Gulf Stream structure, transport, and recirculation near 68°W

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Abstract. An analysis of the structure and transport of the Gulf Stream is undertaken using direct current meter observations from a 13-mooring array deployed near 68°W from June 1988 to August 1990. The analysis is based on a "stream-coordinate" approach, in which velocities are rotated into a local, downstream coordinate frame and averaged according to their relative cross-stream location within the current. The picture so obtained represents the average synoptic structure of the Gulf Stream, rather than the Eulerian-averaged structure in which the current is weakened and broadened by lateral meandering of the current and adjacent recirculations. Many familiar features of the Gulf Stream are reproduced in the analysis, including an asymmetric velocity profile with larger shear on the cyclonic (shoreward) side of the current, an offshore displacement of the velocity core with depth, and a subsurface velocity maximum on the offshore side of the current. Westward recirculations are also seen on both sides of the Gulf Stream. Maximum downstream speeds at the axis of the Gulf Stream reach approximately 2.0 m/s at the surface and 0.7 m/s at 1000 m, roughly twice the corresponding Eulerian-averaged values. The analysis also reveals a deep extension of the Gulf Stream at 3500 m depth with a width of 130 km and average speeds of 3–4 cm/s. The transport of the Gulf Stream in the stream-coordinate frame is 113 ± 8 Sv, approximately 30% larger than the Eulerian-averaged transport of 88 Sv. On the basis of these results and other recent studies the downstream transport increase of the Gulf Stream and the inflow structure to the Gulf Stream are reconsidered. It is concluded that approximately 30 Sv, or over half of the transport increase between Cape Hatteras and 68°W, is fed by inflow from the northern side of the Gulf Stream and that this inflow is concentrated near Cape Hatteras and 68°W, where the Gulf Stream flows steeply across isobaths converging from the north.

1. Introduction

Downstream of Cape Hatteras, the transport of the Gulf Stream is known to increase by more than a factor of 2 due to recirculations on both its northern and southern sides [Worthington, 1976; Halkin and Rossby, 1985; Hogg, 1983]. Knauss [1969] showed that in the first 1000 km downstream of Cape Hatteras the Gulf Stream transport increases from 65 Sv (1 Sv = 10⁶ m³/s) to approximately 150 Sv (at 65°W), an average rate of increase of 8 Sv per 100 km. This large transport, of order 150 Sv, is apparently maintained at least as far as 55°W [Hendry, 1982; Hogg, 1992]. Despite this large increase in transport, hydrographic sections show that the Gulf Stream has a very similar baroclinic structure over this domain, in terms of its horizontal and vertical scales and total density contrast across the current. The evolution of the Gulf Stream downstream of 55°W is not well understood at this time, but near the Grand Banks, there appears to be a distinct change in the character of the Gulf Stream, from a single meandering front to multiple, branching fronts feeding into the Azores and North Atlantic Currents [Krauss, 1986]. The region east of 55°W is also thought to be where most of the detrainment from the Gulf Stream to the bordering recirculations occurs.

It is clear from the available data that much of the downstream transport increase between Cape Hatteras and 55°W is due to an increase in deep velocities beneath the Gulf Stream. In fact, Hogg [1992] found that the transport of the Gulf Stream relative to 1000 m was essentially constant between Cape Hatteras and 55°W, implying that the entire transport increase in this region is due to a growing barotropic component. A point of controversy that remains is from which side of the Gulf Stream this barotropic flow is fed and how the contributions from recirculations to the north and south of the stream vary along the stream's path. Worthington's [1976] circulation scheme had a maximum Gulf Stream transport of 150 Sv near 60°W but indicated that all of the transport increase was fed from a recirculation gyre south of the Gulf Stream, the so-called "Worthington Gyre." More recent measurements [Richardson, 1985; Hogg, 1992] have confirmed that a large recirculation also exists to the north of
the Gulf Stream near 55°W, of roughly equal strength (approximately 40 Sv) as the Worthington Gyre. The western limit of this Northern Recirculation Gyre is not precisely known, although most of its transport is thought to join into the Gulf Stream east of the New England Seamount chain, which cuts across the Gulf Stream near 65°W [Hogg, 1983, 1992].

The large transport increase downstream of Cape Hatteras is closely paralleled by an increasing meander envelope, which grows from about 20 km at Cape Hatteras to 300 km near 65°W [Watts, 1983]. It is generally believed that the downstream transport increase and the associated recirculations are driven in part by finite-amplitude eddy processes related to dynamical instabilities of the Gulf Stream system. To better understand these eddy–mean flow interaction processes, a large observational program (Synoptic Ocean Prediction (SYNOP)) was undertaken during 1987–1990 between Cape Hatteras and the Grand Banks. Two extensive moored arrays were set near 68°W (the SYNOP "central" array) and 55°W (the SYNOP "eastern" array), which were intended to characterize the Gulf Stream structure, transport, and eddy–mean flow interaction in regions upstream and downstream of the New England Seamount chain. Initial results from these moored arrays are presented in articles by Hogg [1993], D. R. Watts et al. (Gulf Stream path and thermocline structure near 74°W and 68°W, submitted to Journal of Geophysical Research, hereinafter referred to as Watts et al., submitted manuscript, 1994), and T. J. Shay et al. (Gulf Stream flow field and events near 68°W, submitted to Journal of Geophysical Research, hereinafter referred to as Shay et al., submitted manuscript, 1994).

The purpose of this paper is to provide a detailed account of the structure and transport of the Gulf Stream near 68°W, derived from the direct current meter observations collected in the SYNOP central array. The analysis presented here relies heavily on so-called "stream-coordinate" techniques of averaging, for which a number of methods have been developed in recent years using both repeat section data [Halkin and Rossby, 1985] and fixed-point current meter measurements [Hall and Bryden, 1985; Hogg, 1992]. The purpose of these methods is to remove the smearing effects of meandering on the time-averaged velocity structure of the current. To accomplish this, the observed currents are rotated into a coordinate frame aligned with the instantaneous path of the Gulf Stream and then averaged according to their relative cross-stream location within the current. The picture so obtained may be thought of as representing the average instantaneous or synoptic structure of the Gulf Stream at a certain location. A strength of this technique is that partial sections can be combined to describe the average synoptic structure, even if they do not individually cross the full width of the current. In the case where counterflows or recirculations are present in close proximity to the Gulf Stream this method also allows the structure of these flows to be investigated apart from that of the main current. This is important because in a simple geographical (Eulerian) average the effect of these counterflows will be to decrease the apparent strength of the Gulf Stream, causing its transport to be underestimated. Similarly, the counterflows may be obscured or disappear entirely in an Eulerian average. A further advantage of this technique is that it allows a clearer interpretation of cross-stream velocity patterns related to inflow or outflow from the Gulf Stream. Halkin and Rossby [1985] were among the first to use stream-averaging methods to describe the structure and transport of the Gulf Stream. Their results, based on a total of 16 absolute velocity sections collected across the Gulf Stream near 73°W, showed a remarkably robust Gulf Stream structure, with well-defined lateral boundaries and with significant inflows on both sides of the stream. They were also able to show that the velocity variance due to structural changes in the Gulf Stream, after meandering effects were removed, was only a few times larger than typical mid-ocean values. Similar methods have been applied to moored current meter data, relying on the fact that in a meandering current a fixed site will occupy various locations relative to the current axis. In these methods the instantaneous down-stream direction is usually determined by the direction of the vertical shear between two or more measurement levels spanning the thermocline. Temperature measurements (either the temperature at a fixed depth [Hall and Bryden, 1985] or the pressure of a given isotherm based on a fit to temperature data at two or more levels [Hogg, 1992]) are then used to determine the cross-stream location relative to the current axis.

A similar approach is used here, although a different method is used for the cross-stream mapping of the velocity field, which incorporates data from an array of inverted echo sounders (IES) deployed simultaneously with the current meter array. Whereas in the past these methods have typically been applied to only one or two current meter moorings in a given region, in this study there are a total of 13 moorings available for a period of over 2 years. The large number of moorings provides an excellent sampling distribution of the current structure and allows a highly resolved and statistically robust picture of the mean flow to be constructed. In what follows we will first describe the data and methods used (section 2), including some comparisons with results obtained using more conventional techniques (appendix). Results are then presented in section 3 on the mean structure and transport of the Gulf Stream, the observed inflow patterns, and the magnitude of the flanking recirculations (primarily on the northern side of the Gulf Stream). These results are then compared with the corresponding results obtained in a fixed (Eulerian) reference frame in section 4. Comparisons with other regional observations and implications on the inflow structure and downstream transport increase of the Gulf Stream are given in section 5, followed by a summary of results in section 6.

2. Data and Methods

Current Meter Observations

The SYNOP central current meter array (Figure 1) was set in two deployments, each of roughly 1-year duration, beginning in June 1988. The moorings were organized along three sections referred to as the G, H, and I lines, containing two, five, and five moorings, respectively. These 12 mooring sites were maintained for a total of 26 months, with the individual mooring turnarounds being accomplished on a series of cruises in summer 1989 such that each site was occupied continuously through August 1990. A thirteenth mooring (M13) was added between the H and I lines from June 1989 to August 1990. Conventional current meter and temperature measurements were collected at nominal depths of 400 m, 700 m, 1000 m, and 3500 m; additional velocity profiles in the
upper 350 m were available from upward looking acoustic Doppler current profilers (ADCPs) at the following three sites: H3, H4, and I2 [Johns and Zantopp, 1991].

Data recovery from the array is summarized by Shay et al. (submitted manuscript, 1994), who also provide a more complete description of the current meter data. The reader is referred to Shay et al. [1994] for details on processing of the conventional current meter data, and to Johns and Zantopp [1991] for a similar discussion of the moored ADCP data. For the purposes of this paper all records were low-pass filtered with a 40-hour Butterworth filter passed forward and backward over the data (to eliminate phase shifts) and subsampled at 1-day intervals. The velocity and temperature measurements were then corrected to constant depths using a modified form of Hogg's [1991] mooring correction scheme [Cronin et al., 1992], and the ADCP profiles were interpolated to constant depths at 50-m intervals over the top 350 m.

Mean velocity vectors for the total experiment are shown in Figure 2, for the 400 m and 3500 m levels. At 400 m the average Gulf Stream can be seen as a moderately strong eastward flow centered between 37°N and 38°N, with maximum mean velocities of about 0.7 m/s. At the northern side of the array the influence of the Gulf Stream begins at about 38.5°N and extends from there to the southern end of the array, where mean velocities still show significant eastward flow components of 0.1-0.2 m/s. Thus in a Eulerian sense the array did not completely span the Gulf Stream. North of 38.5°N, at moorings H2 and I1, the mean flow at 400 m is westward and clearly defines the northern limit of the mean Gulf Stream at this depth. A similar picture holds at the 700-m and 1000-m levels (Shay et al., submitted manuscript, 1994), but with a gradual offshore shift of the eastward flow pattern with increasing depth.

At 3500 m the situation is entirely different; there is a general southwestward flow over the array of magnitude 4-7 cm/s with no clear indication of a deep Gulf Stream. In the southeastern corner of the array (sites I4 and I5) there is, in fact, an eastward component of mean flow, although this feature, as well as the more southerly orientation of the flow vectors in the southwestern part of the array, may be related in part to local topographic influences and to deep cyclonic circulations induced by large meander troughs that frequently develop over this region (Shay et al., submitted manuscript, 1994).

A cursory look at Figure 2 illustrates the difficulties that can be encountered when trying to make estimates of even basic properties such as Gulf Stream transport from Eulerian-averaged data. Despite its rather large meridional extent the array did not fully resolve the boundaries of the downstream flow, and the horizontal resolution across the flow is only barely adequate for transport estimates. Thus a more adaptive approach, such as stream-coordinate averaging, is needed to resolve the Gulf Stream's structure and transport.

Stream-Coordinate Averaging Procedure

Definition of downstream direction. For moored applications the downstream direction is traditionally defined as the direction of the vertical shear between measurement levels spanning the thermocline [Hall and Bryden, 1985]. For a flow in geostrophic balance this direction is parallel to contours of dynamic height integrated over the layer in question and therefore provides a good measure of the orientation of the main baroclinic front.

The standard current measurement levels available on all of the moorings for calculating the shear-derived downstream direction are 400 m, 700 m, 1000 m, and 350 m. The 400-m level is the obvious choice for the upper level, while either 1000 m or 350 m could be used for the deeper level. The 3500-m level is preferable from a signal standpoint, in that the velocity differences between this level and the 400-m level are generally larger, and thus there is less uncertainty in determining the direction of the vertical shear due to...
Figure 2. Average currents at (top) 400 m and (bottom) 3500 m for the total deployment period. All of the vectors
represent 26-month averages, except at sites H5 and M13, which are approximately 1-year averages.

The above definition works well through the swiftly flowing portion of the Gulf Stream but can become unreliable
near the edges of the stream for two reasons. First, owing to the weak vertical shears there, the downstream directions
are noisy, and second, if any vertically sheared counterflows are present, they can be mistakenly interpreted as part of the
downstream flow. For example, a westward flow which decreases with depth would, in this scheme, be assigned a westward “downstream” direction, and the projections onto this would yield an apparently positive downstream flow, which is clearly undesirable. Such situations can be expected to occur frequently in this area associated with Gulf Stream rings and/or weakly sheared westward flows in the flanking recirculations. One way to eliminate this problem is simply to discard any of the daily estimates with “westward” downstream directions. This is an effective way to eliminate rings but can also lead to a biased representation of the average flow near the edges of the Gulf Stream and in the adjacent recirculations.

To overcome this difficulty, a new method was developed to determine the instantaneous downstream direction for sites on the edges of the Gulf Stream and for all situations in which the flow at a given site was not clearly within the Gulf Stream. This method, hereinafter referred to as the IES method, utilizes daily objective maps of the 12°C isotherm depth topography $Z_{12}$ available over a 320 x 340-km area surrounding the current meter array, derived from an array of 24 IESs deployed simultaneously with the current meter array (Watts et al., submitted manuscript, 1994). The methodology for converting IES travel times to $Z_{12}$ estimates and details on the objective mapping procedure are described by Tracey and Watts [1991].

Figure 3 shows examples of the IES method for two particular days. The Gulf Stream is identified in these objective maps by the steeply sloping thermocline surfaces between $Z_{12} = 300$ and 800 m. The 400-m surface lies within the most steeply sloping portion of the baroclinic front and is chosen to represent the nominal axis of the Gulf Stream. The IES method generates a set of curvilinear coordinates for each site relative to this axis as follows. For each day the minimum vector $r$ from each current meter site to the 400-m contour is determined, which, by definition, is normal to the 400-m contour at its closest approach to the current meter site. The downstream direction is then taken as perpendicular to $r$, with the proper downstream sense being determined by whether $Z_{12}$ at the current meter site is shallower or deeper than 400 m. This method is more like the approach used by Halkin and Rossby [1985], in which all velocity measurements in a section are referenced to the downstream direction at the core of the Gulf Stream, rather than a locally defined downstream direction. An important advantage of this method is that rings and other eddy features on the edges of the Gulf Stream can be easily identified in the $Z_{12}$ maps, allowing the data from those sites to be excluded from the analysis.

In general, the downstream directions determined by the IES method and the 400- to 1000-m shear agreed very well with each other through the main part of the Gulf Stream, with typical differences of 5°. These small differences have a negligible effect on the downstream velocity projection. However, a comparison of the two methods showed that the IES method produced a larger scatter in cross-stream velocity estimates within the upper, central part of the Gulf Stream, where the downstream velocity magnitudes are large. The reason for this is apparently that the objectively
analyzed $Z_{12}$ maps yield a slightly smoothed representation of the thermocline topography, which cannot follow all of the smaller-scale path variations that may be present. The shear-derived direction is therefore believed to be a better indicator of the local downstream direction within the main body of the Gulf Stream.

On the basis of the above considerations the following approach was used for assigning daily downstream directions at each of the current meter sites: downstream directions derived from the 400- to 1000-m shear were used, provided that the velocity difference between these levels exceeded a threshold of 5 cm/s and the shear was in the "eastward" sense, otherwise, the downstream direction from the IES method was used. This procedure accomplishes the desired along-stream projection of strong flows within the Gulf Stream, while allowing weakly sheared flows on the flanks of the Gulf Stream, regardless of direction, to be appropriately averaged into a coordinate frame aligned with the instantaneous path of the Gulf Stream. The 5-cm/s threshold was chosen after some experimentation to be the minimum shear necessary to define an accurate downstream direction. Rings and other unrepresentative features on the edges of the Gulf Stream were flagged by computer checks, and these data were excluded from the analysis using a careful hand-editing procedure. This resulted in approximately 22% of the daily current meter observations being excluded from the final stream-coordinate averages.

**Definition of cross-stream coordinate.** Several methods were investigated for determining the cross-stream location of the daily current measurements. Most of these involved simple variations of the Hall and Bryden [1985] method (hereinafter referred to as the HB method), and these are described and briefly intercompared in the appendix. (Readers unfamiliar with the HB method may wish to consult the appendix before proceeding.) The curvilinear coordinate system generated by the IES method also provides an independent way to determine the cross-stream location of each current meter site. Here the magnitude of the vector $r$ from each current meter site to the axis of the Gulf Stream (the 400-m $Z_{12}$ contour) defines the instantaneous cross-stream coordinate, taken to be positive if $Z_{12}$ at the current meter site is less than 400 m (shoreward of the axis) and negative if $Z_{12}$ at the site is greater than 400 m (seaward of the axis). Each daily current meter observation is therefore associated at the outset with a specified cross-stream location, and it is then a straightforward procedure to sort the data at each observation level into cross-stream bins at a selected horizontal resolution and to average them. In all cases the initial sorting and averaging using the IES method was done at 5 km cross-stream resolution.

This method has two significant advantages over the HB type methods. First, the cross-stream mapping involves no assumptions about the dynamics; that is, the thermal wind relation does not need to be invoked in the mapping procedure. Therefore errors in the cross-stream mapping due to noisy estimates of vertical shear near the edges of the Gulf Stream are avoided, and weak barotropic flows adjacent to the Gulf Stream can be mapped in addition to the flow in the Gulf Stream itself. Second, as described earlier, it permits a careful quality control of the data points included in the stream-averaging procedure, which is difficult to do otherwise. A disadvantage of this technique is that it is not-self contained, relying on an independent data set with its own associated errors. Presumably, a similar mapping of the thermocline depth could be produced from the moored temperature measurements themselves, although this would also require a fairly extensive moored array, which is not ordinarily available for this type of analysis.

**Figure 3.** Stream-coordinate examples for (a) March 4, 1990, and (b) June 5, 1990. Contours are the depth of the 12°C isotherm $Z_{12}$ derived from the IES array. The solid, straight line segments show the normal distance from each current meter site to the axis of the Gulf Stream ($Z_{12} = 400$ m). In Figure 3a, site I1 in the northeast corner of the array is flagged and its data excluded from the stream coordinate analysis because of its proximity to the warm (deeper $Z_{12}$) feature just north of it.
As described in the appendix, the IES and HB mapping methods were evaluated on two main criteria, namely, their ability to (1) clearly define lateral boundaries of the downstream flow and (2) produce a robust synoptic structure with the lowest residual variance. On both of these counts the IES method proved to be the most successful and is therefore used in the final calculations and results presented in this work.

Figure 4 shows cross-stream profiles of velocity at 400 m and 1000 m obtained using the above stream-coordinate averaging procedure, at 5 km horizontal resolution. The number of daily current measurements in each 5-km bin and the number of these which satisfied the 5-cm/s threshold for defining a local (shear-derived) downstream direction are shown in Figure 5. The stream-averaged structure at 400 and 1000 m shows maximum core velocities of about 1.3 m/s and 0.3 m/s, respectively, with well-defined edges on both sides of the current and with significant westward recirculation north of the Gulf Stream. Standard deviations about this mean structure reach a maximum of about 0.3 m/s in the cyclonic shear zone at 400 m and decrease to typically 0.1 m/s or less near the edges of the current and at depth. For comparison, Figure A3 in the appendix shows the average velocity structure and standard deviation at 400 m obtained using a modified version of the HB method. (The same quality control was applied in both cases so that the resulting averages include the exact same data points.) It is shown there that the methods agree quite well through the central and shoreward parts of the Gulf Stream but that the HB method does not resolve the offshore edge of the current. The HB method also shows a larger overall variance about the stream-averaged structure compared with the IES method (see appendix for further discussion).

3. Results

Velocity Structure

Figure 6a shows the downstream velocity profiles obtained at all levels using the above stream-averaging method with the ADCP results subsampled to depths of 50 m and 200 m. The original 5-km bins have been smoothed with a
running 25-km boxcar filter and are truncated where these smoothed distributions contain fewer than 50 daily observations. The final distributions at 400 m and deeper span a total width of 270 km, while the ADCP distributions above 400 m span a smaller width of 155 km. Despite the more limited sampling in the upper 400 m the average structure there appears very consistent with the velocity structure at the thermocline levels. Near-surface (50 m) velocities reach a maximum of approximately 2.0 m/s at about 5 km offshore distance, indicating that, on average, the 400-m Z12 contour lies very close to the surface current axis. At deeper levels the velocity core shifts seaward so that by 1000 m depth the maximum is displaced about 35 km offshore of the surface core, which is a well-known feature of the baroclinic front. Also evident is the subsurface velocity maximum on the offshore side of the current, though its structure is not resolved beyond about 50 km by the ADCPs. This is associated with the warm core of surface waters advected from lower latitudes, resulting in a negative thermal wind shear on the offshore side of the current. The stream-averaging also reveals a deep Gulf Stream about 130 km wide centered beneath the anticyclonic portion of the surface flow, with nearly uniform speeds of 3-4 cm/s, a feature which is not at all obvious in the 3500-m Eulerian averages.

Westward recirculations are evident on both sides of the Gulf Stream. The vertical structure of the northern recirculation is essentially barotropic, with very little shear through the water column away from the immediate influence of the Gulf Stream (see also Figure 7). The recirculation south of the Gulf Stream appears to have considerably more shear, decreasing in strength below the thermocline; however, this could well be a near-field effect related to the offshore tilt of the Gulf Stream's velocity structure with depth and may not reflect the true vertical structure of the recirculation farther offshore.

Cross-stream velocities (Figures 6b and 7) show a consistent pattern of inflow from the northern side of the Gulf Stream and also suggest a weaker inflow from the southern side. Between 400 and 1000 m, spanning the thermocline, the convergence of the cross-stream flow component (\(\partial V_y/\partial y\)) is concentrated in a 100-km-wide zone centered near the core of the Gulf Stream (from about -50 to 50 km). The upper level ADCP data show a similar pattern but with larger amplitude. At 3500 m the convergence is spread out over a larger range but interestingly shows largest convergence at about 50-100 km offshore of the surface axis, centered within the deep axis of the Gulf Stream. The vertical structure of the inflow appears to be reversed on the two sides of the Gulf Stream, (i.e., the northern inflow decreases in strength below the thermocline, while the southern inflow increases at depth). Within the error bars, however, either profile could be barotropic (see error discussion in the next section). It should be noted that within the central portion of the Gulf Stream the cross-stream flow in the thermocline, specifically at the 400-m and 1000-m levels, is forced to be barotropic due to the shear-derived definition of the downstream direction. Thus in Figure 6b one can see that the 400-m and 1000-m cross-stream velocities are identical across the core of the Gulf Stream from about -100 km to 50 km. However, on the edges of the Gulf Stream, where the IES method is predominantly used to determine the downstream direction, this constraint is removed.

Statistically, the northern inflow of 3-4 cm/s is highly significant, while the southern inflow is significantly different from zero only at the 3500-m level due to larger scatter and fewer data points available on the offshore side of the Gulf Stream. At each level, however, there is an approximately constant difference of 5 cm/s between the cross-stream velocity components on the northern and southern edges. The size of this inflow convergence is somewhat surprising, in view of other considerations, and will be taken up later in the discussion.

Error Analysis: Streamwise Correlation Functions

Before proceeding further it is necessary to discuss uncertainties in the stream-averaged results. For each cross-stream bin the mapping procedure results in a collection of daily observations from various times and locations throughout the array. To estimate the standard error of the mean value in each bin, some method for determining the degree of independence of the data points is needed. A correlation analysis is traditionally used for this purpose; however, neither a strict Eulerian nor Lagrangian correlation function is appropriate owing to the mixed nature of the space-time sampling. What is needed is a measure of the decorrelation timescale of the flow at a specific cross-stream location.
within the Gulf Stream or, in other words, the timescale on which the current structure itself changes over a spatial domain the size of the current meter array. Intuitively, one might expect this to be longer than the Eulerian timescale, which is dominated in the upper water column by lateral meandering of the Gulf Stream. We are unaware of any previous attempts at such a calculation; presumably, this is one of the few data sets available for which this might be feasible. The results seem potentially interesting in their own right, in addition to their relevance for statistical purposes, and are therefore briefly discussed.

Temporal correlation functions were calculated for a variety of selected cross-stream bins at various depths. For each bin, time series were formed from all available observations from the complete array. At 5-km cross-stream resolution these binned time series are typically quite gappy and can have long periods when no data are acquired for that particular cross-stream location, causing the correlation functions to be rather noisy. Increasing the bin width helps to fill out these time series but can also introduce variance due to smearing of the mean structure in regions where cross-stream gradients are large, tending to artificially reduce the true decorrelation timescale. The original 5-km bin resolution was therefore retained. Smoother correlation functions were produced by ensemble averaging the correlation functions from several nearby 5-km bins. This procedure was carried out at each depth and for five regions, corresponding to the velocity core, the cyclonic and anticyclonic shear zones, and the northern and southern edges of the Gulf Stream. Ensemble averaging over five adjacent bins (25 km) was found to be sufficient to yield reasonably smooth correlation functions for each of these regions.

Figure 8 shows examples of these "streamwise" correlation functions for the downstream velocity component at depths of 400 m, 1000 m, and 3500 m, near the center of the Gulf Stream. The structure of these functions differs from that of a typical space- or time-lagged correlation function, mainly near the origin, where in this case there is a steep, rather than Gaussian, decay in the correlation at short time lags due to added spatial decorrelation. Significant positive correlations occur out to lags of typically 10 or 20 days. Estimates of the integral (decorrelation) timescale $\tau_i$ were made by integrating the correlation functions out to lags of 100 days and recording the maximum value occurring over that range. The integral timescales for the downstream velocity range from about 10 days in the thermocline to roughly 5 days at 3500 m, generally decreasing steadily with depth (Figure 8). Little variation in $\tau_i$ was found with respect to cross-stream location at any of the levels. Timescales for the cross-stream velocity (not shown) were typically shorter, probably due in part to larger measurement noise. The overall impression gained from this analysis is that the streamwise decorrelation timescale is comparable to Eulerian timescales in the Gulf Stream. This is, perhaps, not
surprising in a large-amplitude meandering region such as
the present one, where strong deformations of the frontal
structure are known to occur in association with meandering
(Watts et al., submitted manuscript, 1994).

For the purpose of error estimation we chose to apply a
constant 7.5-day integral timescale throughout; that is, 15
daily observations are required to gain one additional degree
of freedom [Owens, 1991]. Standard errors about the mean
structure are then calculated based on the estimated degrees
of freedom in each cross-stream bin. For example, in Figure
4 there are typically 100 or more data points in each 5-km
bin, and the standard errors about the mean structure shown
there are roughly a factor of 3 lower than the standard
deviations, ranging from a maximum of about 10 cm/s in the
cyclonic shear zone at 400 m depth to less than 4 cm/s on the
edges and at the deeper levels. The profiles shown in Figure
7 are averaged over five adjacent bins and, consequently,
have further reduced errors. For transport errors (see next
section) we have assumed the standard errors at individual
measurement levels to be uncorrelated.

Volume Transport

To obtain transport estimates from the stream-averaged
velocity structure, two types of interpolation/extrapolation
were needed. First, it was necessary to extrapolate the
velocities in the upper 400 m to the offshore side of the
stream, where the ADCPs failed to provide adequate sam-
pling. For this we chose a simple scheme that appeared to be
the most consistent with previous results, in which the
velocities above 400 m were forced to go through zero at
-135 km (the same place where the 400-, 700-, and 1000-m
velocities go through zero), and the velocity difference
between each level and the 400-m level was linearly de-
creased from its observed value at -50 km (the last good
data bin from the ADCPs) to zero at -135 km. Second, a
suitable scheme for vertical interpolation between measure-
ment levels and extrapolation to the bottom was needed.
Above 1000 m the vertical structure is well resolved by the
measurements, and there a cubic spline fit was used. Below
1000 m we used an exponential fit to the measured values
at 1000 m and 3500 m. This vertical interpolation was carried
out for each cross-stream bin, and the resulting profiles were
gridded onto a cross section at 50-m vertical by 5-km
horizontal resolution.

The final stream-coordinate cross section is shown in
Figure 9a, superimposed on the average cross-slope topog-
raphy in the central array. Transports are calculated by
discrete integration over various portions of this cross sec-
tion. In the remainder of this paper these transports are
separated into "Gulf Stream" and "recirculating" trans-
ports, by integrating only those portions corresponding to
downstream (eastward) and reverse (westward) flows, re-
spectively. Transports for the Gulf Stream and for the
northern recirculation are listed in Table 1, along with a
breakdown of the Gulf Stream transport into various deep
ranges. The total Gulf Stream transport is estimated to be
112.8 ± 8 Sv, while the transport in the resolved portion of
the northern recirculation is -16.6 ± 3.2 Sv. Transport
uncertainties include both random errors and probable bias
errors due to interpolation/extrapolation methods. The most
significant source of bias error is due to vertical interpo-
lation in the deep Gulf Stream over the large data gap between 1000
m and 3500 m. To investigate the possible size of this error,
we show in Figure 10 a vertical profile of Gulf Stream
transport-per-unit depth derived from our method, com-
pared with a similar profile obtained independently by Hall
and Fofonoff [1993] using hydrographic data taken across
the Gulf Stream in the central array area during March 1988
(several months before the array was first deployed). The
shape of the two curves is very similar in the upper 1000 m,
but below that they diverge, with the exponential fit appear-
ing to underestimate the deep transport by approximately 7.5
Sv. It should be borne in mind that Hall and Fofonoff's
results are based on assumptions about deep reference
velocities, and, in fact, the current meter derived estimates
at all levels (except 50 m) lie below their curve. Retaining
the shape of their curve but adjusting it lower by 1.8 Sv/km to
close approximately on our observed values at 1000 m and
3500 m (Figure 10, inset) reduces the transport difference
below 1000 m to 1.5 Sv. This is equivalent to a mean
reduction in Hall and Fofonoff's deep reference velocities
across the 150-km width of the Gulf Stream of only 1.2
cm/s. Thus there is a fair measure of agreement between the
two methods, taking into account the referencing uncer-
J

Figure 8. (top) Integral timescales corresponding to (bot-

(cm in the
to zero at -135 km (the same place where the 400-, 700-, and 1000-m
velocities go through zero), and the velocity difference
between each level and the 400-m level was linearly de-
creased from its observed value at -50 km (the last good
data bin from the ADCPs) to zero at -135 km. Second, a
suitable scheme for vertical interpolation between measure-
ment levels and extrapolation to the bottom was needed.
Above 1000 m the vertical structure is well resolved by the
measurements, and there a cubic spline fit was used. Below
1000 m we used an exponential fit to the measured values
at 1000 m and 3500 m. This vertical interpolation was carried
out for each cross-stream bin, and the resulting profiles were
gridded onto a cross section at 50-m vertical by 5-km
horizontal resolution.

The final stream-coordinate cross section is shown in
Figure 9a, superimposed on the average cross-slope topog-
raphy in the central array. Transports are calculated by
discrete integration over various portions of this cross sec-
tion. In the remainder of this paper these transports are
separated into "Gulf Stream" and "recirculating" trans-
ports, by integrating only those portions corresponding to
downstream (eastward) and reverse (westward) flows, re-
spectively. Transports for the Gulf Stream and for the
northern recirculation are listed in Table 1, along with a
breakdown of the Gulf Stream transport into various deep
ranges. The total Gulf Stream transport is estimated to be
112.8 ± 8 Sv, while the transport in the resolved portion of
the northern recirculation is -16.6 ± 3.2 Sv. Transport
uncertainties include both random errors and probable bias
errors due to interpolation/extrapolation methods. The most
significant source of bias error is due to vertical interpo-
lation in the deep Gulf Stream over the large data gap between 1000
m and 3500 m. To investigate the possible size of this error,
Figure 9. (a) Cross section of the stream-averaged velocity structure at 68°W after interpolation/extrapolation (see text). The mean profile of the 12° isotherm across the Gulf Stream is shown for reference. (b) Stream-averaged structure at 73°W [from Halkin et al., 1985].

The remainder of the 8 Sv error is due to random error (3.5 Sv), which is quite small owing to the large number of observations. Conversely, for the northern recirculation, random errors account for most of the uncertainty, since due to its barotropic nature it is relatively insensitive to the vertical interpolation method used.

It is interesting to compare Figure 9a with the stream-averaged structure deduced at 73°W by Halkin and Rossby [1985]. Their results are given for the full water column by Halkin et al. [1985] and are reproduced here in Figure 9b on the same plotting scales as Figure 9a. Overall, the two pictures resemble each other very closely. The surface width of the Gulf Stream is the same at both locations (160 km, between the 10-cm/s isotherms), and the shape of the contours in the upper 1000 m are nearly identical. However, the deep velocity structure is broader at 68°W; for example, at 2000 m the width of the 5-cm/s isotherm is 120 km versus approximately 85 km at 73°W. Also, the core velocities at 68°W are 10–30 cm/s larger in the upper 1000 m, e.g., nearly 2 m/s at the surface versus 1.7 m/s at 73°W.

These differences in velocity structure lead to significant differences in transport. Our Gulf Stream transport estimate of 112.8 Sv at 68°W is 19 Sv larger than the 93.7 Sv computed for the 73°W Pegasus line [Leaman et al., 1989]. The
Table 1. Summary of Volume Transports at 73°W and 68°W and Their Differences

<table>
<thead>
<tr>
<th>Transport, Sv</th>
<th>73°W</th>
<th>68°W</th>
<th>Difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total</td>
<td>93.7</td>
<td>112.8</td>
<td>19.1</td>
</tr>
<tr>
<td>0-1000 m</td>
<td>79.7</td>
<td>99.0</td>
<td>11.2</td>
</tr>
<tr>
<td>0-2000 m</td>
<td>87.8</td>
<td>99.0</td>
<td>11.2</td>
</tr>
<tr>
<td>2000 m-bottom</td>
<td>5.9</td>
<td>13.8</td>
<td>7.9</td>
</tr>
<tr>
<td>Baroclinic</td>
<td>47.4</td>
<td>51.0</td>
<td>3.6</td>
</tr>
<tr>
<td>Barotropic</td>
<td>46.3</td>
<td>61.8</td>
<td>15.5</td>
</tr>
<tr>
<td>Northern recirculation</td>
<td>...</td>
<td>16.6</td>
<td>...</td>
</tr>
</tbody>
</table>

The values at 68°W are taken from this analysis, while those at 73°W are taken from various analyses based on the 73°W Pegasus data [Halkin and Rossby, 1985; Leaman et al., 1989; Hogg, 1992]. The baroclinic and barotropic transports are as defined by Hogg [1992]; see text.

The choice of 1000 m allows direct comparisons with similar results obtained by Hogg [1992] at several downstream locations, including 68°W, which are discussed further in section 5.

Vorticity and Geopotential

The detailed velocity profiles across the Gulf Stream may be differentiated and integrated, respectively, to yield profiles of relative vorticity and geopotential anomaly across the Gulf Stream at various depths. The results are displayed in Figure 11 for the surface and the 400-m, 1000-m, and 3500-m levels. In the stream-averaged representation the surface relative vorticity reaches a cyclonic maximum of $0.60 \times 10^{-4}$ s$^{-1}$ (approximately $0.7 f$, where $f$ is the Coriolis parameter at this latitude $= 0.89 \times 10^{-4}$ s$^{-1}$) at about 20 km from the surface current axis and an anticyclonic maximum of $-0.43 \times 10^{-4}$ s$^{-1}$ (approximately $-0.5 f$), also about 20 km from the axis. This asymmetry in the shear magnitudes on the two sides of the current persists to 1000-m depth but with the ratio of maximum cyclonic to anticyclonic shear gradually weakening from about 1.4 at the surface to 1.1 at 1000 m, where the velocity profile becomes almost Gaussian in shape.

Assuming geostrophic balance, the surface geopotential height difference across the Gulf Stream at this location is 1.25 dynamic meters (dyn m) (equivalent to 1.27-m height difference). We emphasize that this is not based on any assumptions about a reference level or barotropic velocity.
the array must be associated with thermohaline transport in the Atlantic Deep Western Boundary Current (DWBC) that ultimately continues southward past Cape Hatteras. We will return to this issue in section 5, where an attempt is made to estimate the total westward transport north of the Gulf Stream and the portion of that flow that recirculates into the Gulf Stream.

4. Comparisons With Eulerian-Averaged Data

One of the obvious consequences of meandering is its broadening effect on the observed mean velocity structure across the Gulf Stream in the Eulerian frame relative to its synoptic structure. In a large-amplitude meandering region such as the present one this effect should be substantial, and in Figure 12 we show a comparison of these two representations for each of the standard observation levels. Figure 12a shows the mean downstream flow in both frames, and Figure 12b shows the corresponding velocity variances. To properly align these fields, the Eulerian quantities are plotted in terms of their meridional distance relative to 7°50′N, which was the mean location of the Gulf Stream’s axis along 68°W over the 2-year observation period. A summary of the Eulerian and stream-coordinate statistics is given in Table 2.

The Eulerian distributions are reasonably smooth except for a few outliers (e.g., site M13, which shows an anomalously low mean velocity, and site H5, which shows an anomalously low variance), which can be explained by the fact that only 1 year of data was available at these two sites. All other data points represent 26-month averages. Between 400 m and 1000 m the maximum mean downstream speeds observed in the Eulerian frame are smaller than those in the stream-coordinate frame by about a factor of 2, and the maximum variances are larger by about a factor of 4 (note that these are plotted as root-mean-square velocities in Figure 12b for convenience). At 400 m the maximum standard deviation in the stream-coordinate frame is about 0.34 m/s, equivalent to a kinetic energy per unit mass (EKE) of $570 \text{ cm}^2/\text{s}^2$ ($1 \text{ cm}^2/\text{s}^2 = 10^{-4} \text{ J/kg}$). At 50 m depth (not shown) the maximum standard deviation inferred from the ADCP data was 0.45 m/s or an EKE of approximately $1000 \text{ cm}^2/\text{s}^2$. These values are roughly twice as large as those found at the same depths by Halkin and Rossby [1985] at 73°W, indicating an increase in the structural variability of the Gulf Stream between the two locations, probably related to increased meander amplitudes.

By integrating the 400-m variances over the 250-km cross-stream width of the distributions, we find an average EKE of $214 \text{ cm}^2/\text{s}^2$ in the stream-coordinate frame versus $1370 \text{ cm}^2/\text{s}^2$ in the Eulerian frame (Table 2), meaning that only about 15% of the fluctuating energy observed across the array at this level can be accounted for by variability in the Gulf Stream’s structure. The remaining variance is due to lateral meandering of Gulf Stream plus variability due to Gulf Stream rings or other eddy features. At greater depths the ratio of variance in the stream-coordinate and Eulerian frames becomes progressively larger (22% at 700 m, 35% at 1000 m), and by 3500 m there is not a great difference in the total EKE averaged in either frame. It is interesting to note that the variance in stream coordinates at 3500 m is actually higher in certain places (notably on the northern side of the stream), suggesting that the deep current fluctuations there are largely uncoupled from the meandering of the Gulf Stream.

Figure 11. Cross-stream profiles of (a) geopotential anomaly, (b) downstream velocity, and (c) relative vorticity at the surface, 400 m, 1000 m, and 3500 m, derived from the stream-coordinate analysis.
In an attempt to hindcast the observed Eulerian distributions a simple calculation was carried out in which the synoptic Gulf Stream structure derived from the stream-averaging procedure was convolved with the observed distribution of the Gulf Stream axis location ($Z_{12}$ at 400 m) along 68°W, determined from the IES array (see Watts et al., submitted manuscript, 1994). Variability in the direction of the path was modeled by allowing the path direction to vary randomly between specified limits, with the mean direction being eastward. (For small directional changes this is unimportant, but for large directional changes the predicted zonal mean velocity is smaller and the variance is larger). For a realistic calculation it would be necessary to consider the joint distribution of the path location and direction; however, for the present purposes we have assumed the path direction to be statistically independent of path location and to occupy a uniform distribution between ±45°, which is a reasonable approximation for the array region. The resulting distributions of mean zonal velocity and variance across the array are shown in Figure 12 by dashed lines. Included in the variance distributions are the appropriately weighted contributions from the variance of the synoptic structure itself.

The predicted distributions of zonal mean velocity are in generally good agreement with the observations. The peak zonal mean velocities are particularly well reproduced, although the predicted distributions appear somewhat

Table 2. Comparison of Eulerian Versus Stream-Averaged Statistics

<table>
<thead>
<tr>
<th>Depth, m</th>
<th>$U_{\text{max, m}}$, m/s</th>
<th>$\text{EKE}_{\text{max}}$, cm$^2$/s$^2$</th>
<th>$\overline{\text{EKE}}$, cm$^2$/s$^2$</th>
<th>$T_d$, Sv/km</th>
<th>$U_{\text{max, m}}$, m/s</th>
<th>$\text{EKE}_{\text{max}}$, cm$^2$/s$^2$</th>
<th>$\overline{\text{EKE}}$, cm$^2$/s$^2$</th>
<th>$T_d$, Sv/km</th>
</tr>
</thead>
<tbody>
<tr>
<td>400</td>
<td>0.58 (0.58)</td>
<td>2160 (1735)</td>
<td>1370 (1066)</td>
<td>70.8 (75.6)</td>
<td>1.22</td>
<td>570</td>
<td>214</td>
<td>93.3</td>
</tr>
<tr>
<td>7000</td>
<td>0.31 (0.33)</td>
<td>769 (585)</td>
<td>471 (370)</td>
<td>39.9 (42.9)</td>
<td>0.67</td>
<td>205</td>
<td>105</td>
<td>54.5</td>
</tr>
<tr>
<td>1000</td>
<td>0.17 (0.15)</td>
<td>226 (149)</td>
<td>145 (117)</td>
<td>20.3 (21.4)</td>
<td>0.28</td>
<td>70</td>
<td>53</td>
<td>28.7</td>
</tr>
<tr>
<td>3500</td>
<td>0.02 (0.02)</td>
<td>128 (86)</td>
<td>81 (78)</td>
<td>— (92.8)</td>
<td>0.04</td>
<td>111</td>
<td>65</td>
<td>3.8</td>
</tr>
</tbody>
</table>

Quantities listed for each depth are $U_{\text{max, m}}$, maximum downstream velocity; $\text{EKE}_{\text{max}}$, maximum eddy kinetic energy; $\overline{\text{EKE}}$, average EKE over the cross-stream profile; and $T_d$, transport-per-unit depth. The numbers in parenthesis are the modeled Eulerian quantities (corresponding to the dashed curves in Figure 12).
broader than is observed. This is probably due to the
influence of Gulf Stream rings, whose westward flowing
segments adjacent to the Gulf Stream would tend to decrease
the observed zonal mean flow there relative to the predicted
distributions. (Recall that rings and other large eddy features
are specifically eliminated from the stream-averaging pro-
cedure, and their effects can therefore not be reproduced in
this calculation.) The predicted variance distributions also
follow the general structure of the observations but tend to
underestimate the observed variances. For the upper levels
this is probably due to the same reasons mentioned above,
while at deeper levels one can expect topographic Rossby
waves and other quasi-barotropic fluctuations to account for
a significant part of the residual variance.

It is clear from Figure 12a that the mean transport of the
Gulf Stream in the Eulerian frame is substantially smaller
than its mean synoptic transport. At each of the upper levels
the transport-per-unit depth estimated for the observed and
modeled zonal mean flow distributions (after extrapolation
of the observed distributions to zero on the offshore side) is
20–30% smaller than the corresponding synoptic values
(Table 2). At the 3500-m level we were unable to make a
reasonable estimate from the observed zonal mean veloc-
ities, but the modeled transport-per-unit depth there is also
about 25% smaller than the synoptic value. On the basis of
the modeled distributions we estimate the mean transport of
the Gulf Stream in the Eulerian frame to be approximately 88
Sv or about 25 Sv smaller than the mean synoptic transport
of 113 Sv. This should be considered as an upper bound on
the Eulerian transport, since the average mean zonal trans-
ports at the upper levels are about 5% smaller than the
modeled transports. The reduction in transport in the Eule-
rian frame is due directly to the adjoining recirculations; in
the absence of these recirculations the downstream transport
in the two frames should be identical, albeit with very
different distributions. Thus one can infer a similar reduction
in the apparent strength of the recirculations in the Eulerian
frame, though our data are of insufficient meridional extent
to actually prove this.

5. Discussion

Downstream Transport Increase of the Gulf Stream

The mean transport of the Gulf Stream near 68°W derived
from this analysis is 113 ± 8 Sv, representing a 19-Sv increase from Leaman et al.’s [1989] estimate of 94 Sv near
73°W. This corresponds to an average rate of transport
increase of approximately 4.2 Sv per 100 km downstream
distance over this region. This estimate is about a factor of 2
smaller than the average rate of transport increase of 8 Sv
per 100 km found by Knauss [1969] between Cape Hatteras
and 65°W. Thus the region between 73° and 68°W would
appear to be a region of below average transport growth.
Knauss [1969] derived his estimate from a small subset of
hydrographic sections across the Gulf Stream that were
accompanied by direct reference velocities. Given that these
individual sections may be unrepresentative of mean condi-
tions, it is useful to reconsider the Gulf Stream’s transport
increase using only estimates which include a significant
amount of averaging. In addition to the above values at 73°W
and 68°W, Hogg [1992] has produced average transport
estimates of 147 Sv at 60°W and 149 Sv at 54°W, using
stream-averaging techniques (Figure 13). Using his value at
60°W, the average rate of transport increase between 68°W
and 60°W is 4.7 Sv/100 km, which is very similar to the 4.2
Sv/100 km rate of increase between 73°W and 68°W.
Therefore on this coarse scale there is no concrete evidence to suggest
that the transport increase between 73°W and 60°W is
anything other than linear, with an average value of approx-
imately 4.5 Sv/100 km. However, the rate of transport
increase between Cape Hatteras and 73°W is considerably
larger, of order 15 Sv/100 km. In Figure 13 there is also some
suggestion of a nonuniform transport increase between 73°W
and 60°W, with a relatively larger rate of increase beginning
near 68°W, just upstream of the New England seamounts.
This inference is based entirely on the two “synoptic”
transport estimates near 64°W and 69°W [from Knauss,
1969], and as noted above, it is not clear that these estimates
should be given the same weight as the long-term estimates
from moorings or Pegasus. However, this picture also tends
to be supported by inflow observations, as discussed further
below.

Hogg’s [1992] estimate of the Gulf Stream transport near
68°W was 95.5 Sv, considerably lower than our 113 Sv. The
difference appears to be mainly due to much lower average
velocities near 1000 m found in his analysis (Figure 10).
He therefore concluded that the Gulf Stream transport was
nearly constant between 73°W and 68°W. (Note that Hogg’s
results at 68°W were based on data from only one mooring,
the GUSTO mooring, while at his other locations, two or
more moorings were available. The 1000-m record on the
GUSTO mooring also had a 56-day gap that may have
contributed to the low value at that level [see Hall and

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**Figure 13. Summary of observations of the Gulf Stream transport increase downstream of Cape Hattera to 55°W.** Solid circles are from Knauss [1969]; open circles are from Pegasus data at 73°W [Leaman et al., 1989] and from moorings at 68°W (this analysis) and at 60°W and 55°W [Hogg, 1992]. The breakdown of the total transport into baroclinic and barotropic components, relative to 1000 m, is also shown for the Pegasus and mooring locations (from Hogg [1992] at 73°W, 60°W and 55°W and from this analysis at 68°W).
Richardson's [1985] estimate of 93 Sv near 55°W. Thus there
flows from the RISE and SEEP arrays were first projected
results to estimate the total westward transport north of the
to merge these observations with the SYNOP central array
isobath (Figures 14 and 15). In Figure 14 an attempt is made
respectively, have provided a fairly comprehensive descrip-
tion of the top-to-bottom mean flow shoreward of the 3000-m
isobath [Hogg, 1983] and "SEEP" [Aikman et al., 1988] arrays,
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of the top-to-bottom mean flow shoreward of the 3000-m
isobath (Figures 14 and 15). In Figure 14 an attempt is made
to estimate the total westward transport of the Gulf Stream in this region. To accomplish this, the mean
flows from the RISE and SEEP arrays were first projected
onto a common section along 70°W, retaining the depth of
each measurement and placing it over the isobath where it
was collected. These data were then combined with the
SYNOP central array results from Figure 9 into a single
cross section by joining the two sections at the 3500-m
isobath. The assumption in merging these data sets together
is that the flow over the upper continental rise inside the
3000-m isobath is essentially a topography-following flow
that is continuous between 68°W and 71°W, which is a
reasonable assumption. Therefore the picture in Figure 14 is
taken to be representative of the net westward flow north of
the Gulf Stream near 68°W.

Our results indicate a westward transport of at least 16 Sv
of water to the north of the Gulf Stream at 68°W, which
is similar in magnitude and vertical structure to the transport
increase of the Gulf Stream between 73°W and 68°W. Since
not all of this westward flow is resolved by our measure-
ments, questions arise as to what the total westward trans-
port north of the Gulf Stream is in this region, and what
fraction of this water recirculates into the Gulf Stream versus continuing southward past Cape Hatteras as part of the
DWBC.

Long-term current meter observations over the upper
continental rise and slope south of New England show a
persistent westward mean flow with magnitudes of 3–10 cm/s
[Hogg, 1983; Watts, 1991]. Two large moored arrays de-
ployed in the region between 69°W and 71°W, the “RISE”
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The westward flow is characterized by maximum surface
velocities of 8–10 cm/s, with a mild surface intensification in
the upper 1000 m over the 3000-m isobath [Hogg, 1983]
and also adjacent to the Gulf Stream. Below 1000 m the veloc-
ities are fairly uniform, both horizontally and vertically, with
magnitudes of 3–5 cm/s. Surprisingly, there is little direct
indication of the DWBC, apart from a general increase in
near-bottom velocities offshore of the 2000-m isobath and a
weak bottom maximum near the 3500- to 4000-m isobath.
Drawing contours in the simplest possible way and integrat-
ing the flow across the section from the shelf to the edge of
the Gulf Stream, one obtains a total westward transport of
approximately 40 Sv (Figure 14). This value is considerably
larger than previous estimates of the westward transport
north of the Gulf Stream in this region (see Csanady and
Hamilton [1988] for a review); however, we do not believe
that it can be in error by more than about 10 Sv. The largest
source of uncertainty in this estimate is the data gap in the
upper 2000 m between the 3000- and 3500-m isobaths (see
Figure 14). If the flow was taken to be zero here, it would
reduce the net transport by approximately 8 Sv, although

Northern Recirculation and Inflow Structure

In terms of Eulerian transports, it is interesting to note
that our estimate of 88 Sv at 68°W is nearly the same as
Richardson’s [1985] estimate of 93 Sv near 55°W. Thus there
appears to be little change in the Eulerian transport of the
Gulf Stream over this distance, despite the fact that the
synoptic transport increases by O(40 Sv). This can be taken
as further evidence that the downstream transport increase
of the Gulf Stream is fed primarily by tight recirculation
cells, whose meridional scale is comparable to the width of
the Gulf Stream’s meandering envelope, rather than on a
larger (gyre) scale.

Hogg [1992] also partitioned the Gulf Stream
transport into relative baroclinic and barotropic compo-
nents, using a 1000-m reference level (see section 3), at
73°W, 68°W, 60°W, and 54°W. He found that the baroclinic
transport at all locations was nearly constant, at about 47 Sv,
while the barotropic transport increased from values of
46–48 Sv at 73°W and 68°W, to 100–102 Sv at 60°W and 54°W
(Figure 13). Our estimate of the baroclinic transport at 68°W
(using the same 1000-m reference) is similar, at 51 Sv, but the
barotropic transport is considerably larger, 61.5 Sv (Table
1). This suggests an upward revision of about 15 Sv from
Hogg’s [1992] estimate of the barotropic transport at 68°W.

Our estimate of the baroclinic transport is generally consis-
tent with Hogg’s conclusion that the baroclinic transport
changes little downstream of Cape Hatteras, and supports
the notion that the transport increase is primarily barotropic.

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Figure 14). If the flow was taken to be zero here, it would
reduce the net transport by approximately 8 Sv, although
this seems highly unlikely, since the magnitude and vertical structure of the westward flows bordering this region are very similar. The representativeness of the RISE array mean currents might also be questioned due to their relatively short length (8 months); however, Watts [1991] found that these mean flows were quite stable (typically to within ±1 cm/s), and Luyten [1977] showed that the mean flows near the 3000-m isobath from the RISE array were very consistent with multiyear records available there from site "D." According to Figure 14, over half of the total westward transport, or approximately 24 Sv, occurs shoreward of the 3500-m isobath. Assuming a 1-cm/s bias error over this entire region results in a transport error of ±6 Sv. Combining this with our transport estimate for the offshore region (16.6 ± 3.2 Sv), we therefore take 40 ± 10 Sv as a reasonable estimate of the total westward transport north of the Gulf Stream near 68°W. This value is comparable in magnitude to the westward recirculation north of the Gulf Stream near 55°W [Richardson, 1985; Hogg, 1983] and suggests that either a substantial portion of that flow continues westward past the New England Seamount chain (NESC) or that a separate recirculation cell exists between Cape Hatteras and the NESC. In either case a large inflow to the northern side of the Gulf Stream between Cape Hatteras and 68°W is implied. Assuming that the DWBC accounts for roughly 10 Sv of this flow [Hogg, 1983; Watts, 1991], this still leaves order 30 Sv that must join into the Gulf Stream between Cape Hatteras and 68°W.

The dynamical reasons for such a large northern recirculation in this region are unclear, although it is likely that a portion of this flow is associated with a cyclonic gyre in the Slope Water region driven by positive wind stress curl north of the Gulf Stream. On the basis of the annual mean wind stress curl, Leetmaa and Bunker [1978] estimated the strength of this gyre to be 10–15 Sv. On a long-term average one would expect this flow to reveal itself as a baroclinic westward current over the continental slope and rise south of New England, gradually intensifying into a southward current near Cape Hatteras before joining into the Gulf Stream. The westward baroclinic transport relative to the bottom calculated from Figure 14 is approximately 12 Sv, which is consistent with the expected magnitude of this wind driven flow. Therefore up to one third...
of the total westward transport near 68°W might be attributed to a wind-driven recirculation.

The transport of the DWBC south of New England is probably of order 10–15 Sv [Hogg, 1983; Hogg et al., 1986; Watts, 1991], although Richardson [1977] concluded that as much as 24 Sv may cross southward in the DWBC beneath the Gulf Stream at Cape Hatteras. Water mass and tracer observations [Pickart, 1992a] show that the DWBC contains convective source waters ranging in potential temperature from 2 to 5°C, including Norwegian-Greenland Overflow Water (2–3°C), Labrador Sea Water (3–4°C), and a warmer 4–5°C class originally referred to by Weiss et al. [1985] as modified Labrador Sea Water. South of New England, these waters occupy a bottom wedge approximately 1000 m thick extending down the continental margin from 800-m to 4000-m depths [Watts, 1991]. Approximating this distribution by the hatched area in Figure 14, a transport of 9 Sv is estimated for the DWBC, which is in line with the smaller of the above estimates. Thus another one fourth or possibly more of the total westward transport is associated with the DWBC.

The remainder of the transport, of order 10–20 Sv, must be accounted for by a different dynamical mechanism, and a likely candidate is an eddy-driven recirculation of the type described by Holland and Rhines [1980]. The large barotropic component of the westward flow, including the strong and relatively uniform velocity at middepth, is suggestive of this type of recirculation. The fact that a deep recirculation exists north of the Gulf Stream in this region was demonstrated by Pickart [1992b], who found a secondary DWBC tracer maximum offshore of the main DWBC core on several different sections between Cape Hatteras and the New England seamounts, which was embedded in the deep Gulf Stream flowing eastward. He concluded that this water acquires its tracer signature through lateral mixing with the DWBC along westward trajectories in the recirculation before joining the Gulf Stream.

A large westward recirculation north of the Gulf Stream is also supported by direct estimates of inflow to the northern edge of the Gulf Stream. Our results at 68°W indicate a nearly barotropic northern inflow of 3–4 cm/s into the Gulf Stream, with weaker and generally insignificant inflows on the southern side. Halkin and Rossby [1985] also found a large northern inflow to the Gulf Stream near 73°W, although their inflow profile showed a significant reduction in strength below 1000 m. Averaging over the upper 2000 m, they found that about twice as much water enters the Gulf Stream from the north at 73°W as from the south, with an average 0- to 2000-m northern inflow of 4.6 cm/s, similar to our northern inflow estimates. From their data they inferred a local rate of transport increase near 73°W of approximately 15 Sv per 100 km downstream distance, with 10.4 ± 4.4 Sv per 100 km entering from north. Using the same method at 68°W, we obtain an even larger estimate of 12 ± 4 Sv per 100 km entering from the north. Ironically, these inflow estimates actually appear to be too large to be consistent with the observed rate of transport increase of the Gulf Stream. For example, if these values are assumed to be representative of the average inflow rate over the 450-km downstream distance separating the two sections, one arrives at a transport increase of nearly 60 Sv, which is three times greater than the observed transport increase between 73°W and 68°W. Therefore it must be concluded that these local inflow rates are either in error or that they are unrepresentative of the average inflow rate over this region.

To check the consistency of these estimates, it is useful to consider a simple model of how westward flow north of the Gulf Stream might be expected to join into the Gulf Stream. Since the westward flow is largely barotropic, to a first approximation the flow streamlines should follow bottom contours (more properly, *f* contours) southwestward to the location where those isobaths intersect the Gulf Stream. There the flow must either join the Gulf Stream or descend beneath it to follow deeper isobaths [Hogg and Stommel, 1985]. A calculation was therefore carried out in which the westward transport from Figure 14, excluding the estimated portion due to the DWBC, was carried southwestward along isobaths to the point where it intersects the long-term mean path of the Gulf Stream given by Cornillon [1992] (see Figure 15). There the flow is assumed to join the Gulf Stream. The result of this model is a prediction of the local rate of northern inflow to the Gulf Stream between Cape Hatteras and 68°W and the corresponding transport increase of the Gulf Stream due to northern recirculation alone (Figure 16).

Several features of this calculation are worthy of note. First, the predicted northern inflow rate is not uniform but shows a maximum near 69°W, with weaker maxima near 74°W and 71°W. These are general areas where the Gulf Stream is crossing most steeply across isobaths in its descent across the continental rise (Figure 15). The weak inflow predicted between 71°W and 69°W is due to the fact that the Gulf Stream flows over an essentially constant water depth of 4000 m there, parallel to the isobaths. Second, the large inflow rates near 74°W (8 Sv/100 km) and 69°W (13 Sv/100 km) are remarkably consistent with the observed northern inflow rates in those two regions, 10 Sv/100 km and 12 Sv/100 km, respectively. (Note that the northern end of the Halkin and Rossby [1985] Pegasus line and the SYNOP central array are actually located closer to 74°W and 69°W, rather than the nominal 73°W and 68°W locations; see Figure 15). Third, the difference between the two curves in Figure 16, which is the inferred (residual) transport addition from the
offshore side of the Gulf Stream, suggests a strong southern inflow from Cape Hatteras to approximately 73°W, after which this southern inflow becomes much weaker. For example, between Cape Hatteras and 73°W, Figure 16 would imply that roughly 10 Sv of the total transport increase is provided from the north and 25 Sv from the south, leading to a total transport increase of 35 Sv. Conversely, between 73°W and 68°W the transport increase from the north (22 Sv) essentially accounts, within errors, for the total transport increase over this domain (19 Sv), implying negligible southern inflow in this region. The fact that northern inflow already appears to dominate at the Halkin and Rossby section is consistent with a transition between these inflow regimes occurring near 73°W.

The simplicity of this model prevents it from being used for more than a qualitative description of the northern inflow structure. In particular, cross-isobath motion in the westward recirculation due to eddy forcing or vorticity input from the wind is neglected. Uncertainties of the Gulf Stream transport and the magnitude of the westward transport north of the Gulf Stream (including that portion carried in the DWBC) are also considerable, and changes in these numbers will affect the above results. However, the model does appear to be largely successful in rationalizing apparent discrepancies in the measured inflow rates and the observed transport increase of the Gulf Stream and, so far as we are aware, is consistent with all available observations in the region.

We conclude therefore that the large northern inflow rates found near 73°W and 68°W are probably real and that they indicate that these are regions of locally intense inflow. The most likely explanation for the large northern inflow near 68°W is that it is a continuation of the strong westward flow offshore of the 4000-m isobath found near 63°W [Hogg, 1983], which is forced to join into the Gulf Stream near 68°W, where these isobaths converge on the Gulf Stream’s mean path. Water property distributions also show a tongue of water with high DWBC tracer content extending southward beneath the Gulf Stream near 68°-70°W and then eastward [Hogg et al., 1986], suggesting a localized northern inflow to the Gulf Stream here. These observations therefore suggest that the western limb of the abyssal Northern Recirculation Gyre centered near 55°W extends beyond the NESC to at least 68°W.

Part of the deep inflow to the Gulf Stream near 68°W may also be associated with a smaller-scale cyclonic recirculation, indicated in bottom pressure records from the SYNOP central array [Watts et al., submitted manuscript, 1994], that is perhaps embedded in the western reaches of the northern abyssal gyre. It should be noted that the closure of the Northern Recirculation Gyre to the east remains poorly documented, and while we refer to the general westward flow north of the Gulf Stream simply as “recirculation” (excluding that part associated with the DWBC), this does not mean that all of this flow recirculates on the same scale or that a complete closure necessarily occurs at all depths in the vicinity of the Grand Banks. Indeed, the upper portion of this flow, historically referred to as the Slope Water Current, has water properties substantially different from those in the Gulf Stream [e.g., Hall and Fofonoff, 1993], indicating replenishment either through a direct advective connection with the subpolar gyre or by vigorous mixing with subpolar waters near the Grand Banks.

6. Summary

Current meter observations from the SYNOP central array, a 13 mooring array deployed in the Gulf Stream near 68°W, were analyzed to determine the average synoptic structure of the Gulf Stream at this location. Mapping of the current meter observations into daily stream coordinates was accomplished using objective maps of thermocline depth produced from a simultaneous array of inverted echo sounders. This method was found to be superior to more conventional methods [Hall and Bryden, 1985], allowing the stream-coordinate mapping to be extended to the weakly sheared flanks of the Gulf Stream and allowing rings and other unrepresentative eddy features to be readily excluded from the analysis.

The structure of the Gulf Stream derived from this analysis exhibits a number of characteristic features commonly observed in geostrophic sections across the Gulf Stream. These include (1) an offshore shift of the velocity core with increasing depth such that at 1000 m the velocity maximum is displaced 35 km seaward of the surface velocity maximum; (2) a subsurface maximum on the offshore side of the current; and (3) an asymmetric shear structure in the upper 1000 m, with larger horizontal shear on the cyclonic (shoreward) side of the current. The stream-coordinate analysis also reveals a deep extension of the Gulf Stream, with core velocities of 3–4 cm/s at 3500 m depth. In general, the synoptic structure of the Gulf Stream near 68°W was found to be remarkably similar to that deduced by Halkin and Rossby [1985] near 73°W.

The synoptic transport of the Gulf Stream near 68°W determined from our analysis is 113 ± 8 Sv. This value is 19 Sv larger than the synoptic transport of 94 Sv found at 73°W [Leaman et al., 1989], indicating an average rate of transport increase between these two locations of 4.2 Sv per 100-km downstream distance. The transport increase is split nearly evenly above and below 2000 m and is consistent with previous conclusions [Hogg, 1983, 1992] that the transport increase of the Gulf Stream downstream of Cape Hatteras is primarily barotropic. Inflow to the Gulf Stream at 68°W is also found to be nearly barotropic and is dominated by inflow on the northern side of the Gulf Stream. This result is similar to Halkin and Rossby’s [1985] findings at 73°W, where northern inflow to the Gulf Stream was found to be about twice as strong as southern inflow. Taken together, these studies suggest that a considerable portion of the transport increase of the Gulf Stream between Cape Hatteras and the New England seamounts is derived from northern inflow, rather than being supplied primarily by southern inflow, following the original concept of Worthington [1976].

To support this conclusion, an attempt was made to determine the total westward transport north of the Gulf Stream near 68°W by combining results from the SYNOP central array with available long-term mean current observations in the Slope Water region from the RISE and SEEP arrays [Luyten, 1976; Aikman et al., 1988]. The northern part of the SYNOP central array extended shoreward of the Gulf Stream for much of the experiment and showed significant westward mean flow bordering the Gulf Stream, with an estimated transport of 16.6 ± 3.2 Sv between the Gulf Stream and the 3500-m isobath. A similarly large value (23.8 ± 6 Sv) was estimated for the transport inshore of the
3500-m isobath, leading to a total westward transport north of the Gulf Stream near 68°W of 40 ± 10 Sv. Approximately 10 Sv of this flow is assumed to cross under the Gulf Stream near Cape Hatteras in the DWBC. The remaining transport, of order 30 Sv, is assumed to join the Gulf Stream between Cape Hatteras and 68°W. A simple diagnostic calculation in which this flow is carried westward along isobaths, to the point where it intersects the mean path of the Gulf Stream, is shown to produce a northern inflow distribution that is consistent with the observed inflows at 68°W and 73°W.

Concerning the transport increase of the Gulf Stream downstream of Cape Hatteras, the picture that seems to be emerging from the available data is that of a rapid increase in Gulf Stream transport in the first few hundred kilometers downstream of Cape Hatteras, followed by a leveling off of the transport increase, with a further sharp increase in transport beginning at around 68°W (Figure 13). A similar description is suggested by Geosat altimeter results [Kelly, 1991], which indicate a sharp increase in the surface height difference across the Gulf Stream upstream of 72°W and again near 68°W, with an intervening plateau (or possible minimum) near 70°W. The results of the present study suggest that the total addition to the Gulf Stream transport over this domain is split nearly evenly between northern and southern inflow, but that northern inflow is dominant between 73°W and 68°W, while most of the southern inflow is concentrated very near Cape Hatteras.

Appendix: Comparison of Stream-Averaging Methods

Herein we compare the stream-coordinate structure obtained using the IES method, described in the paper, with a more conventional approach based on the Hall and Bryden [1985] method (hereinafter referred to as the HB method). In the HB method the temperature at a selected depth z₀, corresponding to a thermocline measurement level, is used as an indicator of the instantaneous cross-stream location of each mooring within the current. Downstream velocities at all measurement levels are then grouped into classes based on the temperature T₀ at the reference depth z₀ (typically in intervals of 0.5 to 1.0°C) and averaged to produce a mean velocity distribution at each level, u(To). The cross-stream spacing between temperature classes or bins is then determined by inverting the thermal wind relation, to yield

\[
\frac{\partial y}{\partial z} = \frac{10}{\partial D} (\Delta u)^{-1}
\]

where in our application, Δu is the downstream velocity difference between the 400- and 1000-m levels and D is the dynamic height anomaly between 400 and 1000 m,

\[
D = \int_{1000}^{400} \delta(T,S,p) \, dp.
\]

D therefore replaces T₀ as the indicator of relative cross-stream position. The uncertainties due to the estimation of α and (∂u/∂z)z₀ in the conventional HB method are now replaced by having to estimate the 400- to 1000-m dynamic height anomaly from temperature measurements at three levels. The main advantages are that this is an integral operation, and also that D should be a more robust indicator of relative cross-stream location than the temperature at any particular level.

To estimate D, daily pseudotemperature profiles between 400 and 1000 m were created at the mooring sites by integrating upward and downward from bracketing measurement levels and forming a weighted average of the two estimates, namely,

\[
T(z) = \sum_{i=1}^{2} w_i \left[ T_i + \int_{z_i}^{z} \frac{\partial T}{\partial z}(T, z) \, dz \right]
\]

where \( w_i = 1 - |z - z_i|/\Delta z \), \( \Delta z = z_2 - z_1 \) and where \( \partial T/\partial z(T,z) \) is an empirically derived function based on approximately 200 conductivity-temperature-depth (CTD) profiles collected in the central array region (CTD data...
Figure A2. Scatterplots of downstream velocity at 400 m versus (a) cross-stream distance (IES method), (b) 400-m temperature, (c) 700-m temperature, and (d) 400- to 1000-m dynamic height anomaly. The curves in the top panels represent averages over 100 sequential data points, and in the bottom panels the standard deviations about those averages. The offshore edge of the mean distribution in Figure A2b is indicated by the arrowhead at right.
from a modified form of the Hall and Bryden [1985] method. The profiles were then integrated in 10-m vertical increments to obtain daily estimates of the 400- to 1000-m dynamic height anomaly at each current meter site. The most important difference between these indicators, however, is that the IES derived cross-stream position yields the lowest total velocity variance. Of the possible HB method indicators the dynamic height is superior, although it is only marginally better than the 700-m temperature. The 400-m temperature is the poorest due to the large scatter in the 18° range. It can also be seen that most of the increased variance associated with the HB method indicators occurs on the offshore side of the stream, while over the rest of the cross section the variances are similar to the IES method or even slightly smaller (see further discussion below). The most important difference between these indicators, however, is that the IES derived position shows a clear offshore edge of the Gulf Stream, while average downstream velocities at the offshore edge of the other distributions are still of order 0.2 m/s. For these distributions the mean velocity could possibly be brought smoothly through zero by fitting the data to a low-order polynomial, although this would introduce a certain amount of subjectivity to the analysis.

A direct comparison between the IES and modified HB method is shown in Figure A3, where the dynamic height has been mapped into relative cross-stream position according to (2). Variably spaced dynamic height classes, with increments from 0.01 to 0.04 dyn m, were selected to yield a horizontal resolution similar to that for the IES method. Overall, the agreement between the two methods is remarkably good. Both methods show maximum velocities at 400 m of approximately 1.3 m/s and have nearly identical cross-stream structures spanning the core region and cyclonic sides of the stream. The main difference, as noted above, is on the offshore side, where the IES method shows a clear offshore limit of the downstream flow, while the HB method does not. The same general result holds at the other levels. The HB method cannot be extended over the same range as the IES method results, since the integration of (2) cannot proceed past the point where measurable vertical shear between 400 m and 1000 m is present.

Both methods show maximum standard deviations of approximately 30 cm/s but have different distributions. The IES method variances are larger in a narrow region centered on the cyclonic shear zone, while those for the HB method are larger (by nearly a factor of 2) over the entire offshore flank. It can therefore be concluded that dynamic height is a more robust indicator of current speed within the cyclonic shear zone, while distance from the axis is a better indicator on the anticyclonic side of the current. The result of these different variance distributions is to smear the corresponding mean structures in those locations; that is, the cyclonic zone...
is slightly broader in the IES method representation, while the anticyclonic flank is substantially broader in the HB method representation.

The reason the HB method representation fails to show the offshore zero crossing can be seen from Figure A1. On the offshore side of the Gulf Stream there is a relatively wide offshore zero crossing can be seen from Figure A1. On the offshore side of the Gulf Stream there is a relatively wide offshore distance, obscuring the nearly linear decrease though zero seen in the IES method representation. The same result was found using the temperature at 700 m to map the current structure (the best choice among the three available thermocline levels), confirming that this is not due to some unforeseen noise in estimating the 400- to 1000-m dynamic heights.

On the basis of the above considerations we believe the IES mapping method is more robust and produces the best overall representation of the Gulf Stream's structure, and we have therefore used it exclusively for transport estimates and other derived quantities reported in the paper.

Acknowledgments. We wish to thank the captain and crew of the R/V Oceanus and R/V Endeavor for their capable assistance at sea and to the mooring groups at the University of Miami, University of Rhode Island, and Woods Hole Oceanographic Institution for their skillful preparation and execution of the mooring operations. Assistance in data processing was provided by Rainer Zantopp (UM), Sura Haines (UNCG), and Karen Tracey (URI). The mooring motion corrections were performed by Meghan Cronin at URI. Helpful comments and discussion were obtained from Bob Pickart and Melinda Hall. Support for this work was provided under grants OCE-87-16530, OCE-8717144, and OCE-8717141 from the National Science Foundation and initial contracts N00014-89-J1139, N00014-88-K0465, and N00014-87-K0235 from the Office of Naval Research.

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