of (8), the agreement in Figure 11 would improve. This is not

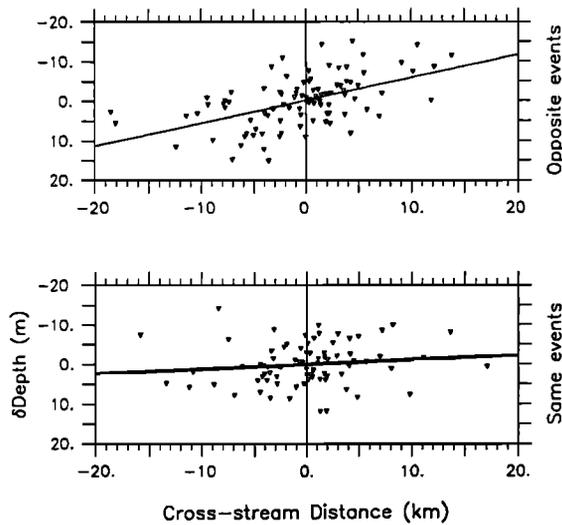


Fig. 12. Scatter plot of band-passed Gulf Stream frontal depth versus cross-stream position of the Gulf Stream, for (top) deep cross-stream velocity directed oppositely to the translation of the temperature front and (bottom) deep flow in the same direction as the frontal translation. The solid lines are the least squares fits.

the case, however. Similarly, if these terms are included for the same events, there is still no correlation. This seems to imply that downstream advection of relative vorticity (not measurable) is required to complete the balance. However, it may be that the other measurable terms are simply too noisy to extract a meaningful signal. Clearly, these dynamical balances need to be studied further, although we apparently have determined the cause of the observed frontal shoaling and deepening.

5. CONTINUOUSLY STRATIFIED GULF STREAM

We return to the IES-derived temperature and geostrophic velocity sections described in section 3, which enable us to look beyond the two-layer description and consider the implications of the frontal shoaling and deepening on the

fully stratified Gulf Stream. Figure 4a showed the mean geostrophic velocity section overlaid on the mean temperature section. We have a similar pair of sections for each individual day, and in Figure 13 we show two such pairs: year day 175, when the Gulf Stream was located onshore of its mean position, and year day 65, when it was offshore. One sees that there are pronounced differences in the Gulf Stream at the two locations, most notably, that the main thermocline is compressed and the vertical shear of velocity is stronger when the Gulf Stream is offshore. We believe that these changes are consequences of the frontal shoaling described above. Simply, if the front is pushed upward as it translates offshore, this causes the thermocline to compress, which in turn increases $\partial T/\partial x$ (and hence $\partial u/\partial z$). Using the temperature sections, we can calculate the width of the thermocline (at the center of the Gulf Stream) as a function of time. We find that it varies with onshore-offshore position of the Gulf Stream by roughly 100 m, which is comparable to the vertical extent of shoaling described above. It should be noted that there is no such cross-stream trend in thermocline thickness observed at the Pegasus line (as there was no shoaling).

Upstream of Cape Hatteras numerous experiments [e.g., Webster, 1961; Bane *et al.*, 1981; Brooks and Bane, 1981] have revealed that Gulf Stream meanders are “skewed”: onshore meanders occur more slowly than offshore meanders. In addition, when the current moves onshore it has a sharper temperature front than when it meanders offshore. This frontal tilting is associated with a compaction of the isotherms as the current meanders onshore, so one may wonder if this is related to the effect observed at Cape Hatteras. Most likely it is not: lateral displacements of the Gulf Stream at Cape Hatteras are not skewed (they are more sinusoidal), and the sense of thermocline compaction is opposite at Cape Hatteras; i.e., the isotherms compact when the current meanders offshore. This is not to say that the shoaling observed at Cape Hatteras can not occur upstream, but the steepest bottom slope occurs right at Cape Hatteras where the Gulf Stream flows into deeper water.

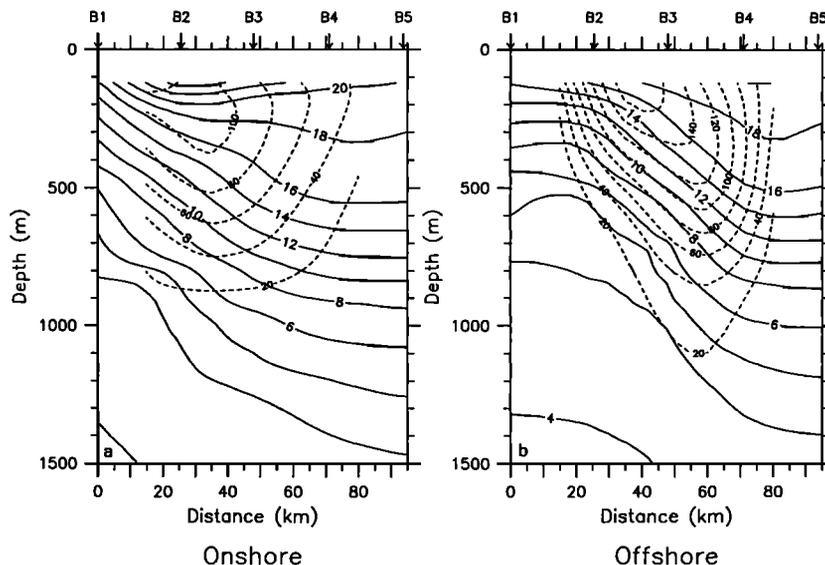


Fig. 13. IES-derived temperature section (solid lines) overlaid on the geostrophic velocity section (dashed lines) for (a) year day 175, when the Gulf Stream was onshore, and (b) year day 65, when the Gulf Stream was offshore.

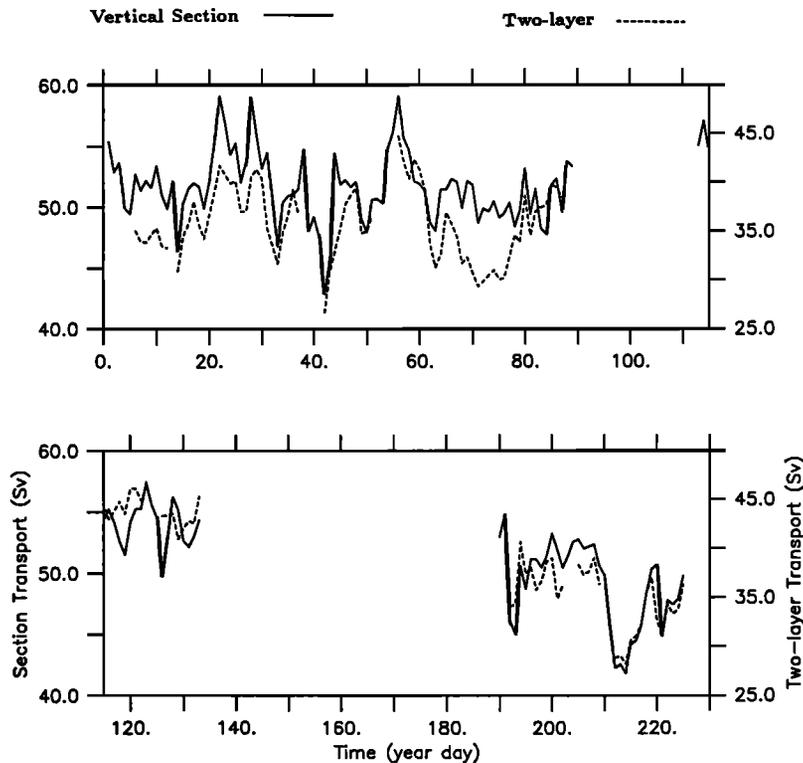


Fig. 14. Gulf Stream transport obtained from integrating the geostrophic velocity sections (solid line) compared with that obtained from the two-layer formulation (dashed line). Gaps in the time series are when the Gulf Stream meandered partially out of the array.

The corresponding change in $\partial u/\partial z$ due to the compression of the thermocline at Cape Hatteras results in an increase in core velocity at 100-m depth of roughly 50 cm s^{-1} as the Gulf Stream meanders offshore. Note that the earlier two-layer representation is not adequate to describe vertical compression of the thermocline. It is of interest then to compare the Gulf Stream transport computed by integrating the IES velocity sections, with the baroclinic transport obtained using the interface profiles. In the two-layer model this transport is equal to $(g'/2f)(d_{\text{off}}^2 - d_{\text{on}}^2)$, where d_{off} is the upper layer thickness at the offshore edge of the Gulf Stream and d_{on} is the thickness at the onshore edge. Figure 14 shows the comparison between the vertical section transport and the two-layer transport. As was mentioned earlier, when the Gulf Stream meandered substantially onshore, it passed partly out of our array and we were unable to compute transports; hence the gaps in Figure 14. One sees, however, that the two time series are highly correlated, indicating that a two-layer description of the Gulf Stream is adequate for capturing the dominant variability in transport. Thus while the vertical compression of the thermocline strongly influences the shallow core velocity of the Gulf Stream, it has little effect on overall transport.

6. SUMMARY

An analysis of IES data from the Gulf Stream near Cape Hatteras has revealed that meanders in the current there are not simply translations of a rigid temperature front. Rather, as the Gulf Stream moves offshore, the thermocline shoals and correspondingly compresses. These systematic variations of the thermocline are seen in daily temperature

sections over an 8-month period which were constructed from the IES data, following the methodology of *Pickart and Watts [1990b]*. A comparable set of geostrophic velocity sections indicates that the geostrophic vertical shear is increased as the thermocline compresses, which significantly influences the upper core speed of the Gulf Stream but not its overall transport.

A similar analysis of historical data collected further downstream indicates that the shoaling of the Gulf Stream front is unique to the region near Cape Hatteras and related to the presence of the steep topography there. As the front translates offshore, deep water columns progress onshore (upslope) and, to avoid being compressed, displace the interface upward. This effect can only occur when the topographic slope is steeper than the frontal slope, which is true near Cape Hatteras where the Gulf Stream flows over the continental slope.

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