Pacific Equatorial Turbulence: Revisited

J. N. MOUM

College of Oceanography, Oregon State University, Corvallis, OR 97331

T. R. OSBORN

Chesapeake Bay Institute, Shady Side, MD 20764

W. R. CRAWFORD

Institute of Ocean Sciences, Sidney, B.C. V8L 4B2, Canada

(Manuscript received 25 April 1985, in final form 27 January 1986)

ABSTRACT

Turbulence measurements from the central equatorial Pacific in February 1982 have been analyzed and compared to synoptic CTD and current velocity profiles and current meter data. These suggest considerably more time (if not space) variability than had previously been anticipated. Above 300 m at the equator the turbulence levels were greater but less than previous equatorial measurements, and turbulent patches occurred more frequently than elsewhere in the open ocean. Below 300 m the occurrence of turbulent patches was less frequent than in other regions of the ocean, except for the persistence of a patch at 500 m.

1. Introduction

Studies of equatorial turbulence in both the Pacific (Gregg, 1976; Crawford, 1982; Crawford and Osborn, 1981a) and Atlantic Oceans (Crawford and Osborn, 1979; Osborn and Bilodeau, 1980) have made an important contribution not only to basic turbulence studies, but also to our understanding of the processes which determine large-scale ocean dynamics. In fact, Crawford and Osborn (1979, 1981a) have shown that the energy dissipated by the turbulent friction in the shear flow, above the Equatorial Undercurrent (EUC), is sufficient to balance the work done by the zonal pressure gradient set up by the easterly wind stress. This unique feature of equatorial currents inspired additional measurements in 1982, described here, and the extensive microstructure measurements in 1984, during the Tropic Heat experiment—preliminary results from Tropic Heat are presented by Moum and Caldwell (1985) and Gregg et al., (1985).

During the Pacific Equatorial Ocean Dynamics (PEQUOD) experiment in February 1982, a set of velocity microstructure measurements were made to depths of 1000 m using the free-fall profiler, Camel III. Coincident White Horse profiles (from which horizontal currents, temperature, and salinity were obtained) gave us the opportunity to compare turbulence measurements with features of the large-scale flow field. We emphasize in this paper the measurements below 300 m depth.

2. Data

The instrumentation and data processing used for this experiment are described by Moum and Osborn (1986). A complete description of Camel III is given by Moum (1984). Mounted on Camel III were two airfoil probes (Osborn and Crawford, 1980) from which an estimate was obtained of the rate of viscous dissipation of turbulent kinetic energy, ϵ, (Osborn, 1974).

The February 1982 PEQUOD cruise involved the recovery of a number of current meter moorings plus a predetermined pattern of White Horse stations (or nets, as defined by the positions of the arrays of bottom transponders used to track the White Horse; see Luyten et al., 1982, for a discussion of the White Horse). In addition, Camel III casts were made coincidently with many White Horse profiles, providing a unique equatorial dataset which encompasses a depth range much deeper than the strong near-surface shear flow. J. Luyten kindly made these data available to us. Locations of current meter moorings, White Horse profiles, and Camel III casts are shown in Fig. 1. The majority of the profiles were deeper than 900 m.

a. Dissipation profiles

Two representative, nearly coincident Camel/White Horse profiles are shown in Figs. 2 and 3. Estimates of turbulent dissipation, ϵ, were made over approximately 2 m vertically (1024 data points), and are plotted as a bar graph on a $\log_{10}$ scale. Drop 2 (Fig. 2) was made

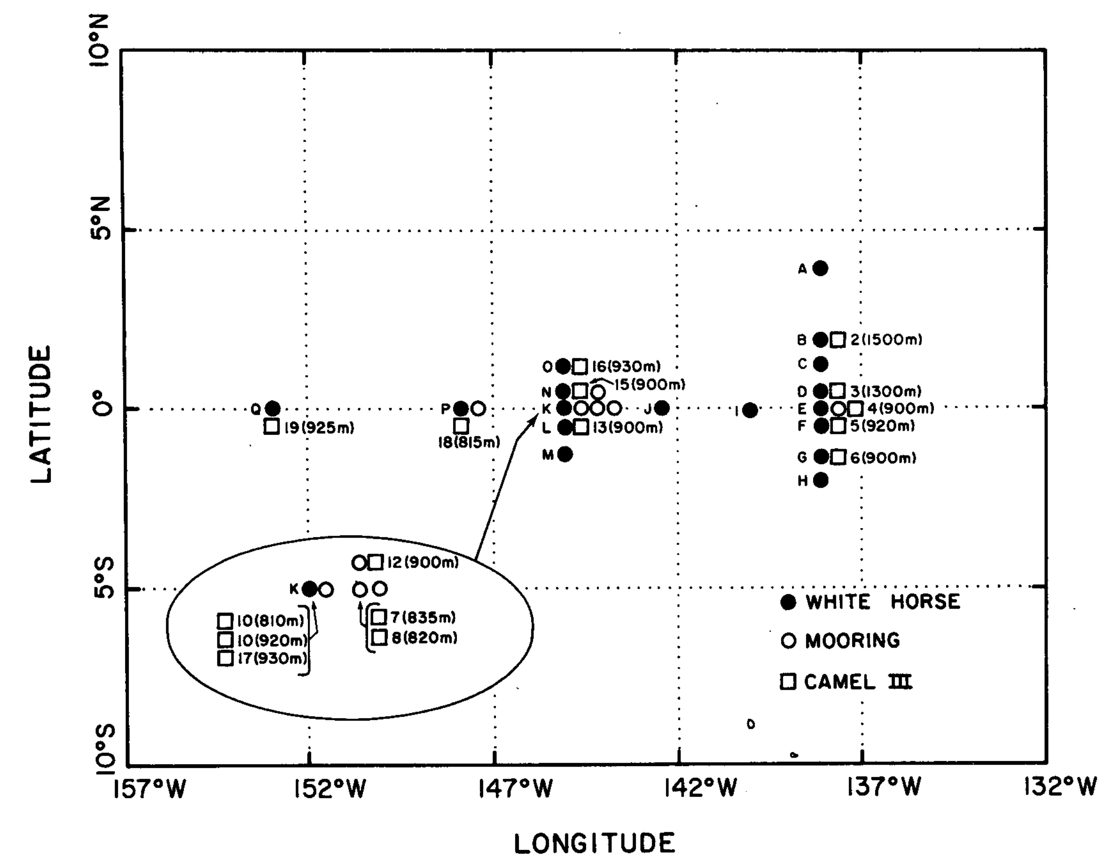

FIG. 1. Locations of White Horse nets, current meter moorings and Camel III profiles (February 1982). Depths of Camel III profiles are in parentheses beside the position.

at 2°N, 138°W. The White Horse velocity profile does not show an eastward undercurrent. There is, however, a significant southward velocity component centered at 100 m. At the maximum of the southward flow, the dissipation is a minimum, while in the high vertical shear above and below the maximum, are turbulent patches. Below 200 m, the patches are characteristically thin (2–10 m) with peak values of ϵ between 10^{-9} and 10^{-8} m^2 s^{-3}. (The kinematic unit m^2 s^{-3} represents energy rate of change per unit mass, and is equivalent to W kg^{-1}. The comparable dynamic unit, W m^{-3}, representing energy rate of change per unit volume is obtained by multiplying the kinematic unit by the fluid density).

A representative profile on the equator at 145°W is Camel drop 17 at White Horse net *K* (Fig. 3). The surface current is weak, and the White Horse data reveal a peak undercurrent of 1 m s^{-1} eastward. Centered at 492 m, is a 35 m-thick patch of turbulence which has a patch-averaged dissipation of 4×10^{-9} m^2 s^{-3}. We observed this feature in seven of the eight microstructure profiles made within ½° of the equator at 145°W during the cruise, but not elsewhere, either along or outside of ½° of the equator. Of these seven profiles, five show 15 to 35 m-thick patches centered within 20 m of 500 m, and have patch-averaged dissipations of $(1-5) \times 10^{-9}$ m^2 s^{-3}, while two show weaker patches at 500 m.

b. Averaged profiles

From the White Horse CTD data, buoyancy frequency, *N*, was computed over 20 m depth intervals. Individual profiles of *N* are shown in Fig. 4. The thick line represents the value averaged over all of the profiles at that depth. The data are shown only for those profiles made within ½° of the equator and for which a coincident Camel profile was made. The peak thermocline value was situated at about 110 m.

Computation of shear at 25 m intervals from White Horse velocity profiles was done by smoothing the velocity profiles with a three-point running mean, calculating the velocity magnitude as the square root of the sum of the squares of the two horizontal velocity components, and first-differencing over 25 m intervals. Resultant profiles are shown in Fig. 5 where, again, the black line is the averaged shear profile. The minimum in the averaged shear profile coincides with the peak value of averaged *N* from Fig. 4, and with the core of the EUC. The mean shear above the EUC core has a

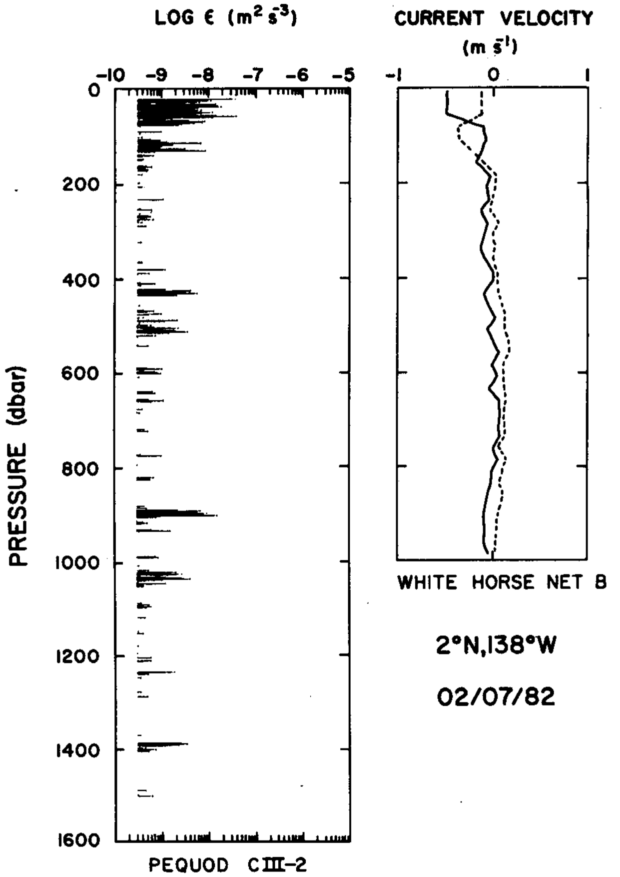

FIG. 2. Depth profiles of $\log_{10}$ of turbulent kinetic energy dissipation, ϵ, and current velocity at 2°N, 138°W on 7 February 1982. The solid line represents E–W velocities ($E > 0$), and the dashed line represents N–S velocities ($N > 0$).

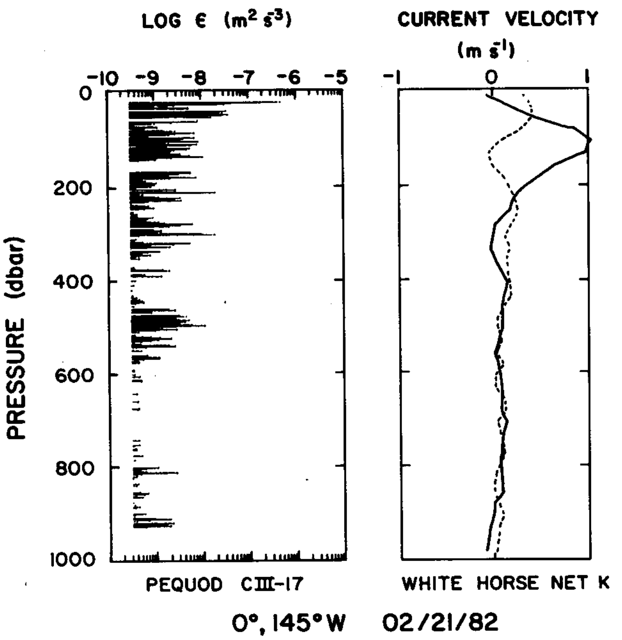

FIG. 3. As in Fig. 2 for 0°N, 145°W on 21 February 1982.

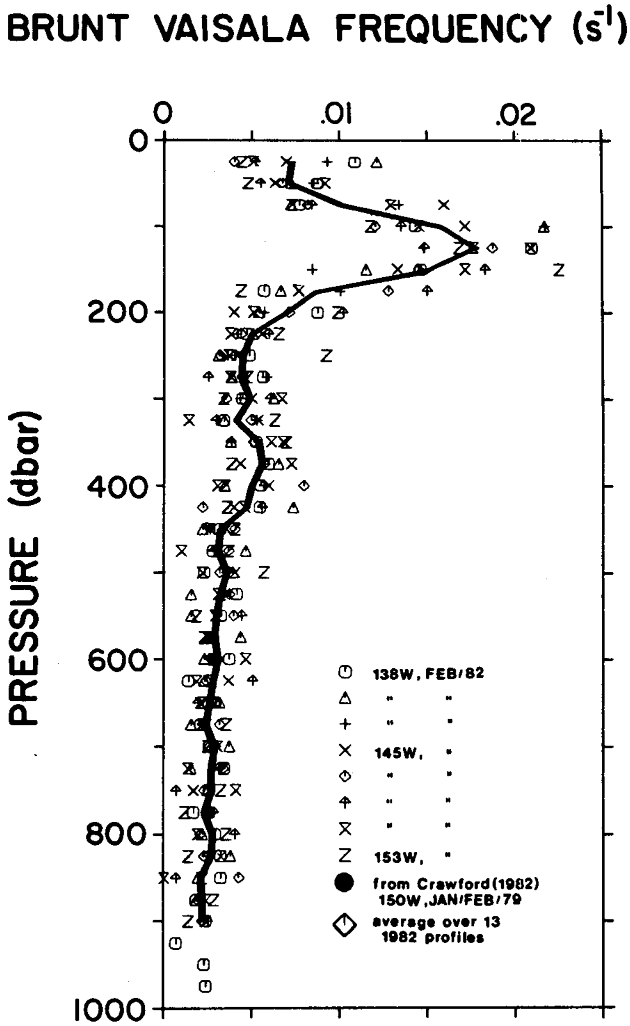

FIG. 4. Eight vertical profiles of Brunt–Väisälä (buoyancy) frequency within ½ degree of the equator in February 1982. The thick line is the average over the eight profiles. Longitudes listed in the key pertain also to Figs. 5 and 6.

peak value of about 0.014 s^{-1}. The averaged difference Richardson number (Ri), then, computed from N and shear values noted above, was less than one over the depth range 40 to 85 m. In fact, occasional estimates of Ri < ¼ were found in individual profiles. Below the EUC core, the mean shear was smaller by a factor of two, and estimates of Ri were considerably larger than above the core.

To compare with the data of Figs. 4 and 5, 2 m estimates of ϵ from those same eight profiles were averaged over 25 m depth intervals, averaged (black line) and plotted in Fig. 6. The dynamic range in 25 m averages of ϵ spanned five decades, from 2×10^{-7} m^2 s^{-3} in the large mean shear region above the EUC core to 3×10^{-11} m^2 s^{-3} below 700 m. Although Moum and Lueck (1985) have estimated the instrumental noise level to be approximately 3×10^{-10} m^2 s^{-3}, smaller

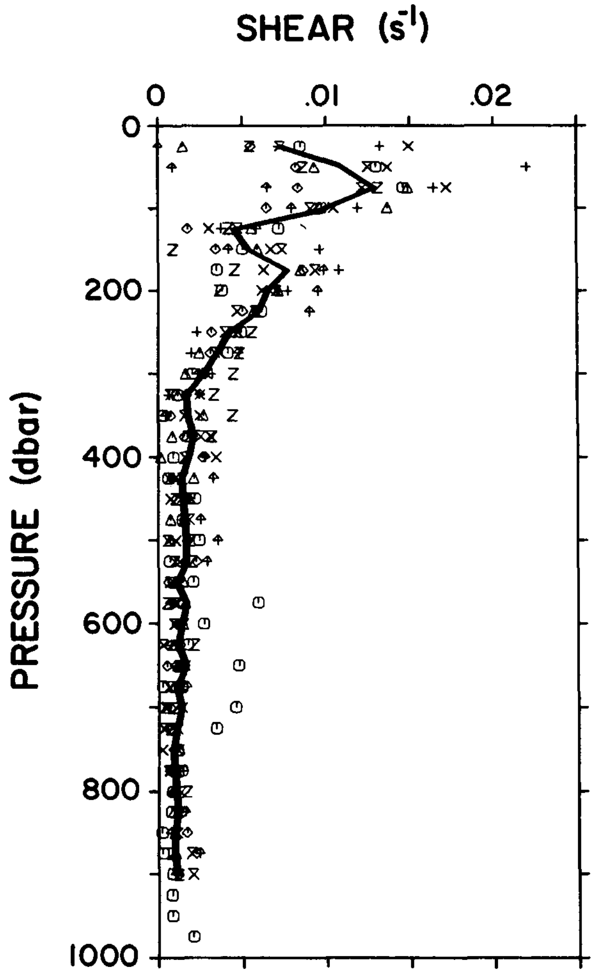

FIG. 5. Eight vertical profiles of 25 m shear estimates from White Horse velocities coincident with the profiles of Fig. 4. The thick line is the average.

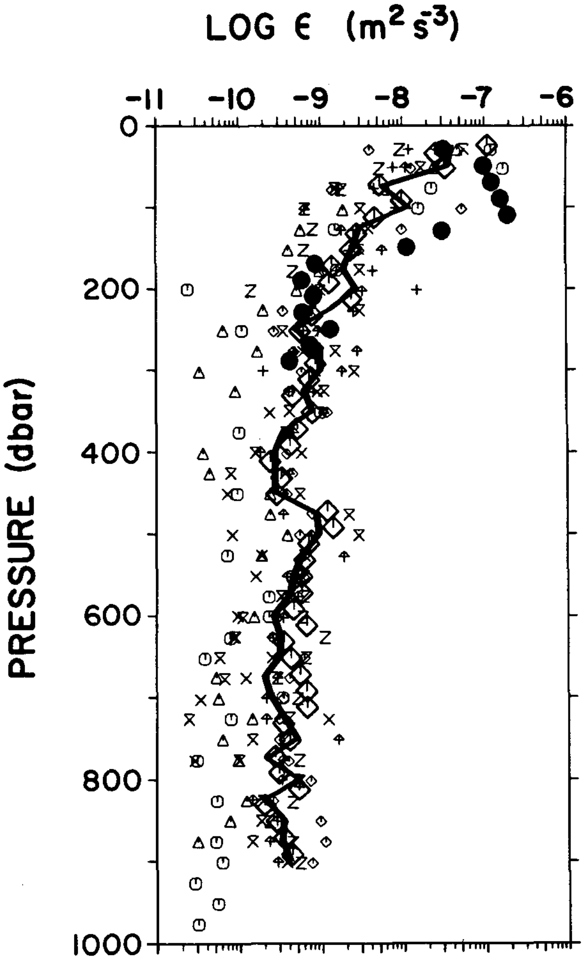

FIG. 6. Vertical profiles of ϵ averaged over 25 m depth, which are nearly synoptic with the data of Figs. 4 and 5. The thick line is the average over the eight profiles. Large solid dots are 20 m averages over nineteen profiles taken at 150°W, and within ½° of the equator, by Crawford in January–February 1979. The large diamonds are 20 m averages over thirteen equatorial profiles from February 1982.

averages were obtained by setting individual estimates of ϵ which were less than the noise level equal to zero when averaging. We believe, that by extending the dynamic range at low values of ϵ, we obtain a more representative distribution of ϵ averaged over small intervals (<100 m). Averages over intervals larger than 100 m are consistently dominated by comparatively thin, but highly energetic patches, and are changed only slightly by setting noise values equal to zero prior to averaging (see Moum and Lueck, 1985, for a discussion of the instrumental noise and Lueck et al., 1983, for a discussion of the merits of the practice of setting noise values to zero.)

The averaged equatorial profile of ϵ shows a near-surface maximum and decreases steadily below, in a manner not well-related to the current structure. Most interesting was the 500 m peak in ϵ, noted earlier as a persistent feature not dominated by a single profile. The depth of the peak in ϵ was well below a local subsurface maximum in N, at about 350 m. Inclusion of the five Camel profiles within ½° of the equator and not coincident with White Horse profiles into the averages (and reaveraged over 20 m intervals), indicates no significant deviation of this pattern, but actually shows enhancement of the 500 m peak.

Between 300 and 1000 m, only 10% of the 2-m averaged values of ϵ were greater than three times the noise level (or 10^{-9} m^2 s^{-3}). This is significantly less than the 22% found in the same depth range from our recent Western Pacific measurements between 21° and 42°N along 152°E (Moum and Osborn, 1986). Above 300 m, 49% of the PEQUOD ϵ values were greater than three times the noise level, compared to only 11% of the Western Pacific values. We have interpreted the more frequently-occurring microstructure patches below 300 m in the Western Pacific to be associated with enhanced internal wave activity in the main thermocline there.

Crawford (1982) reports similar equatorial microstructure measurements at 150°W, made in January and February 1979. We have averaged these estimates of ϵ from Crawford and Osborn (1981b), for comparison with our observations, and plotted these in Fig. 6. Below 160 m, the estimates from 1979 and 1982 are comparable. Above the EUC core (as indicated by the minimum shear in Fig. 5), the 1982 estimates are smaller by a factor of ten at 70, 90 and 110 m. An indicator of the relative differences between the two datasets is given by the 20 to 140 m depth averaged dissipations. Thirteen equatorial profiles from 1982 indicate a depth averaged value of 2.8×10^{-8} m^2 s^{-3}, compared to 12×10^{-8} m^2 s^{-3} from nineteen 1979 equatorial profiles. It is important to consider the influence of Crawford's profiles 11 to 17 (Fig. 6 from Crawford, 1982), which were made within a seven hour time span. Removing these values from the 20 to 140 m depth average reduces the value to 6.2×10^{-8} m^2 s^{-3}, or a factor of 2 larger than the 1982 values, rather than a factor of four. Although Crawford (1982) finds no strong correlation between winds and ϵ over the complete observation period, it is noteworthy that the 1979 profiles 11 to 17 were made near the end of a prolonged period of high winds, following which both the winds and ϵ decreased substantially. Moum and Caldwell (1985), with a considerably more extensive dataset, have found strong evidence for wind-influenced mixing above 100 m depth at the equator, although the mechanism by which this occurs is not yet clear.

c. ϵ versus latitude

A notable result from Crawford (1982) was the large peak in depth-averaged (surface to EUC core) dissipation values near (<½°) the equator from both the Pacific, in 1979, and the Atlantic, in 1974. The result is reproduced here in Fig. 7, upon which is plotted the 1982 PEQUOD results. Although somewhat larger values of averaged ϵ were found near the equator in 1982, the results cannot be considered a confirmation of the previous findings. In light of the Tropic Heat results, in fact, we expect the depth averages to be related to both the diurnal heating cycle, and to the wind and hence highly intermittent when sparsely and irregularly sampled (as was done in 1979 and 1982).

3. Discussion

A review of the literature reveals that at depths where the large scale shear due to the undercurrent is strong (0 to 300 m), levels of turbulence and finestructure activity are enhanced over those levels observed off the equator. Below these depths (in the equatorial Pacific), microstructure activity was either weakly enhanced (Toole and Hayes, 1984), little different (Gregg, 1976), or weaker than observed off the equator (our results). There is some evidence that the shear at these depths is dominated by equatorially trapped waves of periods of several days or longer, and that the internal wave energy decreases with depth at a rate more rapid than predicted by the Garret-Munk model due to critical-layer absorption (Eriksen, 1985). Hence, the source of

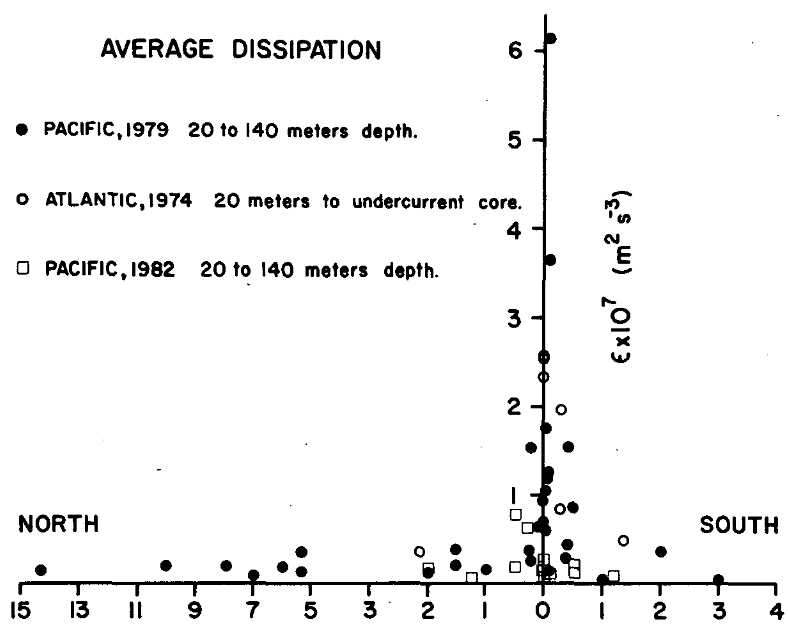

FIG. 7. Averaged dissipations observed during the *Parizeau* cruise in 1979, the *Atlantis II* cruise in 1974, and the *Thomas G. Thompson* cruise in 1982. The 1982 data were added to Fig. 3 of Crawford (1982).

internal wave energy may be weak, resulting in microstructure levels significantly less than observed off the equator.

Most results to date show enhanced small-scale activity at the equator, above the core of the undercurrent. Crawford (1982) observed a peak in dissipation at 0 to 140 m at 150°W; Gregg (1976) observed enhanced Cox numbers at 0 to 400 m at 155°W; McPhaden (1985) finds enhanced levels of a finestructure Cox number at 0 to 400 m over a period of one year at 150° to 158°W; Toole and Hayes (1984) find a significant peak at the equator of frequency of occurrence of Richardson numbers below ¼, averaged over 5 m layers, for profiles at 110°W between 150 to 250 m; Jones (1973) observed very low Richardson numbers between 0 to 400 m at 97°W. Moum et al. (1986) have more recently obtained a more intensively-sampled dataset (1 n mi spacing of ϵ profiles from 3°N to 3°S) and found no obvious peak in ϵ within the sampling domain aside from those associated with nighttime mixing. However, enhanced fine- and microstructure activity was consistently observed above the thermocline (at night) and the sampling domain may not have encompassed an off-equatorial regime.

It is more difficult to comment upon mixing below 300 m. Of the results noted above, only Toole and Hayes (1984), Gregg (1976) and the present study examined deeper waters, with conflicting results as noted above. To find a pattern, we must examine observations of finestructure and internal wave kinetic energy, which are only weak indicators of the strength of mixing. Results of vertical profiles are described by Hayes and collaborators. Toole and Hayes (1984) observed a higher frequency of $\mathrm{Ri} < ¼$ at 150 to 250 m at 110°W for equatorial profiles, but an insignificant increase below 600 m. The speed of the undercurrent at this time was faster (150 cm s^{-1}) than in previous profiles at this position in 1979. Hayes (1981) shows that the variance of vertical isotherm displacement at the equator was similar over a period of several days, but varied by a factor of 3.5 throughout the year 1979. The observations in 1979 showed an excess of meridional over zonal kinetic energy at vertical wavelengths less than 50 m. However, Toole and Hayes (1984) report an excess of zonal over meridional energy in 1982, when the undercurrent speed was higher. Hayes (1981) reports of horizontal velocity finestructure at 400 to 1600 m that, "the TOPS spectrum obtained at 4°24′N, 110°W agrees quite well with Sanford's (1975) measurements at 28°N. In contrast, the equatorial spectrum is nearly an order of magnitude higher for wavelengths longer than 100 m". Hayes (1981) notes that "for longer wavelengths (200 to 1200 stretched meters) the kinetic energy density falls as approximately k^{-2} while for shorter wavelengths (10 to 150 m) the slope is more nearly k^{-3} where k is the vertical wavenumber".

Eriksen (1985) describes current meter measurements at 500 to 3000 m at 140°W in 1982, reporting two significant results: for wave energy at periods of less than one-half day, there was an excess of meridional over zonal kinetic energy, and a decay of kinetic energy with depth according to roughly $N(z)^{1.2}$, significantly different from $N(z)^{1.0}$, predicted by the Garrett–Munk theory (Munk, 1981). After an exhaustive analysis he suspects that a possible explanation "may be that the primary source of internal wave energy is at or near the surface and that critical layers absorb waves oriented parallel to the vertical shear preferentially, thus producing a deficit of energy, an excess of meridional over zonal wave flux, and a significant coherence between temperature and zonal current at depth".

These results can explain our very low turbulence levels observed below 300 m and the concentration of high dissipations near 500 m. If high vertical wavenumber, high frequency internal wave energy diminishes with depth due to critical layer absorption, but low frequency jets of vertical wavelength 100 to 200 m are much more energetic at the equator (Eriksen, 1985), then one might expect a pattern of very low turbulence levels, except near these jets.

It is desirable to try to correlate the times and regions of strong mixing below 300 m with the strength of the undercurrent or the longitude of the observations. The indicators of strong mixing are all at or near 100°W (Toole and Hayes, 1984; Jones, 1973), while normal mixing rates (Gregg, 1976), or weaker rates (our results), are at 140° to 155°W, suggesting a strengthening of turbulence from west to east.

Acknowledgments. The captain and crew of the R/V *Thomas G. Thompson* provided invaluable assistance in the deployment of Camel III. J. Richman and C. Eriksen allowed us to join the cruise on short notice. R. Noel provided shipboard technical assistance, and S. Milaire made the airfoil probes. We are grateful to J. Luyten for permitting us access to the White Horse data. Discussions with J. Richman were very helpful. JNM wishes to acknowledge the support of the Natural Sciences and Engineering Research Council of Canada in the form of a Postgraduate Scholarship while a student at the University of British Columbia. This work was sponsored by the United States Office of Naval Research.

REFERENCES

Crawford, W. R., 1982: Pacific equatorial turbulence. *J. Phys. Oceanogr.*, **12**, 1137–1149.

——, and T. R. Osborn, 1979: Energetics of the Atlantic equatorial undercurrent. *Deep-Sea Res.*, **26**, (GATE supplement II), 309–323.

——, and ——, 1981a: Turbulence in the equatorial Pacific Ocean. Ms. Rep. 81-1, Institute of Ocean Sciences, Sidney, B.C., 63 pp.

——, and ——, 1981b: Control of equatorial ocean currents by turbulent dissipation. *Science*, **212**, 539–540.

Eriksen, C. C., 1985: Some characteristics of internal gravity waves in the equatorial Pacific. *J. Geophys. Res.*, **90**, 7243–7255.

Gregg, M. C., 1976: Temperature and salinity microstructure in the

Pacific equatorial undercurrent. *J. Geophys. Res.*, **81**, 1180–1196.
——, H. Peters, J. C. Wesson, N. S. Oakey and T. J. Shay, 1985: Intensive measurements of turbulence and shear in the Equatorial Undercurrent. *Nature*, 318.
Hayes, S. P., 1981: Vertical fine structure observations in the eastern equatorial Pacific. *J. Geophys. Res.*, **86**, 10 983–10 999.
Jones, J. H., 1973: Vertical mixing in the equatorial undercurrent. *J. Phys. Oceanogr.*, **3**, 286–296.
Luyten, J. R., G. Needell and J. Thomson, 1982: An acoustic dropsonde design, performance and evaluation. *Deep-Sea Res.*, **29**, 499–524.
Lueck, R. G., W. R. Crawford and T. R. Osborn, 1983: Turbulent dissipation over the continental shelf off Vancouver Island. *J. Phys. Oceanogr.*, **13**, 1809–1818.
McPhaden, M. J., 1985: Finestructure variability observed in CTD measurements from the central equatorial Pacific. *J. Geophys. Res.*, **90**, 11 726–11 740.
Moum, J. N., 1984: Velocity microstructure measurements in the western and central equatorial Pacific. Ph.D. dissertation, University of British Columbia, 270 pp.
——, and D. R. Caldwell, 1985: Local influences on turbulence in the equatorial ocean shear flow. *Science*, **230**, 315–316.
——, and R. G. Lueck, 1985: Causes and implications of noise in oceanic dissipation measurements. *Deep-Sea Res.*, **32**, 379–390.
——, and T. R. Osborn, 1986: Mixing in the main thermocline. *J. Phys. Oceanogr.*, **16**, 1250–1259.
Moum, J. N., D. R. Caldwell, C. A. Paulson, T. K. Chereskin, and L. A. Reqier, 1986: Does ocean turbulence peak at the equator? *J. Phys. Oceanogr.* (in press).
Munk, W. H., 1981: Internal waves and small-scale processes. *Evolution of Physical Oceanography*, B. A. Warren and C. Wunsch, Eds., The MIT Press, 264–291.
Osborn, T. R., 1974: Vertical profiling of velocity microstructure. *J. Phys. Oceanogr.*, **4**, 109–115.
——, and L. E. Bilodeau, 1980: Temperature microstructure measurements in the equatorial Atlantic. *J. Phys. Oceanogr.*, **10**, 66–82.
——, and W. R. Crawford, 1980: An airfoil probe for measuring turbulent velocity fluctuations in water. *Air-Sea Interaction*, F. Dobson, L. Hasse and R. Davis, Eds., Plenum, 369–386.
Sanford, T. B., 1975: Observations of the vertical structure of internal waves. *J. Geophys. Res.*, **80**, 3861–3871.
Toole, J. M., and S. P. Hayes, 1984: Finescale velocity-density characteristics and Richardson number statistics of the eastern equatorial Pacific. *J. Phys. Oceanogr.*, **14**, 712–726.