Eddies, Islands, and Mixing

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As part of a field study of the relation between fine scale and large-scale variations of water properties in the western North Atlantic, the waters in the vicinity of Bermuda were investigated in detail. Previous work in the area had revealed regions of intense temperature fine structure confined to the sides of the island. Generally quieter levels of activity elsewhere in the midocean have suggested that significant mixing might only occur at the solid and fluid boundaries of the ocean. During the course of our investigation, two Gulf Stream rings were found in the vicinity of the island. The exchange of water between them caused three regions of strong alongshore flow. In these three areas we find elevated levels of temperature fine structure in the upper 800 m as measured by the variance in the temperature gradient normalized by the square of the mean temperature gradient over the interval. The normalized temperature variances on small scales (0.2–1 m) are most energetic in patches tightly bound to the island sides, whereas the fine structure on larger scales (5–25 m) is also energetic away from the island in a region of outflow. Velocity profiles show that vertical scales shorten as one approaches the island, and the energy increases in the counterclockwise component. There is no correlation evident between the shear measurements of the internal wave field and the intensity of the fine structure. Possible mechanisms for the production of fine structure are explored within the context of these observations.

INTRODUCTION

Thermal radiation and atmospheric forcing at the ocean surface are globally nonuniform. A fundamental question in oceanography is, 'How does the ocean redistribute this energy input?' One school of thought is that small-scale irreversible processes increase the effective diffusion rates in the ocean: water properties are transferred down their gradients more rapidly than molecular processes would allow but in a manner that is, effectively, diffusive in nature. Another school holds that even this effective or eddy diffusion is unimportant: mixing occurs predominantly at boundaries, and property redistribution results through density differences and pressure forces acting on fluid particles.

Using a diffusion upwelling balance to model temperature and salinity profiles within the main thermocline, *Robinson and Stommel* [1959] compute vertical mixing rates of the order of $1 \text{ cm}^2/\text{s}$. In a similar study of deep Pacific Ocean property gradients, *Munk* [1966] also finds that this balance between vertical upwelling and down-gradient turbulent diffusion models measured property distributions satisfactorily with indicated mixing rates of the order of $1 \text{ cm}^2/\text{s}$. Such estimates are model dependent and are, effectively, ocean basin averages. *Munk* remarks on the potential importance of enhanced mixing at boundaries with subsequent penetration into the interior along isopycnals with little mixing. This idea was stimulated by an earlier observation of *Iselin* [1939] that the vertical TS relation in the Sargasso Sea is similar to that of North Atlantic surface water. Thermocline theories with no vertical mixing [*Welander*, 1959] seem equally able to fit measured water properties. In this paper we investigate mixing processes near an isolated ocean boundary, namely, the island of Bermuda, and look at their relation to the ocean interior.

Eddy diffusion coefficients result from attempts to model small-scale processes, generally unresolved, in terms of Reynolds stresses [*Prandtl*, 1952]. Under conditions of steady state and lateral homogeneity, *Osborn and Cox* [1972] show that it is possible to estimate the additional vertical mixing by eddy processes. If K_e and K_m are the eddy and molecular diffusion coefficients, then

$$K_e = \frac{\langle T_z'^2 \rangle}{\langle \bar{T}_z \rangle^2} K_m \quad (1)$$

where $\bar{T}_z$ is the mean temperature gradient (measured over scales much greater than the dissipation scale) and T_z' is the deviation from the mean gradient. (The derivation actually is in terms of the potential temperature, but the difference for measurements in the main thermocline is slight.) Angle brackets indicate the ensemble average usually computed over a depth interval. This relation permits an estimate of the rates of actual vertical stirring by turbulent processes from measurement of the small-scale temperature fluctuations. The contributions to the total variance from each decade of wave number increase with wave number up to scales at which molecular processes can act to smooth out gradients (typically a few centimeters). *Gregg and Cox* [1972] have constructed instruments capable of measuring to these centimeter lengths, and their work in the Central Pacific main thermocline has shown, through use of (1), that eddy diffusion rates there are substantially less than $1 \text{ cm}^2/\text{s}$. Higher rates, associated with lateral intrusions, have been found in the California Current (see also *Gregg* [1975]). The limited observations from the Pacific imply that substantial mixing occurs only at coasts or at boundaries between differing water masses.

In the fall of 1975 a group of investigators conducted a study of the distribution of fine structure and microstructure at several locations in the western North Atlantic (North Atlantic fine structure and microstructure cruise [*Sanford and Hogg*, 1977]). The scientific objectives of this research were to explore regional differences in property gradients by using instruments of overlapping resolution with special emphasis on establishing the context or large-scale environment in which the fine structure and microstructure measurements are made. Part of the study concentrated on the waters around Bermuda as an example of an isolated ocean boundary well removed from boundary currents; this paper is primarily concerned with that aspect of the study. Previous work in the area, presented by *Wunsch* [1970, 1972] and *Hogg* [1972], had revealed isolated areas of enhanced fine structure activity ('fine structure' is used to denote scales of variation from 0.5 m to tens of meters in the vertical). On three separate cruises to Bermuda the anomalous region of fine structure was located on the left side of the island

Fig. 1a. Cruise track of R/V Eastward, October 1975.

if one looked downstream of the low-frequency, or 'mean,' flow. Wunsch [1972] explored a variety of possible explanations for the phenomenon, the most promising being one proposed by Hogg [1972] that the observed fine structure was a result of mixing induced by dynamic shear instability as the mean flow diverged around the island.

In the years since Munk suggested the importance of boundary mixing and the first work near Bermuda, we have come to realize that the middle of the ocean is not sluggish and slowly moving but that the low-frequency flows are dominated by mesoscale eddies (see, e.g., Crease [1962], Swallow [1971], Koshlyakov and Grachev [1973], Richman et al. [1977], Schmitz

[1976], and Freeland et al. [1975]). Products of mixing at the island can penetrate into the ocean interior not only by spreading along isopycnals but also through advection by the eddy field. Our observations are presented in the next two sections and concern the relation between fine structure, the eddy field, and the island of Bermuda. In the final section we attempt to evaluate possible mechanisms for fine structure production in terms of our observations.

MESOSCALE EDDY FIELD

The R/V Eastward departed Beaufort, North Carolina, on October 12, 1975, and sailed east along latitude 34°30'N until

Fig. 1b. XBT isotherms from Beaufort, North Carolina, to Bermuda, October 14-17, 1975.

Fig. 2. Dynamic height anomaly at 300 dbar with respect to 1000 dbar. Solid circles indicate CTD station positions. Numbered stations are referred to later in this paper. Open circles are mooring positions, solid arrows giving current vectors at 300-dbar depth at the beginning of the CTD survey and dashed arrows giving the current toward the end, 9 days later.

Fig. 3. Horizontal velocity at the surface as measured by a towed GEK at the end of the CTD survey.

Fig. 4. Current meter moorings (stick diagrams) and temperature series measured at about 300-m depth at three locations around Bermuda. Vertical lines straddle the period of the field experiment.

a ring was encountered at 70°W, whereupon it turned toward Bermuda (see Figure 1a). Expendable bathythermographs (XBT) were taken about every 2 hours along this track; from these a temperature section is plotted in Figure 1b. There are three prominent features: the Gulf Stream to the west, a Gulf Stream ring in the middle, and yet another ring at the eastern end of the track. This last ring (henceforth called ring A) is somewhat broader than the first and likely belongs to a class of rings whose source is suspected of being pinched off meanders in the Gulf Stream Extension [McCartney *et al.*, 1978].

Upon arrival in Bermuda waters on October 17 the *Eastward* commenced a detailed hydrographic survey of the island using the Woods Hole Oceanographic Institution (WHOI)-Brown conductivity, temperature, and depth (CTD) microprofiler [Brown, 1974] on a regular grid of 16 sites on each of four circuits around the island: along the 1000- and 2000-m isobaths and two outer circles of 50- and 75-km radius. The positions of these stations are shown in Figure 2 along with the dynamic height calculated at 300 dbar relative to 1000 dbar, contoured by using a quadratic interpolation algorithm. Ring A, previously observed in the XBT section, is responsible for strong currents on the northwestern slope of the island. The dynamic topography also rises toward the southeast into what we found later to be yet another ringlike eddy (henceforth

denoted ring B). Rings A and B appear to be exchanging water to the northeast and southwest of the island. We are not certain that the warm eddy indicated in this region is actually a closed circulation pattern rather than another area of cross flow between the two rings; just one station defines the associated minimum.

A geomagnetic electrokinetograph (GEK) was deployed at the end of this CTD survey and was towed in a zigzag pattern around the island. The inferred velocities, superimposed on the cruise track in Figure 3, confirm the eddylike flow off the southwest and the northwesterly flow across the northeast of Bermuda. It appears that by this time, ring A had moved further offshore toward the northwest, while the warm eddy had progressed along the coast toward the northeast.

In order to set the field work in the perspective of the low-frequency midocean flow, three moorings were deployed in April 1975 at the locations shown in Figure 2 and were recovered in January 1976 by the WHOI buoy group. Low-passed (24-hour Gaussian filter) temperatures and velocities at 300 m are presented in Figure 4 for the full 9-month exposure. The vertical lines straddle the period of the field experiment, and the times of passage of rings A and B are indicated. In terms of the ring phenomenon these are remarkable time series. The current and temperature records indicate the presence of four rings in their 9-month duration. The annual production of rings by the Gulf Stream has been estimated as being of the order of 10 [Lai and Richardson, 1977]. It would appear that as many as half of these pass close enough to the island of Bermuda to have some interaction with it.

Rings are derived from cutoff meanders of the Gulf Stream and retain their water mass characteristics in the upper water column [Wiebe and Ortner, 1976; McCartney *et al.*, 1978], where the azimuthal speed is greater than the translational speed so that there can be closed streamlines [Flierl, 1976]. Figure 5 presents potential temperature versus salinity plots for four stations within ring A and four within ring B. A contrast in water properties exists between the two rings, ring B being generally saltier in the water column. The four stations from ring B were taken about 1 week later in a repeat of the 50-km circle at which time the 15° isotherm had risen to a depth of 474 dbar, amounting to an upward displacement of about 100 m relative to the other stations on the circuit (150 m is the criterion used for ring identification by Lai and Richardson [1977]). The shallowest depth for the 15° isotherm in ring A is 497 dbar; it seems surprising that this ring has a fresh slope water core while ring B has not, at least within the region of our observations. A possible explanation is that ring B is older than ring A and has a smaller core region (P. L. Richardson, personal communication, 1977). In spite of rather vigorous efforts in recent years to track rings, ours is the only known sighting of ring B.

Another illustration of the water property contrast between the rings and the eddy flow pattern is given in Figure 6a, which displays the average salinity between the 26.0 and 26.5 potential density surfaces (roughly 100- to 350-m depth, the so-called 18° water region). Relatively fresh water is contained within ring A, and there is, apparently, transfer occurring between the two rings with freshwater flowing east on the south of the island and saltier water flowing west on the north, in a manner consistent with the geostrophic streamlines of Figure 2 and the GEK survey of Figure 3.

Qualitatively similar results are obtained deeper in the water column. In Figure 6b we show the average salinity between the 27.0 and 27.5 potential density surfaces (covering depths of

Fig. 5. Potential temperature versus salinity for stations taken within rings A and B. The portion between 12° and 15°C is not displayed.

about 700–950 m in the main thermocline), and in Figure 6c the average salinity is given between 27.7 and 27.8 potential density (approximately 1200- to 1600-m depth, bracketing the Mediterranean water). This latter map indicates that the 1000-dbar reference level chosen for dynamic height computations is probably too shallow, since there is still cross flow of saline water from southeast to northwest. (Flierl [1976] has demonstrated that rings transport water only above the level at which azimuthal velocities are greater than the translational speed of the ring. Generally, this level is somewhere in the main thermocline, so that one would not expect a ring to transport the deeper Mediterranean water. A plausible interpretation of Figure 6c therefore is that the passage of the rings is temporarily distorting a background salinity distribution, which contains a rather sharp gradient region passing through Bermuda, in which the salinity changes by about 0.05‰ over distances of the order of 20 km. The 4°C potential temperature surface of Worthington and Wright [1970] shows a similar change occurring over a distance of several hundred kilometers. The WHOI

CTD group has taken CTD sections along 28°N latitude, where a similar sharp transition occurs between 68°W and 65°W, and along 60°W longitude, where a noticeable change occurs at 34°N (R. C. Millard and G. H. Volkmann, personal communication, 1976). What maintains this Mediterranean water 'front' is unknown, but it is perhaps related to the horizontal shear in the return flow of the general circulation gyre that Worthington [1976] postulates to exist north of Bermuda.)

FINE STRUCTURE FIELD

Ship-Lowered CTD Observations

A principal motivation behind the field program near Bermuda was to investigate the source of previously observed fine structure and its relation to the larger-scale flow field. From the preceding section we conclude that the large-scale flow field is a result of mesoscale eddies or rings interacting with the

Fig. 6a

Fig. 6. Salinity averaged between potential density surfaces for (a) $26.0 < \sigma_\theta < 26.5$, (b) $27.0 < \sigma_\theta < 27.5$, and (c) $27.7 < \sigma_\theta < 27.8$. Arrows are inferred direction of water flow from Figure 2. Dashed line in Figure 6a is a towed CTD track line.

island. Rings are, almost certainly, also the cause of the strong currents observed in Wunsch and Hogg's earlier work near Bermuda.

In Figure 7a we plot temperatures versus depth between the surface and 1000 dbar for the inner two circuits on the 1000- and 2000-m isobaths starting at the south and working around the island clockwise. Some station positions are indicated in Figure 2. In Figure 7b are the first differences for the same stations taken from 0.1-dbar edited values further averaged over 0.5 dbar. We do not observe the 50- to 100-m-thick isothermal regions seen in the earlier surveys, but we do note, qualitatively, a change in character around the island with more structure present on the northwest slope, as evidenced by larger gradients and more negative values (inversions).

A quantitative way of presenting this variability is to compute wave number spectra of the first difference records and calculate variances over specified wave number bands. After correction of the first difference operation so that it represents a gradient, these variances, normalized by the square of the mean gradient to give the fine structure contribution to a 'Cox number' (see (1), where averaging is done over depth), are presented in map form (Figures 8 and 9). Two wave number bands, corresponding to wavelengths of 0.2–1 m and 5–25 m, are shown for the pressure intervals 200–400 dbar (Figure 8) and 600–800 dbar (Figure 9). (The standard WHOI editing procedure [Fofonoff et al., 1974] is based on a first difference criterion: if the difference between sequential data points exceeds a specified value, the offending point is replaced by linear interpolation. On the basis of considerable experience with open ocean profiles the threshold differences for temperature

and conductivity have been set at $\Delta T = 0.015^\circ\text{C}$ and $\Delta C = 0.03$ mmho below 200 dbar and values 5 times these selected for shallower depths. We discovered that a large number of data points for the near-island profiles were being removed by this procedure. Visual inspection of the records revealed nothing obviously wrong with the measurements, so these editing limits were increased fivefold everywhere above 2000 dbar.)

Several conclusions can be made. First, fine structure intensity in both larger and smaller scales increases as one approaches the island from any direction: the normalized variance levels are higher by an order of magnitude near the island than they are away from it. Second, there is a consistent variation along the island, largest variances being in the three locations where the low-frequency eddies force longshore currents (compare Figure 2 with Figures 8 and 9). Finally, there is a suggestion in the larger scales that relatively unstructured water exists near the island in regions of inflow and that structured water is found offshore in areas of outflow.

Profiles of temperature, salinity, and density between 200 and 400 dbar for representative CTD stations within the three active areas are given in Figure 10. Locations are shown in Figure 8b. Station 12 on the northwest is what Gregg [1975] calls 'irregularly steppey,' while stations 17 and 35 contain frequent inversions. These inversions are not strongly emphasized by the chosen plotting scale for temperature but are easily seen in salinity, which compensates the temperature to produce a monotonic density profile. Gregg [1975] has found in the California Current that substantial mixing occurs only in connection with these inversions. Although the CTD is not capable of resolving all the variance (Gregg finds that more

Fig. 6b

Fig. 6c

than 90% is contributed by scales smaller than those we resolve), it is intriguing that the fine structure intensity near the island is not directly correlated with the presence of inversions except perhaps in our larger-scale band (5–25 m).

Towed CTD Measurements

The R/V *Chain* arrived in the area at the time of the CTD survey. In response to preliminary *Eastward* findings it was

Fig. 7. (a) Temperature and (b) temperature difference versus depth for stations on the 1000- and 2000-m isobaths. Some stations are referenced in Figure 2. They start in the southwest and proceed around the island in a clockwise manner.

Fig. 8a. Contribution to the Cox number by small scales (0.2–1 m) for the 200- to 400-dbar depth interval. Contours are drawn to the 100-fathom (~200 m) contour.

decided that the *Chain* would tow its CTD on two lines through the indicated active area over the northwest slope. The depth-controlled tow fish system [Hess and Nowak, 1974], which consists of two fish, separated vertically by 8.5 m with

pressure and temperature sensors in both fish and a conductivity sensor in the lower fish, was operated so as to bracket the 17.5°C isotherm. This isotherm is at an average depth of 350 m and is likely to be representative of the water layer (150–

Fig. 8b. Same as Figure 8a except for large scales (5–25 m). Dashed line is towed CTD track line.

Fig. 9. Same as in Figure 8 except for the depth interval 600–800 dbar. (a) Small scale. (b) Large scale.

500 m) of low Väisälä frequency (2 cycles/h) between the seasonal and permanent pycnocline. The location of the continuous 24-hour tow is indicated in Figure 8b and in the insert of Figure 11a, each track covering about 100 km. The tracks are roughly parallel at a 15-km separation.

The results of the tow are shown in Figures 11a and 11b. In each figure the bottom graph indicates the interpolated/extrapolated depth of the 17.5°C isotherm whenever the tow fish either bracketed or one fish was within 5 m of the isotherm. Data are recorded at least once every 3 m along the towpath.

The next graph above is a trace of the temperature difference between the vertically separated sensors. Since the separation was constant to within a couple of percent, this represents a local gradient. Above this graph is the average salinity in the temperature band $17.460 \pm 0.005^\circ\text{C}$ computed for every half hour of tow. The vertical length of the line is one standard deviation about the mean. Finally, the uppermost graph records two methods of quantifying the temperature difference time series in a way which is not sensitive to the details of tracking (assuming statistical vertical homogeneity in a small band around the 17.5°C isotherm). One graph (solid line) is the variance of the temperature difference normalized by the square of the mean difference over a $\frac{1}{2}$ -hour interval (~ 5 km), while the bar graph is a measure of the percentage of the time the tow fish bracketed a temperature inversion (defined liberally as $\Delta T < 0.005^\circ\text{C}$).

Four segments of the tracks indicate an above-normal level of vertical fine structure by either of the two measures, as well as by simple examination of the ΔT time series. From Figure 11a, the track nearest the island, there is a 25-km track length of high variance and inversions. It is found where the track was closest to the 2000-m depth contour of the island. (A mean value of 0.09 for the normalized variance was obtained from a 14-hour track at a similar depth 300 n. mi. (~ 560 km) to the east of Bermuda in open ocean.) On the return track (Figure 11b) this active region is again well defined, though it is now confined to a 15-km segment, with even higher variance. (The increase in variance in the active regions was not accompanied by changes in the mean gradient.) Two smaller regions of activity are also noted, near the western end of the first pass and during the turn. They differ from the wider active area in that the variance is caused by a single inversion which, during the turn, is 500 m long.

The CTD station closest to the horizontal tows is station 40, for which the temperature, temperature gradient, salinity, and density are shown in Figure 12, plotted on the same scale as Figure 10. We note here the presence of numerous temperature inversions stabilized by salinity inversions. By referring to the dynamic height streamlines of Figure 2 and the salinity map of Figure 6a it can be inferred that the frequent inversions seen in this area result from an interleaving of water properties between rings A and B. These 10- to 20-m-thick features contribute greatly to the variance of the towed CTD and the lower wave number band of the lowered CTD but have a much subdued influence on the higher wave number band, which has a pattern much more tightly confined to the sides of the island.

Velocity Profiler Observations

The R/V *Knorr*, carrying a variety of fine structure and microstructure instrumentation (see Sanford and Hogg [1977] for details), conducted a three-component program consisting of observations in an open ocean area, around Bermuda, and within the Gulf Stream. One of the purposes of the open ocean observations was to provide a reference data set to compare with the Bermuda and Gulf Stream measurements.

After a 5-day series of measurements at the open ocean site (35°N , $66^\circ 30'\text{W}$) the *Knorr* arrived at Bermuda just as the CTD survey from the *Eastward* was being finished. Guided by the CTD information from the *Eastward*, the *Knorr* participants profiled an area to the north of the island and initially

Fig. 10. (Opposite) Temperature, salinity, and σ_θ versus depth for *Eastward* stations 12, 17, and 32. Locations are in Figure 8b.

proceeded counterclockwise around the island. Locations of the Bermuda stations are shown in Figure 13.

It was immediately apparent from the electromagnetic velocity profiler (EMVP) [Sanford *et al.*, 1974, 1978] observations that there was only weak flow around the island. There was, however, considerably more energetic, small-vertical scale shear here than had been seen at the open ocean site. Figure 14 presents all of the usable Bermuda EMVP data in terms of centroid wavelengths. The centroid wavelength is the inverse square root of the second moment of the one-sided shear spectrum normalized by the total shear variance. This quantity is a single wavelength for the velocity variations. For a Gaussian-distributed variable it is twice the expected depth interval between zero crossings of the detrended zero vertical mean velocity profile [Rice, 1945; Middleton, 1960].

Figure 14 shows that the centroid wavelength varies considerably even at one site (e.g., open ocean and mooring site). There is a trend toward shorter vertical scales in shallower water closer to Bermuda. This behavior is not in accord with WKBJ scaling of wavelengths proportional to N^{-1} , where N is the Brunt-Väisälä frequency given in Table 1. The shortest wavelengths and the largest shears were observed on drops 283–285 along the 1200-m isobath to the NW of the island in

the area midway between the zones of high normalized temperature variances as outlined from the *Eastward* survey (Figures 8 and 9). Curiously, however, the CTD profiles accompanying the velocity profiles show only a slight enhancement of temperature variability.

The shear spectra of Figure 15 show the trend toward more energetic high-wave number structure as a function of the four categories. The spectra of 80 consecutive shear values have been averaged over 13 wave numbers (zeroth and fortieth estimates are ignored) and over the profile set. The open ocean shear spectra decrease as wave number to the power -0.6 , a result consistent with the power law of approximately -2.5 for velocity spectra at these wave numbers found elsewhere in the open ocean profiles [Sanford, 1975]. There seems to be a transition at wave number ~ 14 c/km in that for smaller wave numbers the open ocean is more energetic than the other sites, while for large wave numbers the open ocean site is the least energetic. The conspicuous result of Figure 15 is that the slope of the spectra differ most between the first and second band-averaged estimates, while the slope differs little between the second and third estimates.

Since the results of Figures 14 and 15 are dominated by the structure of internal waves, rather than the longer vertical

Fig. 11a. Towed CTD measurements for near-island tow. Description is in text. Ticks on track on the insert chart are 20 km apart.

Fig. 11b. Same as in Figure 11a except for outer tow.

wavelength geostrophically balanced shear, we have computed the two-sided spectra for these data. *Leaman and Sanford* [1975] used rotary spectra to describe the sign of the energy flux in the open ocean. A spectral component is clockwise (CW) or anticlockwise (ACW) depending on the sense of the rotation of the velocity vector viewed from above of the velocity contribution of that component. In Figure 16 the CW shear variance changes less from site to site than that of the ACW. In the open ocean the ACW energy is a little greater than half the CW, while in shallow water (<2000 m) the ratio is 1. A clearer presentation of these results is obtained when the mooring site and the >2000-m profiles are combined. The reasons for combining these data are that the mooring site is part of the >2000-m profiles around Bermuda and that the number of profiles in each set is only 5 and 11, respectively. It is easy to see that the average of all 16 >2000-m profiles have a CW/ACW ratio intermediate between the open ocean and <2000 m.

CTD casts were made from the *Knorr* at times roughly coincident with a number of the EMVP drops (Figure 13). Although the quantity and the spacing of these drops are not adequate for mapping purposes, the normalized temperature variances show a pattern similar to that obtained by the *Eastward* with two exceptions. The active region to the northwest is less prominent, presumably a result of the later arrival of the

Fig. 12. Temperature, salinity, and σ_θ versus depth for *Eastward* station 40. Location is in Figure 8b.

Fig. 13. Locations of *Knorr* EMVP (upright numbers) and CTD (italicized numbers) stations in the Bermuda area.

Knorr and further translation of ring A offshore (velocity profiles show virtually no large-scale shear). Also, toward the end of the work, temperature activity increases to the southeast, where previous casts were relatively quiet. We believe that this is associated with further translation of ring B onshore (here velocity profiles show the development of substantial large-scale shear).

We had the preconception that the processes producing temperature fine structure were related to velocity shear and have attempted to find a relation between measures of shear and temperature variance. From Figures 8, 9, and 16 it can be seen that small scales of both fields exhibit the greatest island influence. In Figure 17 we show plots of contribution to the Cox number shear variance for the highest measured wave number bands in the depth intervals 200–400 and 600–800

dbar. There is no discernible correlation between the two quantities at either depth. In the deeper interval there is a hint that larger temperature variances are proportional to shear, but the converse is not true (higher shear variances are not related to higher temperature).

DISCUSSION

The large-scale temperature field, during our field experiment, was dominated by several Gulf Stream rings. Through interacting with each other and with the island of Bermuda, smaller (vertical) scales of temperature structure were produced. Temperature fluctuations at the smallest scales (0.2–1 m) resolvable by our CTD show greatest activity in regions close to the island, where the strongest alongshore currents occur. Energy in larger fine structure scales (5–25 m) is high

Fig. 14. Centroid wavelength of velocity variations from EMVP observations between 200 and 1000 dbar versus water depth at drop site. Solid squares are the group averages.

further into the ocean in a region of offshore flow. This outward extension may be an advective effect, but more plausibly it results from interleaving intrusions of contrasting water masses between the rings and the surrounding water. (Similar intrusive features have been observed elsewhere at water mass boundaries, for example, by *Gordon et al.* [1977] in the Antarctic Circumpolar Front, by *Gregg* [1975] in the California Current, by *Katz* [1969] in the Sargasso Sea, and by *Woods and Wiley* [1972] in the Straits of Malta.) Measurements of velocity shear also show an island influence with more energy present in smaller vertical scales and in anticlockwise motion near the island. Other than this change toward the island there appears to be little correlation between temperature and velocity variabilities. We interpret this as indicating that even in a field of active fine structure production, most of the fine structure observed is a remanent of previous mixing.

In viewing our observations as examples of boundary mixing that could occur wherever eddies interact with sufficiently steep slopes (e.g., islands, seamounts, continental slopes), it is useful to evaluate possible mechanisms in relation to the observations. Two general categories and several subclasses are considered below.

We have no evidence that doubly diffusive processes (salt fingering) are active in the region. In the ocean it has been observed in only a few special areas such as the Mediterranean outflow [*Williams*, 1974, 1975; *Magnell*, 1976] and under an Arctic island [*Neal et al.*, 1969]. *Gregg* [1975] suggests that salt fingering occurs in association with intrusions in the California Current. By inference, double diffusion may be occurring in the region of large-scale intrusions extending out from Bermuda to the northwest in our work. However, the intensity of our small-scale variability is not restricted to areas of large-scale intrusions (Figures 8 and 9).

Wunsch [1972] has suggested another diffusive mechanism which may be active around Bermuda. Wunsch studied the buoyancy boundary layers which result from the adjustment of a resting interior to zero heat and salt fluxes into a sloping bottom. The presence of strong geostrophic rings in the interior will certainly overwhelm this effect. A more likely diffusive mixing mechanism is the adjustment of the velocity field to zero velocity conditions at the boundary. The resulting turbulent boundary layer has been suggested to be responsible for producing well-mixed layers on the ocean floor [e.g., *Armi and Millard*, 1976]. Inspection of the near-bottom region of the

TABLE 1. Observational Data

	Number of Profiles	Average Water Depth, cm	Shear Variance, 10^{-6} s^{-2}			$\overline{N^2}$, 10^{-6} s^{-2}	Centroid Wavelength, m	Bulk Richardson
			CW	ACW	Total			Number, $\overline{N^2}/\text{Total}$
Open ocean	19	5150	4.77	2.49	7.26	20.0	147	2.75
Mooring site	5	4575	3.55	3.62	7.17	18.3	133	2.55
Combined*	16	3582	4.19	3.43	7.62	18.2	113	2.40
> 2000 m	11	3130	4.48	3.34	7.82	18.2	104	2.33
< 2000 m	23	1450	4.32	4.31	8.63	18.7	92	2.17

*Denotes combination of mooring site and >2000-m data.

Fig. 15. Shear spectral density versus vertical wave number in four different regions. Spectra have been calculated from data covering the depth interval 200–1000 dbar and averaged over wave number bands and drops. The 95% confidence limit is derived from a significance test between the open ocean and <2000-m groups. About the same limit would apply to a comparison of the average of the mooring site and >2000-m results with either of the other two groups.

CTD profiles, XBT on the 500-m depth contour, and the EMVP profiles has rarely revealed well-mixed bottom layers. Nevertheless, bottom-generated turbulence seems a viable mechanism and could account for our observed correlation between the eddy flow and the fine structure intensity. In this regard the hint of a correlation between temperature and shear variances at high temperature variances found for the deep water in Figure 17 is intriguing. The shallowest of the active stations is EMVP 298 at 916-dbar depth, where the bottom is still some 116 m beneath the 800-m level (or about 500 m to the side).

Dynamic shear instability is the other broad class of physical processes that has been invoked to explain enhanced vertical stirring rates in the ocean. The field work near Bermuda was undertaken on the hypothesis that the fine structure activity was a direct result of shear instability of the mesoscale flow field in preferential locations at the side of the island, as had been suggested by Hogg [1972]. Using the density data from the CTD survey, we have calculated a geostrophic Richardson number and find minimum values of about 20 in the regions of strong flow to the west and northwest of the island. The Chain tows have superior horizontal resolution, which is helpful in eliminating smaller-scale aliasing in estimating the isopycnal slope. A geostrophic Richardson number of about 75 is estimated from Figure 11 over the strong gradient region. The velocity profiles also show little large-scale shear except at the end of the Bermuda field work when ring B had entered the area. It seems unlikely that the eddy flow became unstable over depth intervals of 100 m or more.

There are, however, small Richardson numbers associated with smaller-scale variations of velocity, as is shown by the low bulk Richardson number column in Table 1. We believe that these are associated with the internal wave field, and there is a slight but consistent decrease in Ri as one approaches the island.

In the presence of a sloping bottom, internal waves of sufficiently high frequency will be transmitted toward shallower water [Wunsch, 1969] (the critical period for Bermuda is roughly 2 hours). Provided that these high frequencies are modal in character, they shoal much like surface waves:

shortening in wavelength, increasing in amplitude, and refracting so that crests more closely parallel the bottom contours. Eventually, the amplitude will be sufficiently large for the waves to be unstable, whereupon they break, causing mixing and fine structure (Cacchione and Wunsch [1974] have demonstrated this process in the laboratory). Waves having periods greater than the critical period will be reflected from the slope. A plausible interpretation of the spatial variation of the velocity structure observed by the EMVP is that the low-frequency clockwise polarized components are reflected, while the high-frequency waves are transmitted and amplified. From moored temperature and velocity measurements on the Bermuda slope, Eriksen [1976] has observed frequent occurrences of low (i.e., less than 1) Richardson number events coupled with the appearance of temperature inversions. He concludes that internal wave breaking is the agency for fine structure production. This process has been a favored one for the explanation of open ocean activity as well [Munk, 1966; Garrett and Munk, 1972b; Orlanski and Bryan, 1969].

The internal wave field in the open ocean is believed to be almost isotropically distributed [Katz, 1973; Garrett and Munk, 1972a, 1975]. Local, near Bermuda, topographic features could lead to a geometrically dependent focusing of energy and anisotropic fine structure production. Two facts speak against this possibility's being significant. First, the internal wave field, as measured by the EMVP, has no more activity in these active fine structure regions than in others close to the island. Second, at the end of our island survey, ring B had moved north, so that a strong westerly flow was being directed toward the island from the east. At this time, CTD stations (corresponding to EMVP drops 301, 304, and 316 in Figure 17) by the Knorr on the southeast coast recorded high levels of fine structure activity at a site where the records from two previous casts were quiet. This suggests coupling with the low-frequency flow rather than purely geometrical enhancement of open ocean internal waves. The low-frequency flow will cause additional refraction and focusing of energy (A. Garrett, personal communication, 1976) and could perhaps lead to breaking activity connected to both geometrical effects and strong flow (e.g., critical layer absorption).

It seems more likely that the internal wave field is involved in energy transfers between the low-frequency eddy flow and the fine structure. Here two mechanisms are suggested. Müller [1976] asserts from theoretical arguments that internal wave energy increases in strong low-frequency shears and predicts

Fig. 16. Shear variances, between 200 and 1000 dbar, decomposed into clockwise and anticlockwise components versus region over which drops are averaged. For each region the 95% confidence limits are shown for significant differences between CW and ACW variances.

Fig. 17. Normalized variance in small-scale (0.2–1 m) temperature fluctuations versus shear variance in small scales (20–30 m) at two-depth intervals. Numbers refer to EMVP stations shown in Figure 13. Station 290 in the indented region is part of the lower graph. Units for the shear variances are 10^{-6} s^{-2} .

effective vertical viscosities of the order of $10^8 \text{ cm}^2/\text{s}$ through relaxation of this field back to its undisturbed state. Interaction with the island will intensify eddy shears and therefore amplify this mechanism. Typical relaxation times are of the order of a few days, comparable to the residence time near the island. Ruddick [1977] has searched for evidence of this strong interaction in moored time series. Being unsuccessful, he proposes a countertheory in which the internal wave field relaxes not to its state in the absence of mean flow but to one appropriate to internal waves in the presence of a mean flow. This is a much weaker interaction consistent with measured vertical eddy viscosities of the order of $10^2 \text{ cm}^2/\text{s}$ in the main thermocline and relaxation times of the order of 10 days. It does not seem plausible that such a weak, slow interaction, one subject to rapid advection, would result in fine structure patterns so tightly bound to the island.

A second possible eddy-island interaction is the generation of lee waves by flow over the rough bottom. This process is well documented by laboratory experiments [Davis, 1969], theoretical work [Miles, 1968], and observations in the atmosphere [Smith, 1976] where it has been connected with the production of clear air turbulence. Two requirements for efficient lee wave generation are that the horizontal flow velocity u be of such strength that the Doppler-shifted wave frequency $u \cdot k$ be in the internal gravity wave frequency range and that there be topographic variations on the scale $|k|^{-1}$. By match-

ing the horizontal phase speed of the internal waves to u one can estimate the vertical scale as $\lambda_v \approx 2\pi|u|/N \approx 120 \text{ m}$ for a 10-cm/s current with $N = 5 \times 10^{-3}/\text{s}$. (This argument was suggested by M. Stern (personal communication, 1976).) Theoretically, Miles [1968] has shown for two-dimensional topography that if the height exceeds a value of the order of $2|u|/N$ ($\sim 40 \text{ m}$), rotors or overturnings will form in the lee. We have no detailed bathymetry of the Bermuda slope, but corrugations of this size are known to exist. The lack of correlation between temperature and velocity structure might seem to rule out lee waves as a viable mechanism. However, it must be remembered that the internal waves propagate, spreading their energy over great distances, while the fine structure products of wave breaking can only spread slowly through density contrasts and advection by the mean flow. Thus at times comparable to a wave period, we cannot expect a high correlation between temperature gradient and shear, and at times comparable to $L/|u - c_g|$ (c_g is the horizontal group velocity, $\approx 10 \text{ km/d}$ for a high-frequency wave with a vertical wavelength of 100 m), the internal wave energy will be spread over a distance L , and we should not expect even a statistical correlation between the two quantities.

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