Gulf Stream Dynamics. Part I: Mean Flow Dynamics at 73°W

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ABSTRACT

During 1984, five current meter moorings measured velocity and temperature in the Gulf Stream anticyclonic flank at a location approximately 250 km downstream of Cape Hatteras. Here, these data are used to analyze the energy budgets of the Gulf Stream mean flow with a view towards examining Gulf Stream vertical structure and inertial character.

We find that Gulf Stream dynamics exhibits considerable vertical structure at our experiment site. At 380 m, the kinetic energy flux appears to be convergent, but the eddies augment mean kinetic energy. Order of magnitude estimates of processes involving vertical velocity appear to be too small to account for this mismatch; hence, we conclude that a conversion of mean kinetic to mean potential energy, via flow up a mean pressure gradient, must be occurring. Opposite tendencies are found at 880 m, leading to the conclusion that the Gulf Stream is flowing down a mean pressure gradient at this depth. Evidence supporting a situation similar to the latter is also found at 1880 m. We appeal to recent baroclinic general circulation theories to explain these observations in terms of northward shifts of the gyre structure with depth. Of course, our observations are undoubtedly influenced by lateral topography, bottom topography and eddies and the effects of these are poorly understood from a theoretical point of view.

We also present evidence that the Deep Western Boundary Current is restoring energy to the deep potential energy field by flow up a mean pressure gradient. The rates are considerably smaller than those in the Gulf Stream but structurally resemble our results at 380 m.

1. Introduction

It is generally accepted that western boundary currents are of fundamental importance to the dynamics of steady basin-scale circulations. It has been suggested that such currents are regions of enhanced loss of potential vorticity and energy and act to restore global balances between forcing and dissipation. Effects such as the role of Gulf Stream heat transport and its subsequent feedback on the structure of the general circulation have also been examined (Behringer et al. 1979; Huynh and Veronis 1981). Such studies emphasize the need for examining the time-mean Gulf Stream when engaging more global and basin scale questions. Nonetheless, although transient events and eddy dynamics in the Gulf Stream have been the subject of several previous observational programs (Webster 1961a,b, 1965; Schmitz and Niiler 1969; Brooks and Niiler 1977; Hood and Bane 1983; Lee and Atkinson 1983; Dewar and Bane 1985; Hall 1986b; Rossby 1987; Dewar and Bane 1989), Gulf Stream mean flow has received considerably less observational attention and much less is known about its dynamics.

The role of the Gulf Stream in theoretical general circulation calculations is highly dependent upon model dynamics which, depending on parameter settings, vary considerably. In the simplest barotropic models (e.g., Stommel 1948; Munk 1950), the Gulf Stream assumes a frictional role and the boundary currents are regions of energy loss. At the opposite extreme, in Fofonoff's (1954) unforced model, boundary currents are regions in which total energy is conserved and cycled between kinetic and potential energy. We are still at the stage, in our examination of the lowest order dynamics of the oceanic Gulf Stream, where the simplest classes of models, such as those described above, need to be compared with data. Distinguishing observationally between simple models, however, is not a straightforward task. The fundamental difference between frictional (i.e., Stommel and Munk) and conservative (i.e., Fofonoff) Gulf Stream models lies in the importance of nonlinearity. Such models are thus relatively distinct dynamically; however, all three yield circulations which are kinematically similar. Velocity measurements do not yield a clear picture of which model is most appropriate. On the other hand, a rel-

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atively sensitive test of individual models can be realized observationally by examining boundary current energetics. To the extent that nonlinear, inertial dynamics are appropriate in the boundary currents, the lowest order dominance of geostrophy is broken and flows in the direction of mean pressure gradients ensue. These are connected to potential energy release and with suitable measurements are detectable. Energetics analyses also benefit from the cancellation of the usually dominant Coriolis effect in the equations.

A second issue is that the best studied existing Gulf Stream models have relatively coarse vertical resolution (Schmitz and Holland 1982). Little is therefore known theoretically about the vertical structure of western boundary current dynamics. With a few notable exceptions, this is true also of our field knowledge of the Gulf Stream.

In this paper, data from the Gulf Stream Dynamics Experiment are used to examine the energetics of the Gulf Stream mean flow at a location approximately 250 km downstream of Cape Hatteras with a view towards addressing the above issues. The results of this analysis help to define the effects of eddies on the mean flow, quantify the inertial character of the Gulf Stream and point to some rather surprising results regarding the vertical structure of stratified western boundary currents. We further examine Deep Western Boundary Current dynamics and find a nontrivial exchange between mean potential and mean kinetic energy. Our results are viewed from the perspective of simple theoretical models.

a. Background

An important contribution to our empirical knowledge of Gulf Stream energetics was made by Fofonoff and Hall (1983), who combined a theoretical jet model with the Gulf Stream '60 data (Fuglister 1963) in order to compute mass, momentum and energy fluxes. The Gulf Stream '60 data were obtained from 53° to 68°W and consisted primarily of hydrographic observations. Fofonoff and Hall suggested that these data were consistent with a downstream rise in pressure. The Bernoulli function

$$B = \frac{u^2 + v^2}{2} + P$$

where u and v are velocity components and P is pressure, is conserved along streamlines in inviscid, inertial models. Reductions in kinetic energy are balanced by increases in pressure which can eventually be related to potential energy variations (pressure is hydrostatic). Thus, Fofonoff and Hall's arguments are consistent with an inertial restoration of potential energy in the Gulf Stream '60 area.

More recently, Dewar and Bane (1985) examined data from the Gulf Stream Deflection and Meander Energetics Experiment, which was located well upstream of the Gulf Stream '60 experiment (i.e., in the South Atlantic Bight south of Cape Hatteras). Based on direct measurements of velocity, they suggested that the Gulf Stream in the South Atlantic Bight was flowing down a mean pressure gradient and hence being accelerated by a conversion of potential energy to kinetic. Furthermore, their pointwise conversion estimate, when applied all along the South Atlantic Bight, yielded a net conversion comparable in magnitude but opposite in sign to Fofonoff and Hall's net conversion. Hence, they concluded that they were seeing evidence of a recycling of energy, with the South Atlantic Bight being a region of potential energy release and the Gulf Stream extension being a region of potential energy gain.

Most previous, long-term, direct current measurements in the deep water Gulf Stream system (i.e., downstream of Cape Hatteras) have been obtained along the continental rise. Luyten (1977) and Thompson (1977, 1978) summarize these. One of the reasonably well established observational features of the mean Gulf Stream is its downstream increase in transport (Watts 1983), although the dynamical reasons for this increase are not clear. Thompson (1978) suggested that eddies were transferring momentum between the Gulf Stream and adjacent countercurrents, and that this "eddy-driving" could possibly be responsible for the observed transport variations. Support for this statement came from his analysis of current meter data at 69° and 70°W, which indicated momentum flux gradients of the right sign to accelerate both the Gulf Stream and deep countercurrent flows.

More recently, Hogg (1983, 1985) reexamined the issue of eddy effects on the deep circulation of this region from the point of view of vorticity dynamics. He suggested that lateral relative vorticity and thickness fluxes are of comparable strength in the deep water and, given relatively large error bars, appear to have gradients of the proper sign to drive at least part of the observed multigyre deep circulation.

The present paper builds upon these past efforts in the following ways. First, the Gulf Stream Dynamics Experiment occurred at a location upstream of the Gulf Stream '60 experiment region and downstream of Cape Hatteras. Thus, the data are from a region which has not been previously studied with moored current meters. Second, we have long-term direct velocity measurements and can directly estimate several of the terms in the mean energy equation (as opposed to inferring them from hydrographic data). Third, we have observations at several depths and can therefore examine the vertical structure of the Gulf Stream system. Finally, our array permits us to compute both alongstream and cross-stream derivatives, which enables us to estimate more components of the mean energy equations (and hence assess the balances more thoroughly) than has been previously possible.

Accordingly, we find that the Gulf Stream at 73°W (our experiment site) is apparently in a transition zone between a region to the west of net acceleration and a region to the east of net deceleration. The transition manifests itself baroclinically, such that the near surface (380 m) Gulf Stream is recycling energy into potential energy although the deeper flow is still accelerating. This picture suggests a cross-stream mean velocity profile which seems consistent with other Gulf Stream observations.

We also find a net tendency for the eddies to accelerate the Deep Western Boundary Current. This is in accord with Thompson's (1978) results, but we measure a convergence in mean kinetic energy flux. A balance in the mean kinetic energy equation at this level apparently requires a conversion from kinetic to potential energy.

We interpret the results from our upper three levels to suggest two things about the mean Gulf Stream. First, we find that the Gulf Stream exhibits considerable inertial character in that it is actively recycling energy. Second, we argue that our observations are suggestive of a downstream shifting with depth of the inertial "center" of the Gulf Stream. We are unaware of any theoretical Gulf Stream models that display such structure.

We interpret our deeper level observations as pointing to the pressure gradient as the momentum source behind the downstream transport increase in the Gulf Stream. This picture differs slightly from that of Thompson and is based on some estimates which, due to data limitations, he was unable to make.

The Gulf Stream Dynamics experiment and data processing methods are discussed in the next section. The mean energy equation is introduced in section 3 and our results are presented. Our picture of the mean Gulf Stream is compared and contrasted with the results of models and other field programs. A discussion and conclusions section closes the paper.

2. The Gulf Stream Dynamics Experiment and data processing

The Gulf Stream Dynamics Experiment took place from mid-January 1984 to mid-January 1985. The experiment was located approximately 250 km downstream of Cape Hatteras along the mean Gulf Stream path (see Fig. 1). Five moorings, each supporting four current meters, formed one of the two major components of the field work. The total water depth in the study area ranged from 3500 to 4000 m, and the nominal depths of the three top current meter levels were 380 m, 880 m, and 1880 m. The fourth current meter level was designed nominally to be 500 m off the bottom. A schematic of the current meter array with actual current meter depths is given in Fig. 2.

The second component of the Gulf Stream Dynamics Experiment consisted of bottom mounted pressure

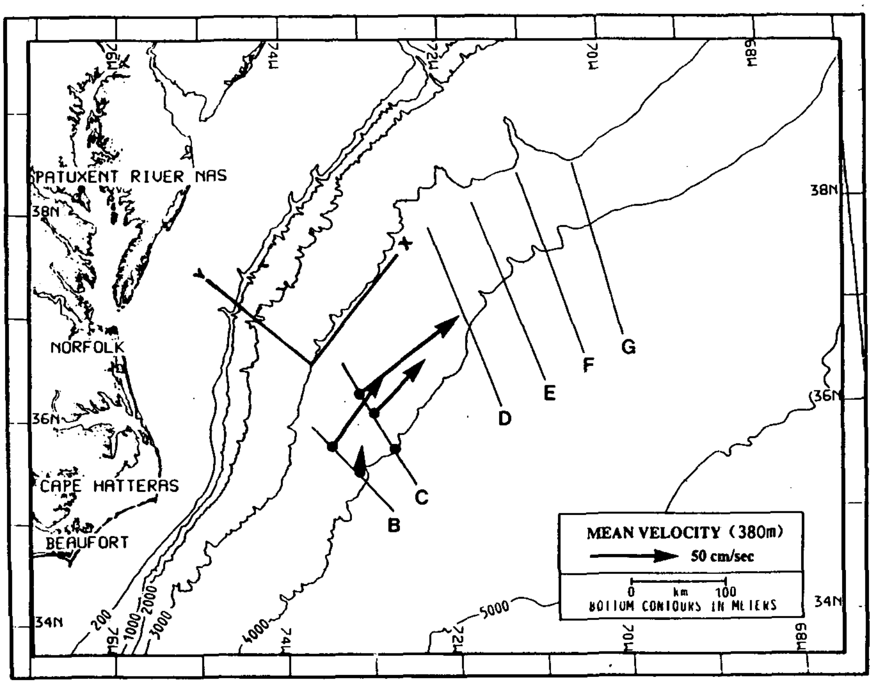

FIG. 1. The Gulf Stream Dynamics Experiment region. Identifiable land marks include Cape Hatteras and Chesapeake Bay. The five current meter moorings are indicated by the black dots and the arrows indicate the observed mean flows at level I, i.e., 380 m. The lines indicate the locations of the inverted echo sounders, and the coordinate system used in our analysis is indicated by the axes. The array is evidently in the Gulf Stream anticyclonic shear zone.

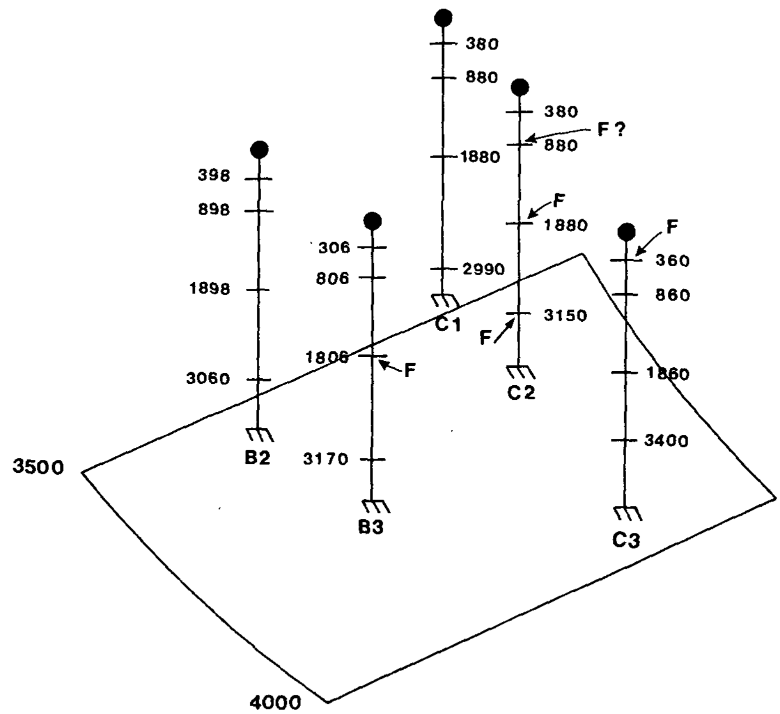

FIG. 2. A detailed view of the array. Each mooring supported four current meters at nominal depths of 380 m, 880 m, 1880 m and D-500 m, where D is the local fluid depth. The actual depths of the current meters are indicated and meters which failed are denoted by "F".

gauges and inverted echo sounders (IESs), which were placed along six lines, each of which runs roughly normal to the mean path of the Gulf Stream (see Fig. 1). The IES data will be reported separately. This study will focus on the current meter data only.

Note that two of the IES lines (the B and C lines), correspond roughly to the lines on which the current meters were mounted. Hence, the current meter moorings are referred to as the B2, B3, C1, C2 and C3 moorings. Individual current meters are named according to their position on the mooring, with "1" denoting top mooring and "4" denoting bottom mooring (hence "C2-3" denotes the current meter at 1880 m on the C2 mooring).

Fourteen out of the 20 current meters returned complete datasets of roughly 380 days duration in both velocity and temperature. In addition, current meter C1-1 operated successfully for roughly 200 days before failing. These 15 datasets are the focus of the present work. The remaining five current meters, denoted by an F in Fig. 2, returned time series of varying quality and were excluded from our analysis. It is necessary in the evaluation of the energy equation to compute derivatives in both horizontal directions and this requires at least three data points at a given depth. Note that this condition was met at all levels (and exceeded at levels I, II and IV) in spite of data loss (see Fig. 2).

The raw data were low-pass filtered using a Lanczos-type filter with a quarter-power point at 1 cycle/40 hours and an energy rejection factor of 10^{-6} at 1 cycle/12 hours. The filtered time series had an equivalent sampling interval of 6 hours. Further, velocity measurements are expressed in a Cartesian system with the x axis pointing in a "downstream" direction (see Fig. 1). The x axis in this coordinate system is rotated 40° anticyclonically from true north. Complete documentation of data processing methods can be found in Bane et al. (1988).

The top instrument on each mooring measured pressure which was, in turn, converted to a time series of the depth of the topmost current meter. Using these data, the low-passed temperature time series at the top two current meters of each mooring were corrected for vertical movement following the procedure of Hogg (1986). Velocity measurements were not corrected (also in accordance with Hogg 1986) and the two deepest current meter records were left unaltered in all variables (reflecting both reduced mooring motion and temperature gradients at those levels).

Average depth levels for the top current meters were estimated by averaging the top meter depth record. The second and third level instruments were then assigned average depths which were 500 and 1500 m greater, respectively, than the top instrument. Bottom depth fixes were obtained at both mooring deployment and recovery. These readings were averaged, and the average used to estimate the average depth of the fourth level instruments (recall that the bottom meters were 500 m off the bottom).

Density was computed from temperature by using an equivalent coefficient of compressibility which takes into account standard Gulf Stream T–S characteristics; i.e.,

$$\rho = \rho_0[1 - \alpha^*(T - T_0)]. \quad (1)$$

Here ρ_0 and T_0 are reference temperature and density, and α^* is the equivalent compressibility coefficient. For the observed temperatures, α^* was approximately 1.5 $\times 10^{-4}$ °C^{-1}, and this value was used throughout the analysis. Vertical mean density gradients at given depths were computed from a least squares fit of a quadratic polynomial to the mean density data.

All statistics (i.e., means, variances and covariances) from a given current meter were assigned to the average depth of that meter. The computation of horizontal derivatives of these statistics, as is necessary in the evaluation of the energy equation, requires data at common depths. So-called "standard" depth statistics were computed for the top two levels by linear interpolation from the average current meter depths to depths of 380 m and 880 m. These depths were chosen subjectively, but required minimal interpolation at two of the moorings.

In an attempt to linearly interpolate the data at level III to a standard depth of 1880 m (using levels II and III), it was found that the results were dominated by the results at level II. This occurred even at moorings where the level III instrument was only a few meters away from 1880 m. These results were judged unacceptable; hence, the original and uninterpolated level III data are used in this analysis. Similarly discouraging results were found at level IV. In view of this, and given that the deep current meters appear to reside in a dynamically distinct current regime (i.e., the Deep Western Boundary Current), the original bottom data were analyzed.

Errors for the statistics at the average current meter depths were computed using a modification of the technique employed by Hogg (1983), or see the Appendix. Errors at the standard depths were calculated using standard formulas for error propagation (Bevington 1969).

Horizontal derivatives of a statistic q were computed at the standard depths using the formula

$$\delta q_{i,j} = q_x \delta x_{i,j} + q_y \delta y_{i,j} + O(\delta x^2, \delta y^2) \quad (2)$$

where $\delta q_{i,j}$ is the measured difference in q between the ith and jth current meters and $\delta x_{i,j}$ and $\delta y_{i,j}$ denote the mooring separation in x and y. A least-squares fit was used to calculate the gradients and errors were computed in the standard way from the uncertainties in the estimates of q_i. Physical data pertinent to the current meter moorings are given in Table 1. All relevant means and covariances, as measured by the current meters, are listed in Table 2.

TABLE 1. Mooring data. (Separations: C1-C2, 24.1 km; C1-B2, 58.1 km; C2-C3, 42.6 km; C2-B3, 66.7 km; C3-B3, 46.0 km; C1-C3, 66.7 km; C1-B3, 84.0 km; C2-B2, 52.3 km; C3-B2, 62.3 km; B2-B3, 41.6 km.)

Mooring	Location	Deployment (1984)	Recovery (1985)	Meter	Depth (m)
B2	35°48'N 73°26'W	10 Jan 1984	21 Jan 1985	1	398
				2	898
				3	1898
				4	3060
B3	35°31'N 73°08'W	10 Jan 1984	21 Jan 1985	1	306
				2	806
				3	1806
				4	3170
C1	36°15'N 73°10'W	11 Jan 1984	21 Jan 1985	1	379
				2	879
				3	1879
				4	2990
C2	36°06'N 72°59'W	11 Jan 1984	22 Jan 1985	1	380
				2	880
				3	1880
				4	3150
C3	35°48'N 72°43'W	11 Jan 1984	22 Jan 1985	1	360
				2	860
				3	1860
				4	3400

3. Results

Here, we briefly derive the mean energy equation and define its terms. We assume the ocean is adequately described as a Boussinesq fluid and ignore terms involving friction. It should also be noted that our analysis will of necessity be incomplete. For example, we have no direct measurements of vertical velocity (w) and will routinely ignore quantities in the mean energy equation which involve w. Although this neglect is suspect, we provide some scaling arguments which support our interpretations. Finally, the results of the calculations are summarized in Table 3.

a. Mean kinetic energy

The equation which is the focus of this paper is the mean kinetic energy (MKE) equation and is obtained by ensemble averaging the horizontal momentum equations and vector multiplying the result with the ensemble averaged velocity. The result is

$$\underbrace{\frac{\partial}{\partial x_i}\left[\langle u_i\rangle \frac{\langle u_\alpha\rangle^2}{2}\right]}_{\text{DMF}} = \underbrace{-\langle u_\alpha\rangle \frac{\partial}{\partial x_\alpha}\langle P\rangle}_{\text{MPW}} \underbrace{- \langle u_\alpha\rangle \frac{\partial}{\partial x_j}\langle u'_\alpha u'_j\rangle}_{\text{AME}} \quad (1)$$

TABLE 2. Flow statistics. The upper number in each category is the measured value of that statistic and the lower number is the uncertainty.

Current meter	u	v	T	$u'u'$	$v'v'$	$u'v'$	$u'T'$	$v'T'$	$T'T'$	$u'T'T'/2$	$v'T'T'/2$	$u'u'u'/2$	$u'v'v'/2$	$v'u'u'/2$	$v'v'v'/2$	z
B21	51.3	2.97	17.7	597	314	−241	4.12	−2.22	.42	−7.70	3.79	−3055	−3510	1676	4416	398
	6.9	3.09	.08	99	90	65	3.15	1.97	.20	5.52	3.14	4564	2596	2656	2556	9.6
B22	21.0	1.73	7.89	104	54.9	−27.5	2.7	1.3	.73	−1.7	−.03	−426.1	−139.7	118.8	90.6	897.7
	2.29	1.61	.16	22	8.2	11.0	1.1	.70	.14	1.1	.7	384.2	123.5	171.8	96.8	8.5
B23	5.6	2.09	3.86	38.7	20.2	−.7	−.117	.093	.006	.001	−.002	−42.8	−3.1	13.9	10.03	1896
	1.3	.73	.018	9.8	2.9	1.7	.040	.049	.001	.003	.003	86.7	10.9	12.7	17.02	5
B24	.67	1.0	2.6	42.4	18.0	13.0	−.23	−.06	.02	−.05	−.007	21.9	50.2	42.5	53.8	3060
	1.1	.45	.04	5.2	3.0	2.9	.13	.08	.004	.01	.005	50.3	20.4	20.4	20.4	2
B31	11.6	8.0	18.3	301.4	440.7	−92.4	2.1	−3.6	.35	−.62	2.7	−1939.	−672.6	−1444.	6846.	306.3
	4.0	2.7	.06	90.8	154.7	84.3	1.2	1.0	.13	1.0	1.4	2814.	582.2	728.8	3604.	3.0
B32	8.3	4.1	11.	93.3	82.9	−33.5	2.9	−1.9	1.2	−2.1	1.09	−136.1	−127.1	55.6	163.	806.8
	2.0	1.5	.14	19.0	22.6	9.7	1.2	.7	.39	1.2	.75	256.7	46.4	46.8	118.	1.5
B34	−3.2	1.6	2.4	61.5	16.7	8.6	−.07	−.04	.005	−.003	−.001	151.2	19.2	65.8	20.3	3170
	.02	.47	.02	12.9	2.5	5.6	.04	.02	.003	.003	.002	117.7	14.6	44.0	13.9	1.0
C11	67.0	−11.7	17.3	1431	489.4	−618.2	−12.5	7.6	.72	7.8	−5.8	−8697	−826.2	3891	−1175	379.3
	13.7	7.4	.26	253.7	109.9	118.4	6.6	4.7	.53	12.1	9.1	23225	6196	9022	5895	24.6
C12	22.6	.76	6.4	119.8	68.0	7.0	1.6	4.6	11.3	18.1	9.8	−440.1	−29.2	183.6	57.4	879.0
	1.8	1.4	1.1	32.5	9.3	11.7	3.9	2.3	4.9	13.2	6.6	712.2	85.7	123.1	116.3	20.0
C13	6.7	−.13	3.8	41.8	36.8	−10.9	−.28	.24	.01	.002	−.0001	−49.4	−10.5	−21.2	−8.1	1879
	1.2	1.0	.03	11.8	7.0	1.6	.09	.05	.001	.008	.007	156.4	13.6	26.7	42.3	12
C14	1.4	−.40	2.8	51.5	25.8	−20.8	−.22	.17	.04	−.02	.01	−.88	40.7	−53.3	−26.4	2990
	1.0	.78	.06	13.0	4.0	3.5	.21	.09	.005	.02	.01	174.2	26.9	58.2	28.5	3
C21	43.4	−4.26	17.9	684.2	329	−248.5	1.43	.60	.34	−4.34	.82	−3657.	−2192.	983.8	4184.	380.2
	8.1	2.87	.04	139.2	92.1	41.2	2.71	.60	.17	5.29	.83	7122.	3123.	2685.	2130.	12.2
C22	19.9	1.31	8.8	135.9	69.5	−27.3	−1.36	2.76	1.19	.69	−.81	−267.7	35.7	7.8	27.7	880
	3.4	1.88	.2	24.2	10.7	13.6	1.29	.60	.30	2.42	1.05	559.3	95.5	96.4	108.0	10
C24	1.0	−.77	2.6	45.0	32.3	−22.0	−.24	.18	.03	.004	−.005	71.5	27.3	−33.3	−.6	3150
	1.3	.70	.05	5.9	3.2	3.0	.15	.12	.003	.018	.013	82.7	37.5	36.6	33.7	3
C32	15.1	3.9	10.0	326.5	197.6	−101.3	.82	−.35	.76	−1.87	2.12	566.2	−221.6	−294.0	1550.	860
	4.0	2.0	.12	41.7	48.2	14.2	2.93	1.06	.24	2.20	1.00	1451.	350.3	411.0	879.	2
C33	1.0	1.0	4.0	36.3	21.8	−13.0	−.11	.02	.001	−.001	−.001	−19.0	−3.5	−7.2	−33.8	1859
	.8	.7	.007	8.5	5.0	3.3	.03	.014	.0004	.001	.0007	23.3	25.8	29.2	63.1	2
C34	−3.3	.60	2.5	34.1	35.6	−17.7	−.14	.05	.006	−.009	.003	−69.7	7.0	29.7	−94.2	3400
	.4	.57	.02	12.8	12.0	8.0	.11	.06	.003	.01	.006	57.8	14.9	25.1	91.3	2

where x_i are Cartesian coordinates, u_i denotes velocity in the ith direction and P denotes pressure. Summation notation is used and the indices i and j can obtain the values 1, 2 and 3. The index α can obtain the values 1 and 2 only. In more standard notation, x_1, x_2 and x_3 are represented by x, y and z, and u_1, u_2 and u_3 are denoted by u, v and w. Equation (1) equates growth and decay of kinetic energy, denoted by the divergence of the mean kinetic energy flux (DMF), to acceleration of the mean flow by the eddies (AME) and conversion from potential energy, here given by mean pressure work (MPW). AME measures the tendency of the eddies to deposit momentum in the jet and augments kinetic energy if the eddy momentum flux is convergent (divergent) in regions where the mean momentum is positive (negative). The structure of MPW shows clearly that it is related to mean flow in the direction of mean pressure gradients. Flow from low pressure to high is proceeding effectively uphill, thus increasing the fluid's potential energy. The opposite holds true for flow down a mean pressure gradient. The data from the Gulf Stream Dynamics Experiment allow us to directly estimate DMF and AME. In what follows, we will compute MPW as a residual of these quantities.

b. Level I

The level I current meters were at a nominal depth of 380 m and the mean flows at that depth are shown

TABLE 3. Summary of mean energy calculations. Units: erg cm^{-3} s^{-1}.

Level	DMF* $\langle\langle u\rangle\langle E\rangle\rangle_x + \langle\langle v\rangle\langle E\rangle\rangle_y$	AME $= -\langle u\rangle\langle u'u'\rangle_x - \langle u\rangle\langle u'v'\rangle_y - \langle v\rangle\langle u'v'\rangle_x - \langle v\rangle\langle v'v'\rangle_y$	MPW
I	$(-.24 \pm .24) \times 10^{-2} + (.85 \pm .91) \times 10^{-3}$	$= (-.20 \pm .09) \times 10^{-2} + (.27 \pm .08) \times 10^{-2} + (-.18 \pm .33) \times 10^{-4} + (.12 \pm .41) \times 10^{-4}$	
	$(-.15 \pm .26) \times 10^{-2}$	$= (.73 \pm 1.2) \times 10^{-3}$	$(-.22 \pm .29) \times 10^{-2}$
II	$(.44 \pm .20) \times 10^{-3} + (.02 \pm .16) \times 10^{-3}$	$= (-.63 \pm .43) \times 10^{-3} + (-.19 \pm .24) \times 10^{-3} + (.11 \pm .39) \times 10^{-4} + (.43 \pm .69) \times 10^{-4}$	
	$(.47 \pm .26) \times 10^{-3}$	$= (-.76 \pm .50) \times 10^{-3}$	$(.12 \pm .06) \times 10^{-2}$
III	$(.35 \pm .72) \times 10^{-5} + (.09 \pm .26) \times 10^{-5}$	$= (-.17 \pm .96) \times 10^{-5} + (-.33 \pm .21) \times 10^{-5} + (.21 \pm .11) \times 10^{-5} + (-.18 \pm .13) \times 10^{-5}$	
	$(.44 \pm .77) \times 10^{-5}$	$= (-.47 \pm 1.0) \times 10^{-5}$	$(.91 \pm 1.3) \times 10^{-5}$
IV	$(.44 \pm .46) \times 10^{-6} + (-.10 \pm .03) \times 10^{-5}$	$= (-.21 \pm .27) \times 10^{-5} + (.10 \pm .11) \times 10^{-5} + (.45 \pm .35) \times 10^{-5} + (.52 \pm .84) \times 10^{-6}$	
	$(-.56 \pm .57) \times 10^{-6}$	$= (.40 \pm .47) \times 10^{-5}$	$(-.46 \pm .47) \times 10^{-5}$

* $\langle E\rangle = (\langle u\rangle^2 + \langle v\rangle^2)/2$.

in Fig. 1. Note that the array resides in the Gulf Stream offshore, anticyclonic shear zone. Current meters B21, B31, C11 and C21 operated at this level. Note that B21, C11 and C21 were clearly in the mean Gulf Stream; therefore, in addition to the least squares calculations (which use all four meters) we report the values of the mean energy terms computed from these three meters. The latter are referred to as the Gulf Stream grouping.

The least squares estimate at level I of DMF was $(-0.15 \pm 0.26) \times 10^{-2}$ ergs cm^{-3} s^{-1}, while, according to the Gulf Stream grouping, DMF was $(-0.71 \pm 1.0) \times 10^{-2}$ ergs cm^{-3} s^{-1}. In view of the estimated uncertainties it is unclear that DMF is measurable. Perhaps the most believable result of these estimates is their sign (negative) since it is common to both estimates. A negative value for DMF implies a convergence in mean kinetic energy flux or, equivalently, that less energy is exiting a control volume surrounding our array than is entering it. A reasonable question is whether the magnitude of the DMF estimate $O(-1.0 \times 10^{-2}$ ergs cm^{-3} s$^{-1})$ is believable in spite of the large error bars. To this end, we suggest that such a DMF value can represent at best a local value. Evidence for this comes from our observation that the dominant contributor to DMF is the downstream gradient,

$$\mathrm{DMF} \approx \frac{\partial}{\partial x}\left[\langle u\rangle \frac{(\langle u\rangle^2 + \langle v\rangle^2)}{2}\right] \approx \frac{\partial}{\partial x}\langle u\rangle\left\langle\frac{u}{2}\right\rangle^2$$

suggesting that the decrease occurs primarily as a deceleration of the downstream flow. The eastward extension of the Gulf Stream beyond the experiment site is at least 1000 km. If either above value were typical of DMF over such a distance, a westward directed flow of 50–100 cm s^{-1} would be expected at the end of the section. This is clearly not reasonable. Perhaps the safest conclusion is that we see a tendency for DMF to be negative at 380 m, but in view of our errors, the magnitude of DMF remains unclear.

Our least squares estimate of AME at 380 m is AME $= (0.73 \pm 1.2) \times 10^{-3}$ ergs cm^{-3} s^{-1}, while the Gulf Stream grouping yields AME $= (5.1 \pm 4.3) \times 10^{-3}$ ergs cm^{-3} s^{-1}. The comparison between estimated value and uncertainty is more favorable here than for DMF (indeed the Gulf Stream estimate appears to be significant), so these estimates are taken to indicate an eddy acceleration of the mean flow.

The dominant contributor in both AME estimates is

$$-\langle u\rangle \frac{\partial}{\partial y}\langle u'v'\rangle$$

or equivalently, the net deposition of eastward momentum in the jet by cross-stream eddy transport. At all meters $\langle u'v'\rangle$ is negative as is

$$\frac{\partial}{\partial y}\langle u'v'\rangle;$$

thus, our estimate is consistent with more southward transport of eastward momentum at jet center than on the southern jet flank.

One of the interesting features of both sets of estimates is that they have the wrong signs to balance. The residual, R, of these terms, formed by bringing AME to the left-hand side of (1), is $R = (-1.2 \pm 1.1) \times 10^{-2}$ ergs cm^{-3} s^{-1}, for the Gulf Stream group and $R = (-0.22 \pm 0.29) \times 10^{-2}$ ergs cm^{-3} s^{-1} for the least-squares estimates. The sign of R is again common to both, and R is significant according to the Gulf Stream estimates.

Recall that we have no direct estimates of quantities involving vertical velocity, w. Examples are vertical kinetic energy flux divergence,

$$\left[\langle w\rangle \frac{(\langle u^2 + \langle v\rangle^2)}{2}\right]_z$$

and mean flow acceleration by vertical Reynold's stress, $-\langle u\rangle\langle w'u'\rangle_z$. It is possible, however, to estimate the order of magnitude of these quantities.

Upstream of Cape Hatteras, Dewar and Bane (1985) estimate $\langle w\rangle \sim -0.02$ cm s^{-1} at 219 m based on observations of mean flow crossing bottom isobaths. Osgood et al. (1987) find a comparable $\langle w\rangle$ estimate at 219 m from an analysis of the heat equation, and further suggest $\langle w'^2\rangle^{1/2} \sim 0.08$ cm s^{-1}. Downstream of Hatteras, Bower (personal communication 1989) finds $\langle w'^2\rangle^{1/2} \sim 0.08$ cm s^{-1} in the upper 500 m from an examination of RAFOS float trajectories in the Gulf Stream, and Hall (1986a) argues for comparable values at 68°W based on the GUSTO mooring observations.

According to our observations, mean kinetic energy changes from $\sim 10^3$ ergs cm^{-3} at 380 m to $\sim 10^2$ ergs cm^{-3} at 880 m. A worst-case estimate of

$$(\langle w\rangle(\langle u\rangle^2 + \langle v\rangle^2)/2)_z \quad \text{is thus:}$$

$$\frac{[(0.02 \text{ cm s}^{-1})\ 900 \text{ ergs cm}^{-3}]}{5 \times 10^4 \text{ cm}} \approx 4 \times 10^{-4} \text{ ergs cm}^{-3} \text{ s}^{-1}$$

which appears to be too small to provide a balance in the mean kinetic energy equation.

The other unknown quantities, for example

$$-\langle u\rangle \frac{\partial}{\partial z}\langle u'w'\rangle$$

involve correlations between horizontal and vertical velocities, for which apparently no direct estimates exist. On the other hand, as a worst-case estimate, assume at 380 m that $C_{u'w'} = 1$. We observe $\langle u\rangle \sim \langle u'^2\rangle^{1/2} \sim 50$ cm s^{-1} at 380 m. If we further assume $C_{u'w}$ goes to zero over 1000 m (again in assumption leaning towards a worst case),

$$\langle u\rangle\langle u'w'\rangle_z \approx 50 \text{ cm s}^{-1} \frac{(50 \text{ cm s}^{-1})(0.1 \text{ cm s}^{-1})}{10^5 \text{ cm}} \approx 2.5 \times 10^{-3} \text{ ergs cm}^{-3} \text{ s}^{-1}.$$

This is most likely an overestimate of vertical eddy acceleration of the mean, since the real value of $C_{u'w'}$ at 380 m is undoubtedly smaller than 1. In any case, the above estimate appears to be too small to provide a balance in mean kinetic energy balance inferred from the Gulf Stream grouping.

The only term remaining in MKE is MPW and it must act to restore a balance. We therefore argue at level I that MPW is negative and O(10^{-2} ergs cm^{-3} s^{-1}).

A negative value for MPW suggests a conversion to mean potential energy from mean kinetic energy and eddy acceleration of the mean, which occurs because of a flow up a mean pressure gradient, or equivalently here, a basically eastward flow proceeding into a region of increasing pressure. Given mean flows of O(50 cm s^{-1}), our MPW estimate can be converted into an estimate for the downstream pressure gradient of $P_x \sim 2 \times 10^{-4}$ gm cm^{-2} s^{-2}. If this is typical of the Gulf Stream extension region, it is consistent with a rise in sea surface of 20 cm over a distance of 1000 km.

The above pressure gradient estimate agrees in magnitude but is opposite in sign to that measured upstream of Hatteras by Sturges (1974) and inferred by Dewar and Bane (1985), both of whom estimated downstream pressure gradients of O[(2–3) $\times 10^{-4}$ gm cm^{-2} s^{-2}]. The only pressure gradient estimate downstream of Hatteras of which we are aware is that of Fofonoff and Hall (1983), who measured a net downstream pressure increase equivalent to a sea surface rise of $\sim$1.8 cm in 1000 km. This is substantially smaller than our estimate, but it should be recalled that Fofonoff and Hall's estimate is an average over the upper 3000 m of the water column. We are thus more heartened by the agreement in sign than we are concerned by the discrepancy in magnitude.

c. Level II

All current meters but C2-2 functioned at 880 m. The mean flows observed at level II are displayed in Fig. 3, from which it is clear that the array was still resident in the Gulf Stream anticyclonic shear zone at this depth.

DMF at level II is positive according to the least-squares estimate [DMF = $(4.7 \pm 2.6) \times 10^{-4}$ ergs cm^{-3} s^{-1}]. A divergent DMF at 880 m indicates that more kinetic energy is leaving a control volume about the array than is entering it.

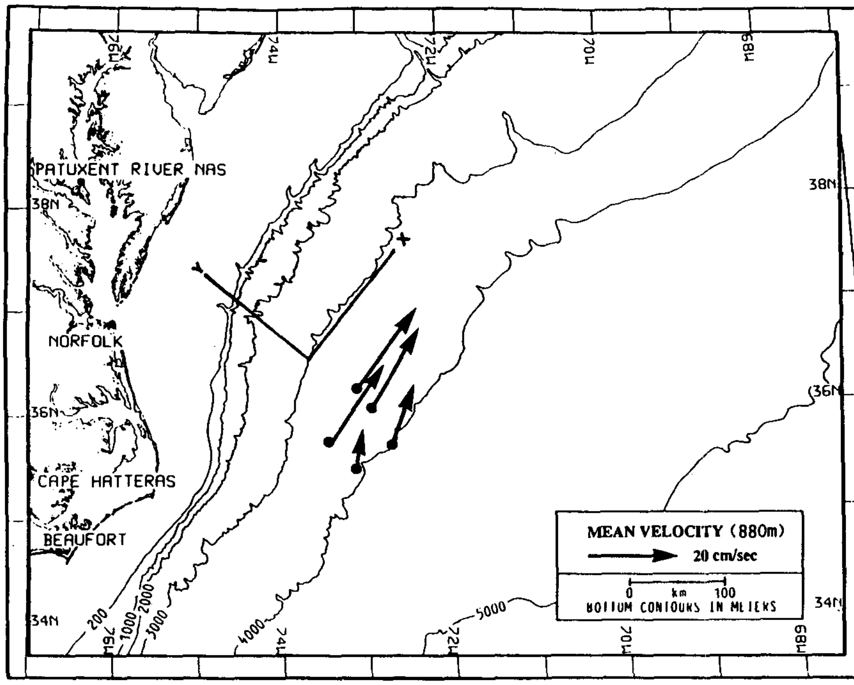

FIG. 3. As in Fig. 1, except the velocity vectors denote the mean flows observed at level II, i.e., 880 m. The mean flow measured at C2-2 is plotted although the quality of the behavior of the time series was suspect and the data were not used in the analysis. Note that at this depth the array is resident in the Gulf Stream anticyclonic shear zone.

The dominant term in the DMF estimate is

$$\frac{\partial}{\partial x}\left[\langle u\rangle \frac{\langle u\rangle^2+\langle v\rangle^2}{2}\right]$$

or, equivalently, a positive DMF is consistent with a downstream increase in mean flow speed. While this tendency seems reasonable, our estimate appears to be larger than an overall average DMF at 880 m. For example, although the Gulf Stream upstream at Hatteras is strongly constrained by topography, it still exhibits a reasonable mean flow on its offshore side at 800–900 m depth. According to Richardson et al.'s (1969) Cape Fear, North Carolina average Gulf Stream section, $\langle u$ (880 m)$\rangle \approx 20$ cm s^{-1}. Our $\langle u\rangle$ observations 250 km downstream are larger than 20 cm s^{-1}, but are less than 25 cm s^{-1}. If our DMF estimate were typical of the region downstream from Hatteras, a net increase in $\langle u\rangle$ of 10 cm s^{-1} would be required. Even allowing for uncertainties, this increase is not observed. On the other hand, our DMF estimate is not too large to be a local value. The average DMF from Hatteras to 73°W required by the above comparison is DMF = 1.6 $\times 10^{-4}$ ergs cm^{-3} s^{-1}, which is only a factor of 3–4 smaller than our local DMF estimate.

The AME estimate at level II is negative [AME $= (-7.6 \pm 5.0)\times 10^{-4}$ ergs cm^{-3} s^{-1}]. It therefore appears that the eddies at 880 m are organized so as to decelerate the mean flow. Two terms dominate the estimate, namely:

$$-\langle u\rangle \frac{\partial}{\partial x}\langle u'u'\rangle, \quad -\langle u\rangle \frac{\partial}{\partial y}\langle u'v'\rangle$$

with the former generally greater in magnitude. These quantities express the tendencies of the eddies to extract eastward momentum from the Gulf Stream jet by an increased downstream flux of downstream eddy momentum and, less importantly, by a removal of eastward momentum by cross-stream eddy transport.

Again, the AME estimate is bigger than larger-scale average values. The downstream momentum flux gradients are positive, in agreement with the intuitive idea that eddy variance grows in the downstream direction; however, the rates are consistent with an increase in $\langle u'^2\rangle^{1/2}$ from $\sim$14 cm s^{-1} at our site, to $\langle u'^2\rangle^{1/2} \approx 35$ cm s^{-1} at roughly 68°W. In contrast, Hall (1986a) reports $\langle u'^2\rangle^{1/2} \approx 22$ cm s^{-1} at 68°W based on the GUSTO mooring measurements.

As at level I, DMF and AME are not structured so as to balance: R at 880 m appears to be positive and significant [$R = (1.2 \pm 0.56)\times 10^{-3}$ ergs cm^{-3} s^{-1}]. Our results indicate the need for other processes in MKE in order to provide a balance.

We can estimate the magnitude of the processes involving vertical velocity in the same manner as was done at 380 m. Here, we take w (880 m) estimates from the work of Hall (1986a). She estimates w (875 db) $\approx 4\times 10^{-3}$ cm s^{-1} and $\langle w'$ (875 db)$^2\rangle^{1/2} = 8 \times 10^{-2}$ cm s^{-1}. Based on our mean flow data, the divergence of vertical energy flux is approximately

$$\frac{\partial}{\partial z}\left[\langle w\rangle \frac{(\langle u\rangle^2 + \langle v\rangle^2)}{2}\right] \approx \frac{(4 \times 10^{-3}\ \mathrm{cm\ s^{-1}})\ 200\ \mathrm{cm^2\ s^{-2}}}{5 \times 10^4\ \mathrm{cm}} \approx 1.6 \times 10^{-5}\ \mathrm{ergs\ cm^{-3}\ s^{-1}}$$

and the acceleration of the mean flow by vertical eddy momentum flux is

$$\langle u\rangle\langle u'w'\rangle_z \approx 10\ \mathrm{cm\ s^{-1}}\ \frac{10\ \mathrm{cm\ s^{-1}}\ 8 \times 10^{-2}\ \mathrm{cm\ s^{-1}}}{5 \times 10^4\ \mathrm{cm}} = 1.6 \times 10^{-4}\ \mathrm{ergs\ cm^{-3}\ s^{-1}}.$$

The latter estimate is again a worst case, having assumed that u' and w' are perfectly correlated. Note, at 880 m both effects are much too small to provide a balance in MKE; therefore, we conclude that MPW must essentially account for the residual: MPW $\approx (1.2 \pm 0.56) \times 10^{-3}$ ergs cm^{-3} s^{-1}.

The inferred positive value for MPW indicates a net conversion from potential energy at this level or, equivalently, that the Gulf Stream is flowing down a mean pressure gradient. Given our $\langle u\rangle$ observations of $\sim$25 cm s^{-1}, we calculate a downstream pressure gradient of

$$\frac{\partial P}{\partial x} \approx -5.0 \times 10^{-5}\ \mathrm{gm\ cm^{-2}\ s^{-2}}$$

which, if it were expressed entirely in terms of free surface variation, translates approximately into a free surface drop along the Gulf Stream from Cape Hatteras to our experimental site of slightly more than 1 cm. In terms of internal isopycnal variations, our pressure gradient suggests drops on the order of 10 m. These estimates are more in line with the magnitude of Fofonoff and Hall's estimates (although they are of opposite sign), and thus represent reasonable pressure variations.

d. Level III

The current meters B3-3 and C2-3 malfunctioned at level III, leaving only one possible grouping, consisting of meters B2-3, C1-3 and C3-3. The estimate of MKE at this level was therefore based on these three meters only. The observed mean flows at level III are shown in Fig. 4. The B2 and C1 meters indicate a weak $\sim$4 cm s^{-1} downstream flow; thus, level III appears to reside in a deep remnant of the Gulf Stream.

DMF at level III is positive and dominated by the downstream gradient. This by itself is indicative of a growth in downstream flow, but the estimate is of questionable significance. At level III, DMF = $(4.4 \pm 7.7) \times 10^{-6}$ ergs cm^{-3} s^{-1}. Nonetheless, independent evidence suggests that this estimate is reasonable in sign and magnitude. Assuming the estimate is believable, it suggests an increase in downstream flow from 4 cm s^{-1} at our experiment site to 7 cm s^{-1} 500 km downstream. This is perhaps a little large, given

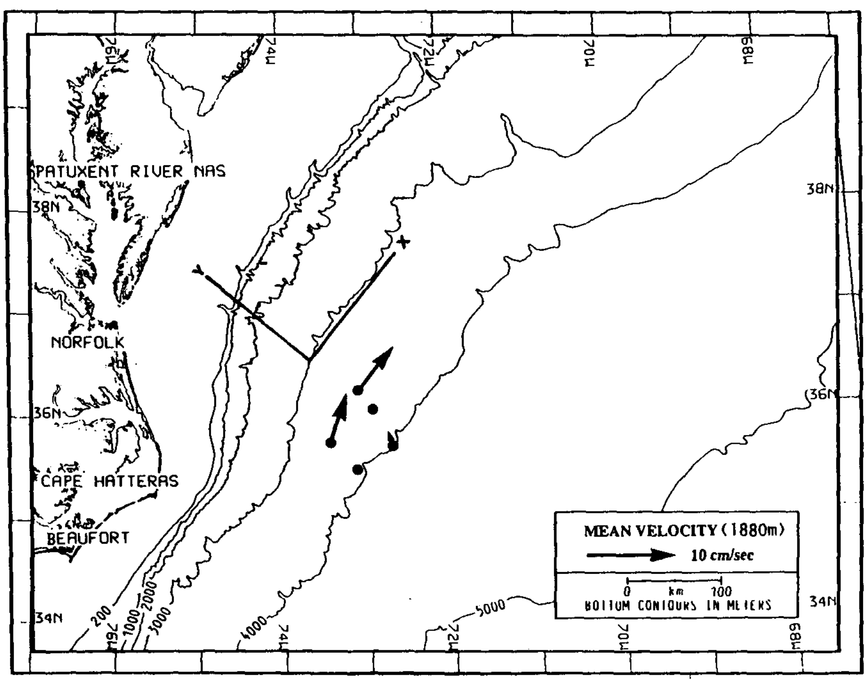

FIG. 4. As in Fig. 1, except the velocity vectors denote the mean flow observed at level III, i.e., 1880 m. Note the evidence of a weak Gulf Stream remnant at this depth.

Hall's observations of $\langle u \rangle \sim 5$ cm s^{-1} at 2000 db at 68°W, but is not entirely unreasonable. At Cape Hatteras (i.e., upstream of our site), the Gulf Stream is easily identified to depths of roughly 1000 m, but it is unclear if it is present at 1880 m. Here, for the sake of argument, we assume a vanishing mean flow at 1880 m off Cape Hatteras. If our DMF estimate is assumed to apply from Cape Hatteras to our experiment site, a change in downstream flow from 0 cm s^{-1} to $\langle u \rangle \approx 6$ cm s^{-1} would be required, and this agrees quite well with our C1-3 and B2-3 mean flow observations. We therefore feel there are reasons for accepting our DMF estimate, even if our computed uncertainty is large.

Our AME estimate at level III is negative, indicating a deceleration of the mean flow by the eddies, but is similarly overwhelmed by its error. At level III, AME = $(-4.7 \pm 10) \times 10^{-6}$ ergs cm^{-3} s^{-1}. This estimate has no clearly dominant term, although

$$-\langle u \rangle \frac{\partial}{\partial y} \langle u'v' \rangle$$

is largest by a factor of two. Thus, to the extent that these noisy estimates are believable, cross-stream transport of zonal momentum is an important contributor to the mean energy balance at level III. The eddies appear to be removing momentum from the weak downstream flow which, in view of our negative values for $\langle u'v' \rangle$, requires a greater southward transport of eastward momentum at the flank of the jet than at jet center.

Again, in spite of errors, our estimates are of opposing signs and require other processes to balance MKE. Restoring the balance with MPW yields MPW = $(9.1 \pm 12.6) \times 10^{-6}$ ergs cm^{-3} s^{-1}. Thus our best guess for MPW is that it is positive, and indicative of a conversion to kinetic energy from potential energy due to flow down the mean pressure gradient. Of course, our error bars do not preclude the possibility that MPW vanishes. Nonetheless, our observations of $\langle u \rangle \approx 4$ cm s^{-1} imply a downstream pressure gradient of $P_x \approx -2.3 \times 10^{-6}$ gm cm^{-2} s^{-2}, which is equivalent to a downstream drop of 1 m over 500 km for an isopycnal surface.

e. Level IV

All current meters except C2-4 operated properly at level IV. The mean flows observed at this level are shown in Fig. 5, from which it is apparent that this level resides on the inshore, asnticyclonic shear zone of the southwestward flowing Deep Western Boundary Current (DWBC). [See Joyce et al. (1986) for a more global picture of the relationship between the Deep Western Boundary Current and the Gulf Stream.] A remarkable characteristic of the region just northeast

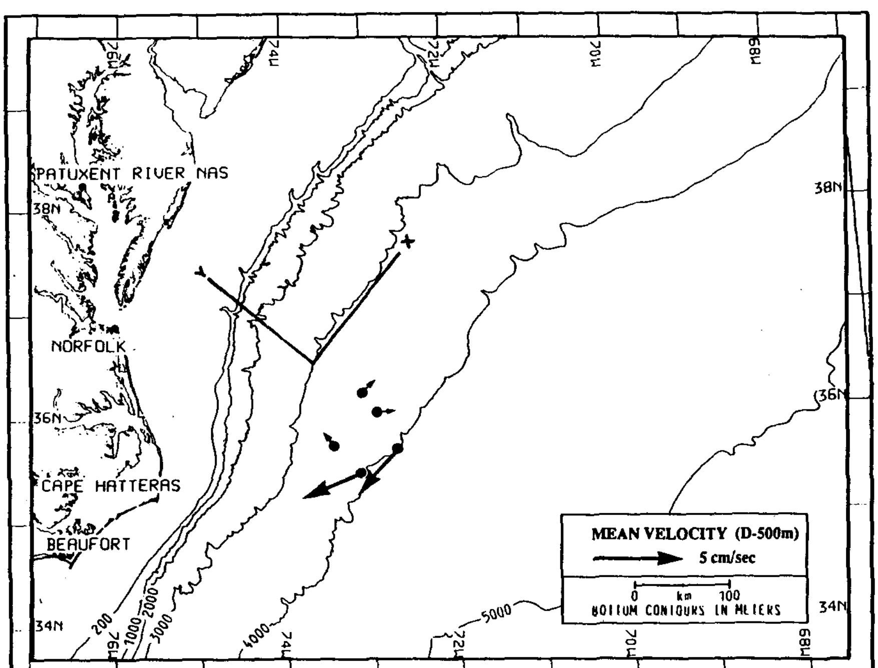

FIG. 5. As in Fig. 1, except the velocity vectors denote the mean flows observed at level IV, i.e., 500 m off the bottom. The mean flow measured at C2-4 is plotted although the time series was suspect and the data were not used in the analysis. Apparently, the array at this level was resident in the Deep Western Boundary Current. The weak northeastward flows at the inshore meters suggest the deep Gulf Stream remnant is inshore of the DWBC.

of Cape Hatteras is that it is apparently a transition region for these currents. For example, the so-called "south" section of Joyce et al. shows the DWBC offshore of the deep Gulf Stream, while their "north" section shows the DWBC inshore of the deep Gulf Stream. This relationship between the DWBC and the Gulf Stream has also been pointed out by Richardson and Knauss (1971) and Richardson (1977). Our experiment site is located physically between the two Joyce et al. sections but seems to measure mean flows more like Joyce et al.'s south section.

Our DMF estimate at level IV is marginally significant and negative [DMF = $(-0.56 \pm 0.57) \times 10^{-6}$ ergs cm^{-3} s^{-1}]. Thus the mean kinetic energy flux vector appears to be weakly convergent at level IV, or less energy leaves a control volume about the array than enters it.

Both x and y gradients in DMF are comparable in magnitude which makes the interpretation of this quantity less clear. On the other hand, the mean flow vectors in Fig. 5 are not particularly well aligned with either coordinate axis, so a reasonable interpretation is one of a downstream (here southwestward) decrease in mean flow speed.

AME at level IV appears to be positive but our estimate is not significant. The least-squares analysis yields AME = $(4.0 \pm 4.7) \times 10^{-6}$ ergs cm^{-3} s^{-1}. A positive AME indicates an acceleration of the mean flow by the eddies.

The dominant contribution to AME at level IV is $-\langle v \rangle \langle u'v' \rangle_x$, which is related to cross-stream momentum transfers, and the subsequent deposition of westward momentum in the jet core. This requires $\langle u'v' \rangle_x < 0$, given our observations of positive $\langle v \rangle$. Our direct estimate is $\langle u'v' \rangle_x = -6 \times 10^{-6}$ cm s^{-2} which, because it is so large, cannot be representative of much of the DWBC. For example, a momentum flux divergence of this magnitude over a distance of 300 km yields a net change in $\langle u'v' \rangle$ of -180 cm^2 s^{-2}, and nowhere at these depths are covariances of this magnitude observed. If these flux divergences are real, they must be local. On the other hand, some of our other eddy statistics are more typical. Our $\langle u'v' \rangle$ estimates are O(-10 cm^2 s^{-2}), which agree with those observed by Luyten (1977) and Thompson (1977), and our other flux divergence estimate, $\langle u'v' \rangle_y \sim 1 \times 10^{-6}$ cm s^{-2} agrees well with Thompson's (1978) estimates at 69°W and 70°W.

Again, DMF and AME are of opposite sign and we advance MPW to restore a balance in MKE. MPW at level IV appears to be negative [MPW = $(-4.6 \pm 4.7) \times 10^{-6}$ ergs cm^{-3} s^{-1}]. A negative MPW is consistent with a mean flow up a mean pressure gradient and an associated conversion to mean potential energy. It appears that this conversion is fed by both a convergent DMF and an eddy acceleration of the mean flow.

In view of our velocity estimates of O(1 cm s^{-1}) to the southwest, the above MPW estimate translates to a pressure gradient at level IV of $P_x \sim -6 \times 10^{-6}$ gm cm^{-2} s^{-2}. This gradient over a distance of 500 km is equivalent to a 3 m southwestward rise in an isopycnal surface.

4. Discussion

We feel our analysis indicates three structural aspects of North Atlantic western boundary currents. First, there is a clear suggestion that both the Gulf Stream and the DWBC exhibit considerable inertial character and are significantly influenced by ageostrophic flows. This interpretation is suggested because of the importance of MPW, an effect which would be absent if the flows were purely geostrophic. Our analysis also suggests at all levels that inertial effects are at least as important as, and perhaps dominate, eddy effects. Second, our results reveal a fair amount about the baroclinic structure of the Gulf Stream–DWBC system. This structure in turn suggests how the Gulf Stream changes between an upstream region of net acceleration and a downstream region of net deceleration. Last, the role of the eddies in mean Gulf Stream dynamics has been somewhat illuminated.

a. Gulf Stream structure

As an example, consider the results of the analyses at levels I and II, both of which were resident in the Gulf Stream. Our analysis suggests that the three quantities DMF, AME and MPW differ in sign between levels I and II. This indicates that the surface Gulf Stream is decelerating at our experiment site, while the deeper flow is still gaining in downstream momentum. This is a somewhat unexpected result about mean Gulf Stream baroclinic structure. Our analysis also suggests a reversal of the role of the eddies, from acting like an accelerator near the surface to acting like a brake at middepth. Commensurate with these role reversals is the role of MPW, which absorbs energy back to potential energy near the surface, and releases it at depth.

In our attempts to understand these results, we have been led to what are by now relatively well accepted ideas about the baroclinic structure of the large scale circulation. For example, Worthington (1976) describes the North Atlantic gyre in several temperature ranges, the results of which are summarized in several mass transport diagrams. Our observed 380 m mean temperatures were ~O(17°C), which places our level I in Worthington's warmest water circulation scheme. At 880 m we observed average temperatures of 7–10°C which places us in Worthington's 7–12°C layer. This layer is also the deepest of Worthington's layers which participates in the wind driven part of the recirculation, thus our upper two levels appear to reside within the wind driven gyre.

It is clear in the comparison of Worthington's layers that the North Atlantic circulation recedes to the north and becomes more zonally confined with increasing depth. Similar characteristics are observed in dynamic height patterns in both the North Pacific and North Atlantic (Reid 1981) and theoretical underpinnings for these observations have recently been advanced (Rhines and Young 1982; Luyten et al. 1983). We argue that our observations suggest a comparable baroclinic structure in the Gulf Stream.

To the extent that the Gulf Stream is an inertial current, consistent with our inferences about MPW, pressure gradients should exist on streamlines. Since the Gulf Stream is a northward flowing western boundary current, those pressure gradients should lead to a local pressure minimum on streamlines. In some sense, this minimum represents the inertial "center" of the boundary current and our data suggest this center varies in position as a function of depth (see Fig. 6). Our shallow observations are consistent with a flow up a pressure gradient and hence a location of the inertial minimum upstream of our site. Our deeper 880 m observations reveal a Gulf Stream which is apparently flowing down a mean pressure gradient, suggesting a location of the deeper Gulf Stream inertial center downstream of our array site. We therefore suggest that the baroclinic structure of the inertial Gulf Stream is characterized by a downstream shift of the inertial center with increasing depth. As further evidence of this

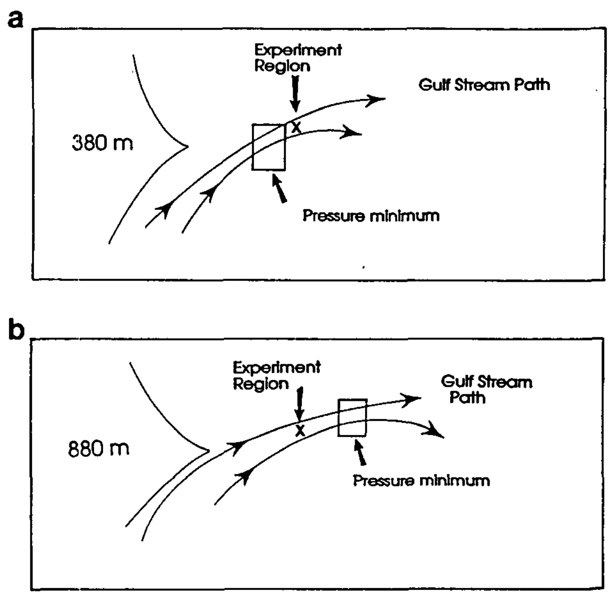

FIG. 6. Schematic of our inferred baroclinic Gulf Stream structure. (a) At 380 m, our results are consistent with an upstream location relative to our experiment site of a local pressure minimum, which we identify as the Gulf Stream inertial "center." (b) At 880 m, our results suggest a downstream location relative to our experiment site of the Gulf Stream inertial "center." We thus infer a barcolinic shifting of Gulf Stream structure reminiscent of the baroclinic shifting of North Atlantic general circulation.

trend, our 1880 m observations, although of questionable significance and apparently underneath the main body of the Gulf Stream at 73°W, indicate a similar balance to that observed at level II.

A characteristic of simple inertial jet models is that in the region of acceleration, the cross-stream length scale decreases in the downstream direction, while in the region of deceleration, the cross-stream length scale broadens downstream. In the real ocean we also expect Gulf Stream inertial pressure gradients to extend inshore to more linear regions where they should balance geostrophic flows. Accelerating regions are thus fed by an inward directed mass flux from both sides, and decelerating regions eject mass.

Halkin and Rossby (1985) have recently published averaged velocity profiles of the Gulf Stream at 73°W, based on observations made using the Pegasus profiler. They display average speeds in directions both normal and parallel to their transects and argue that the actual "mean" Gulf Stream flowed at an angle of $\sim 11°$ with respect to the direction normal to their transect. Interestingly enough, this angle seems to hold in the upper several hundred meters of the Gulf Stream, with the result that correcting their flow vectors yields an inconclusive result with respect to cross-stream divergences. On the other hand, there are indications at depths of 1200 m that mass is being fed to the Gulf Stream on its offshore side. Thus Rossby's data are not inconsistent with our suggested mean flow structure.

Since "downstream" near 73°W has a northward component, our inferred baroclinic Gulf Stream structure resembles the baroclinic changes in large scale structure. On the other hand, the dynamics of western boundary currents differ in important ways from the dynamics of gyre scale interiors, so it is unclear that our inferred northward migration of Gulf Stream structure with depth occurs for the same reasons as the large scale shift. We are unaware of any Gulf Stream models which exhibit such structure.

To our knowledge the only assessment of Gulf Stream mean flow energetics upstream of Cape Hatteras is that by Dewar and Bane (1985), who argued that the South Atlantic Bight Gulf Stream was accelerating. They had estimates at one depth and one location only, but it is tempting to assume that this observation is indicative of the Gulf Stream all along the continental margin. This is consistent with Sturges's (1974) observations of sea level slope in the South Atlantic Bight. The only energetic assessment of the Gulf Stream downstream of our site is that due to Fofonoff and Hall (1983), who argue for an appreciable decrease in downstream momentum flux in the Gulf Stream '60 region. They balance this decrease with a pressure gradient, hence their analysis shows a deceleration of Gulf Stream flow at 68°W.

Our observations thus apparently exist in a transition zone, and it is interesting to note that this transition apparently occurs baroclinically. Deceleration is observed first at the surface and is delayed at deeper levels, which are strongly topographically constrained upstream. Our interpretation of this region as a "transition" zone is also consistent with our observations of "local" DMF and AME at our upper two levels.

b. Level IV interpretations

Our DWBC observations, rather surprisingly, parallel our 380 m Gulf Stream observations, i.e., negative DMF, positive AME and negative MPW. It is interesting to place these deep observations within Thompson's (1977, 1978) theory about mean flow production by eddies. Thompson suggested, based on analysis of deep current meter observations at 69° and 70°W, that eddies were transferring momentum between the Gulf Stream and the nearby inshore countercurrent via cross-stream momentum flux convergence and divergence. Thompson further argued that this was possibly the source for the excess momentum needed to drive the mass transport in the Gulf Stream above the expected Sverdrup flow.

Our observed flow situation at 73°W is somewhat different than Thompson's, in that the DWBC appears here to be offshore of the deep Gulf Stream (see Fig. 5 and Joyce et al. 1986). Nonetheless, at level IV we also observe a net acceleration of the DWBC mean flow by the eddies and our net rates of acceleration are comparable to, if slightly larger than, Thompson's. Our observations thus support some of Thompson's ideas about deep eddy–mean flow interaction. On the other hand, we observe a convergent mean kinetic energy flux, which disagrees with Thompson's suggestion that the eddy momentum forcing feeds directly into fluid acceleration. Thompson argues by means of an averaging process around closed streamlines that pressure forces do not enter into the Gulf Stream momentum balance. While we agree that net pressure forces do vanish when globally averaged, in view of our observations of significant ageostrophic effects in the DWBC and Gulf Stream, we suggest that local pressure gradients might well form the primary sink for eddy momentum forcing.

Higher up in the fluid column (i.e., at 880 and 1880 m), our comparisons with Thompson's theory are more problematic. There is admittedly no Deep Western Boundary Current at 1880 and 880 m, but the so-called inertial recirculation exists at these depths and this resembles the countercurrent Thompson envisioned. Here the eddies attempt to decelerate the mean flow, and thus have an effect which is opposite to that forwarded by Thompson.

Since MPW is required in MKE at all four of our depths, our data suggest that the pressure gradient is involved in the observed downstream increase in Gulf Stream transport, possibly providing the energy necessary to accelerate entrained fluid to Gulf Stream speeds. This requires significant ageostrophic dynamics, but the recirculation region is generally thought of as inertial. This view also does not contradict Hogg's (1985) picture of the eddy-driving of the deep mean circulation, as the pressure gradient is incapable of imparting any net vorticity.

c. Effects of the Gulf Stream–DWBC transition

Our estimate of $\langle u'v' \rangle_x$ at level IV is much too large in magnitude to be characteristic of the DWBC. One can question the validity of this estimate, but it appeared to be significant [the least squares estimate was $\langle u'v' \rangle_x \approx (-6.2 \pm 0.7) \times 10^{-6}$ cm s^{-2}.] Aside from discarding this estimate, a possible explanation of it might focus on the strong changes which apparently occur in this region between the positioning of the DWBC and the Gulf Stream. Recall that in the immediate vicinity of 73°W, the DWBC apparently crosses under the Gulf Stream, thus moving from a basically inshore position northeast of our site, to an offshore position southwest of our site. Such a unique region should also be unique in its eddy statistics, consistent with our level IV $\langle u'v' \rangle_x$ measurements.

In closing, we note that our results demonstrate clearly the importance of eddy variability in the dynamics of the mean Gulf Stream, yet we have found the topic of time dependent inertial flow to be relatively unexplored theoretically. We are therefore in a comparatively weak position with respect to examining our results within a larger, dynamically consistent framework. On the other hand, our evidence suggests that the eddies affect the mean flow in relatively straightforward and robust ways. We identify the development of a theory of eddy–mean flow interaction in inertial Gulf Stream models as potentially the area of greatest gain in our continuing efforts to build an understanding of the oceanic Gulf Stream.

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APPENDIX

Error Calculations

The mean of an arbitrary statistic, q, is here defined by

$$\bar{q} = \frac{1}{T} \int_0^T q dt$$

where T is the total record length. As long as the time series is stationary, this is a consistent estimate of the ensemble mean, i.e., $\langle \bar{q} \rangle = \langle q \rangle$, where the angle brackets denote an ensemble mean. The question of error in our estimate of $\langle q \rangle$ involves the variance of q, i.e.

$$\sigma_{\bar{q}}^2 = \langle (\langle q \rangle - \bar{q})^2 \rangle$$

which after writing $q = \langle q \rangle + q'$ becomes

$$\sigma_{\bar{q}}^2 = \frac{1}{T^2} \int_0^T \int_0^T \langle q'(t) q'(s) \rangle dt ds.$$

Again assuming we are working with stationary time series, $\langle q'(t)q'(s) \rangle = R_{q'q'}(t-s)$, where $R_{q'q'}$ is the lag covariance of q'. After some manipulation,

$$\sigma_{\bar{q}}^2 = \frac{1}{T^2} \int \left(1 - \frac{\tau}{T}\right) R_{q'q'}(\tau) d\tau$$

which is an exact statement of the uncertainty.

Denoting the energy spectrum of q' by $E(f)$, the above becomes

$$\sigma_{\bar{q}}^2 = \frac{2}{T} \int^T \left(1 - \frac{\tau}{T}\right) \int_{-\infty}^{\infty} E(f) e^{2\pi \text{ if } \tau} df d\tau$$

$$= \int_{-\infty}^{\infty} E(s) \frac{\sin^2(\pi f T)}{\pi^2 f^2 T^2} df$$

after some algebra.

We have computed $E(f)$ for all statistics and performed the above integral numerically to represent our error.

Hogg (1986) uses the expression

$$\sigma_{\bar{q}}^2 = \frac{2I}{T} R_{q'q'}(0) \quad \text{where} \quad I = \int_0^{\infty} \frac{R_{q'q'}(\tau) d\tau}{R_{q'q'}(0)}$$

to estimate error. Our formula reduces to this if $E(f)$ is slowly varying for small f, in which case

$$\sigma_{\bar{q}}^2 \approx \mathrm{E}(0) \int_{-\infty}^{\infty} \frac{\sin^2(\pi f T)}{\pi^2 f^2 T^2} ds = \frac{E(0)}{T}$$

$$= \frac{2I}{T} R_{q'q'}(0)$$

recalling the Fourer transform pair relationship between $E(f)$ and $R_{q'q'}$.

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