Wind Direction and Equilibrium Mixed Layer Depth in the Tropical Pacific Ocean

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ABSTRACT

The normal equilibrium state of the tropical Pacific mixed layer is explained by the steady state solution for the maximum vertical penetration of oceanic turbulence generated by local atmospheric forcing. The previously overlooked interaction between planetary rotation and the zonal wind stress is believed to increase the vertical turbulent kinetic energy, causing the deep mixed layer in the central and western equatorial Pacific Ocean. The unique conditions of the tropical Pacific provide a test for a revision to the basic equilibrium theory for turbulent mixing in stable oceanic planetary boundary layers.

1. Introduction

As noted by Moore and Philander (1977) and Niiler (1982), there has been no explanation for the unusually deep surface mixed layer in the central and western part of the equatorial Pacific, compared to the very shallow mixed layer in the eastern equatorial Pacific. This zonal structure can be seen in Fig. 1 from Lemasson and Piton (1968) and in similar sections reported by others (Colin *et al.*, 1971; Halpern, 1980), as well as in climatological layer depths (Fig. 2) of Robinson (1976).

The climatological zonal tilt of the upper part of the equatorial thermocline, and the anomalous dynamical response of the equatorial thermocline during the course of El Niño events are generally understood (Wyrtki, 1975; Gill, 1975). What is not clear is why the surface turbulent boundary layer is so deep toward the west in the tropical Pacific. This thick upper layer is well mixed in spite of the stratifying buoyancy forces associated with a moderately strong downward net surface heat flux. The surface buoyancy flux (Fig. 3) is computed from the heat flux and radiation fields of Weare *et al.* (1981). Because of the stabilizing effect of this downward surface buoyancy flux, the wind stresses (Figs. 4a, b) of Wyrtki and Meyers (1976), as found in the vicinity of the equator near the dateline, would normally be expected to generate a mixed layer having a depth of a few tens of meters at most. However, both density and dissipation profiles (Crawford and Osborn, 1981) indicate a turbulent boundary layer or mixed layer having a depth of 100 to 150 m in the central Pacific. This situation prevails over a wide area spanning several degrees north and south of the equator from about 150°W to about 160°E.

Pacanowski and Philander (1981) also noted the zonal structure in the thermocline. They suggested that the Kraus and Turner (1967) parameterization as applied by Hughes (1980) was inadequate to explain this zonal structure. Their objection was based on the one-dimensional limitation of the Kraus–Turner model. However, even with the inclusion of advection in a three-dimensional dynamical model, Pacanowski and Philander find that they cannot explain the zonal structure of the mixed layer and thermocline evident in Figs. 1 and 2. They tested a vertical mixing parameterization with an eddy viscosity/conductivity dependent upon Richardson number as well as a more complex turbulence closure model. Neither of these mixing parameterizations could reproduce the deep, well-mixed layers in the central and western tropical Pacific.

There are valid questions concerning the definition of "mixed layer." This is particularly true at the equator. Although temperature is well mixed to 100 m or more, momentum is not because of the undercurrent shear which is generated by the opposing forces of the westerly wind stress and the zonal equatorial pressure gradient. Away from the equatorial zone, the mixed layer currents are in fact much more vertically homogeneous (Firing *et al.*, 1981). In any case, it is the *turbulent* boundary layer that is of interest here ("mixed layer" will be used interchangeably with "turbulent

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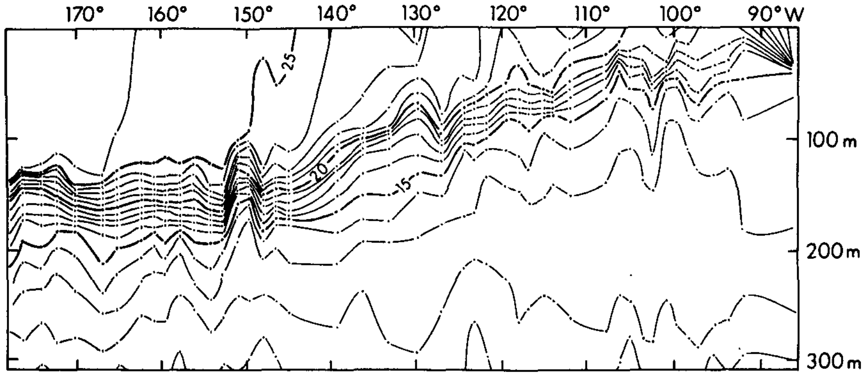

FIG. 1. Equatorial temperature section from Lemasson and Piton (1968) showing a deep mixed layer west of about 145°W, transition zone between about 145°W and 110°W, and a shallow mixed layer east of 110°W.

boundary layer"). Observations of turbulent dissipation by Crawford and Osborn (1981), and the calculations of effective vertical diffusivity (Gregg, 1976; Osborn, 1980) from microstructure profiles in the near-equatorial domain, clearly show the turbulent boundary layer to be very deep and very energetic. This region of high turbulent kinetic energy corresponds well to the thermally quasi-homogeneous layer which overlays the core of the undercurrent. Thus the depth of penetration of the turbulence appears to correspond well to the surface mixed layer of Fig. 1, with a depth in excess of 100 m near the dateline, and a depth of less than 20 m east of about 115°W. This apparent mixing depth undergoes a marked zonal transition between about 115°W and 145°W.

Meridional temperature sections from the Hawaii-to-Tahiti shuttle experiment (Wyrtki and Kilonsky, 1984) show the mixed layer depth increases somewhat both north and south of the equator, but with an asymmetrical structure. For meridional sections between 150°W and 158°W, the maximum mixed layer depth on the north side of the equator occurs at about 5°N, and the maximum on the south side is near 8°S. Most of this asymmetry is due to the countercurrent dynamic topography north of about 5°N.

Between 4°S and 4°N (a distance of about 900 km), the mixed layer temperature and current structures of the western and central Pacific are more symmetrical. This reflects the natural symmetry of the equatorial response to an approximately zonal wind stress and an approximately symmetrical climatological surface heat flux pattern. A key feature of this dynamical response is the upwelling, which is at least a partial explanation for the meridional change in mixed layer depth immediately north and south of the equator. Schopf and Cane (1983) and Muller *et al.* (1984) find that the depth of the mixed layer can be considerably reduced by even a small upward velocity at its base. Thus the layer depth is reduced in the near vicinity (±50 km) of the equator. In any case, the problem of explaining the deep tropical mixing is exacerbated since the turbulent boundary layer is even deeper in much of the central and western tropical Pacific than is indicated by Fig. 1, which is right on the equator.

Several modeling efforts are directed toward the understanding of equatorial dynamics. It is recognized, however, that such models will need to incorporate realistic mixed layers to be successful (Leetma *et al.*, 1981). Also, to be useful for studies of air–sea interactions, these models must be capable of realistic thermodynamic feedback between the ocean and the atmosphere (Philander *et al.*, 1984; Haney and Rennick, 1985). Such realism cannot be achieved unless, firstly, the basic equilibrium state of the oceanic turbulent boundary layer is understood and can be simulated.

Pacanowski and Philander (1981), Hughes (1980),

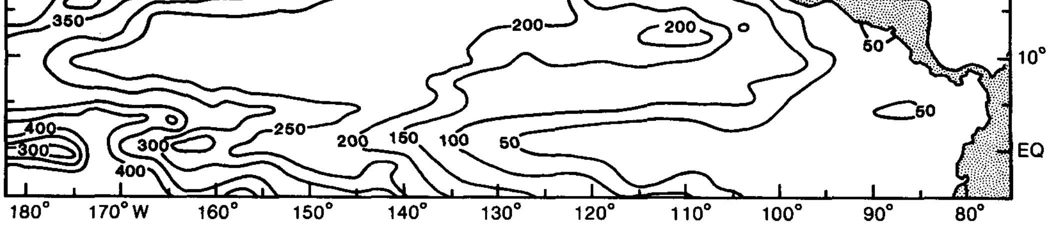

FIG. 2. Climatological depth (ft) of the top of the thermocline for month of March, from Robinson (1976).

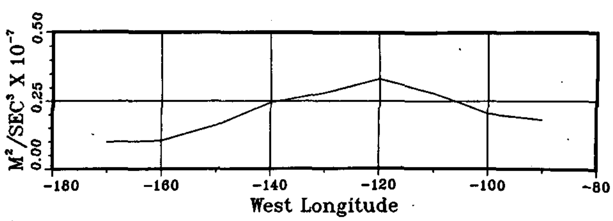

FIG. 3. Least squares, third-order polynomial fit to annual mean downward surface buoyancy flux, $B_0(x)$, computed from Weare *et al.* (1981) climatological values for surface heat flux and radiation.

and Schopf and Cane (1983) have employed turbulent mixing models in the study of equatorial dynamics. As previously discussed, Pacanowski and Philander used a mean flow dynamical instability (Richardson number) parameterization. Both Hughes (1980) and Schopf and Cane (1983) employ versions of the Kraus and Turner (1967) mixed layer model. Hughes was largely concerned with the meridional structure, and he attributed the contrasting mixed layers in the western and eastern Pacific to differences in upwelling velocity. Schopf and Cane showed that a three-dimensional equatorial model can be adapted to include mixed layer physics, which overcomes the earlier criticism of Pacanowski and Philander concerning the use of the Kraus–Turner model. Schopf and Cane, however, modeled only the eastern tropical Pacific, and their numerical simulations do not address the question of zonal variations in mixed layer depth.

2. Zonal atmospheric forcing and (lack of) zonal mixed layer response

Weare *et al.* (1981) and an updated (Wyrtki, personal communication, 1984) version of Wyrtki and Meyers (1976) are used as best estimates of the annual mean surface heat flux, radiation and wind stress for the near-equatorial Pacific Ocean between 90°W and the dateline. Figure 3 shows the resulting longitudinal variation of annual downward buoyancy flux (B_0), assuming that 50% of the incident shortwave radiation is absorbed in the top meter, and the remainder is absorbed exponentially with an extinction coefficient of $(6\ \mathrm{m})^{-1}$. Figures 4a and 4b show the eastward (τ_x) and northward (τ_y) wind stress components. Weare *et al.* (1981) indicate a possible absolute error of $\pm30\ \mathrm{W\ m^{-2}}$ for the net surface heat flux Q_0, which precludes putting too much faith in the smaller scale horizontal variability in their net heat flux calculations. However, the zonal equatorial net surface heat flux gradient, $\partial Q_0/\partial x$, was found to be statistically significant. Also, the zonal average of Q_0 near the equator is consistent with independent heat budget analyses that show the region between the Galapagos Islands and New Guinea to be an area of net heat gain (Wyrtki, 1981; Hastenrath, 1980). Although uncertainty in the drag coefficient may cause error in the annual average of the wind stress on the order of $\pm20\%$, the zonal dependence of the direction of the wind stress is expected to be accurately represented.

An additional uncertainty in applying these boundary conditions to an ocean model is the neglect of seasonal and shorter time scale changes. Seasonal fluctuations in the central and western equatorial Pacific are relatively small, but there are significant seasonal fluctuations in wind stress and other atmospheric parameters in the eastern tropical Pacific. Elsberry and Camp (1978) showed that atmospheric forcing of a synoptic period can influence significantly the longer-term oceanic mixing response. However, the primary intent of this study is to explore possible causes of the mean zonal tendencies, and the depth of the central and western equatorial mixed layer in particular. Therefore, the present exclusion of the unsteady components of the surface boundary conditions does not seem unreasonable. Nevertheless, the basic question addressed here must be a qualified one: "What major features of the near-equatorial mixed layer can be explained by the local annual mean surface forcing?"

Without horizontal or vertical advective effects, a surface-heated mixed layer would normally be expected to have an asymptotic maximum depth comparable to the Obukhov length scale,

$$L = \frac{2u_*^3}{B_0}, \tag{2.1}$$

where B_0 is the net downward surface buoyancy flux, and $\underset{\sim}{\tau} = \rho u_*^2$ is the magnitude of the surface wind stress.

Figure 5 shows the longitudinal dependence of the Obukhov scale $L(x)$, determined from the computed values of B_0 (Fig. 3) and the Wyrtki and Meyers values for $\underset{\sim}{\tau}$ (Figs. 4a, b). Comparisons of Fig. 5 with Figs. 1 and 2 show that L is similar to the mixed layer depth in the east, but it is much too shallow farther west. Although there is a zonal variation in L, it is not nearly large enough. Including the effects of equatorial up-

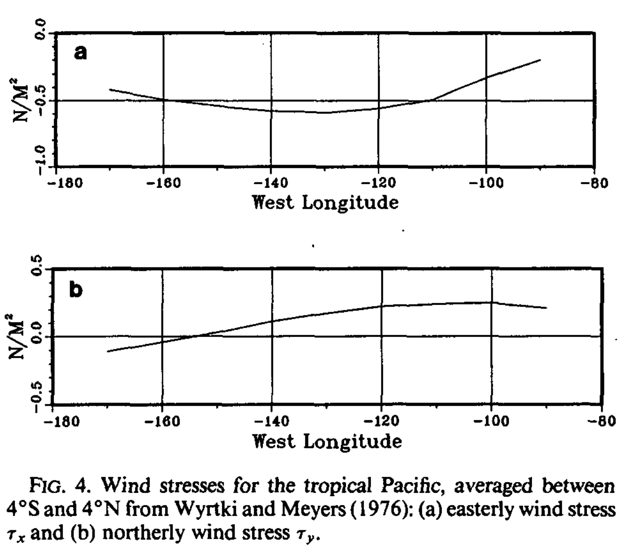

FIG. 4. Wind stresses for the tropical Pacific, averaged between 4°S and 4°N from Wyrtki and Meyers (1976): (a) easterly wind stress τ_x and (b) northerly wind stress τ_y.

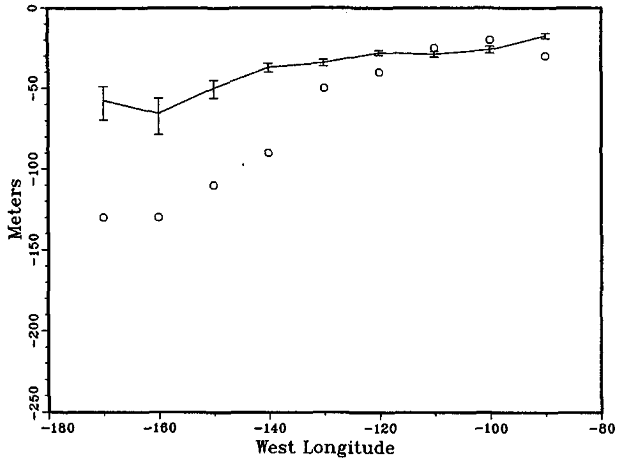

FIG. 5. The Obukhov length scale dependence upon longitude at the equator computed using $Q_0(x)$ and $\tau(x)$ illustrated in Figs. 3 and 4, respectively. Vertical bars show the effect of varying Q_0 by ± 3 W m^{-2}. Circles approximate the depth of the mixed layer in Fig. 1.

welling would reduce the predicted equilibrium h to a value even less than would be computed from (2.1). Thus the Kraus–Turner model, which has (2.1) as the equilibrium state, appears to approximate the mixed layer depth in the eastern tropical Pacific, but it underestimates the mixed layer depth over the remainder of the tropical Pacific. This is consistent with the results of Pacanowski and Philander (1981), who employed a different parameterization for vertical mixing.

There are several other factors that may influence the equilibrium mixed layer depth: (i) advection, (ii) variability in the absorption of radiation below the surface, (iii) effect of salinity, (iv) turbulence generated by shear not directly related to the local wind stress, and (v) the effect of a neglected physical process. A possible candidate process in this latter category is the effect of the usually neglected component of the Coriolis force upon turbulent mixing. The basic theory for this mechanism is contained in the first part of this study, Garwood *et al.*, 1985. We shall refer to this latter process as the rotation–stress interaction. This process has been shown to increase vertical mixing by augmenting the vertical component of the turbulent kinetic energy for situations with westward surface wind stress (easterly winds).

As mentioned earlier, upward vertical advection can only reduce the depth of the turbulent boundary layer. As Hughes (1980) postulated, horizontal variability in vertical motion can cause horizontal variability in mixed layer depth. However, positive vertical velocity at the equator is not likely to accentuate the east–west contrast in $L(x)$ (Fig. 5). The value of L is already small in the east, and upwelling will make the mixed layer even shallower in the west. Even downwelling will not increase the mixed layer depth for surface-heated boundary layers because the time scale of mean advection is much greater than the dissipation-controlled time scale of the turbulence (Muller *et al.*, 1984). Thus turbulent kinetic energy can not be advected below the equilibrium depth for nonadvective cases having the same surface boundary conditions. Likewise, horizontal advection can only decrease mixed layer depth, as in the case of an advected shallow plume of buoyant water (Garvine, 1984). When denser water is advected toward less dense water in the presence of a downward surface heat flux, the tendency is for the denser water to slide beneath the more buoyant turbulent surface layer without interacting with the turbulence (Adamec and Garwood, 1984). The energetics of the turbulence for this latter process is not well understood, however, and additional observations are required to confirm this tendency.

From Jerlov (1976), central equatorial Pacific water has an optical type of Ia or II. For such types of oceanic water, penetration of shortwave radiation below the surface may reduce the net downward surface buoyancy flux in comparison with the same radiation incident at the surface of turbid water. This effect has been included in the calculation of B_0, but the value of B_0 may be even further reduced if the effective extinction coefficient is much greater than the estimated value of $(6\ \mathrm{m})^{-1}$. This would increase L, according to (2.1). The effect of penetrating solar radiation is minor if the mixed layer is much deeper than the depth scale of the penetration. However, for very shallow mixed layers, the Obukhov length scale and the resulting depth of mixing may be changed significantly (Simpson and Dickey, 1981). This effect would only be important here for the shallow mixed layer of the eastern region, and it could cause the layer depth there to be greater by approximately 5–10 m. The buoyancy flux in the deeper mixed layer to the west will not be affected appreciably by this penetration of shortwave radiation.

A comparable reduction in effective surface buoyancy flux is caused by evaporation which usually exceeds precipitation east of the dateline in the equatorial Pacific. Inclusion of the negative salinity flux would reduce somewhat the calculated net downward buoyancy flux. This in turn can increase somewhat the depth of mixing (Miller, 1976).

Turbulence may be generated by shear not directly related to the local wind stress via a dynamic instability. Although Pacanowski and Philander (1981) cannot explain the observed $h(x)$ by such a dynamic instability associated with the undercurrent shear, we believe that this shear is important for the turbulent kinetic energy balance right at the equator. The large dissipation rate that is found at depth near the core of the undercurrent is probably attributable to shear production within the undercurrent (Crawford and Osborn, 1981). Here, however, we are concerned with the larger tropical region, most of which is well outside the direct influence of the undercurrent. Gradient Richardson number calculations based upon the profiles of Firing *et al.* (1981) show no evidence of systematic dynamical instabilities away from the immediate vicinity of the undercurrent. Thus, for the general tropical Pacific, none of the above factors by itself appears to explain the depth of the mixed layer in the central and western tropical Pacific. Hence, the possibility of an alternative physical process is considered.

3. The effect of wind direction on the equilibrium depth of mixing

Garwood *et al.* (1985) suggest a modification to the equilibrium mixing depth,

$$h = \frac{C_1 L}{1 + C_2 \Phi} \tag{3.1}$$

where L is the Obukhov length scale as in (2.1), and $C_1 \sim 1$, and $C_2 \sim 12/7$. The function

$$\Phi = \frac{\Omega_y \tau_x}{\rho B_0} \tag{3.2}$$

depends upon the zonal wind stress (τ_x), the northerly component of rotation (Ω_y), and the downward surface buoyancy flux (B_0). For the trade winds, $\tau_x < 0$, and the sign of Φ is negative. Thus the denominator in (3.1) is less than unity, which may increase considerably the equilibrium depth h.

For the equilibrium turbulent kinetic energy budget (Fig. 6), the effect of $\Omega_y \tau_x$ is to convert horizontal turbulent kinetic energy to vertical turbulent kinetic energy when τ_x is negative, and vice versa when τ_x is positive. With x eastward and y northward, the horizontal and vertical turbulent kinetic energy budgets, vertically integrated over the mixed layer are:

$$\frac{\partial}{\partial t} \int_{-h}^{0} \frac{\overline{(u'^2 + v'^2)}}{2} dz = 0 = G - \Pi_{zz} + \frac{\Omega_y h \tau_x}{\rho} - \frac{D}{3}, \tag{3.3}$$

$$\frac{\partial}{\partial t} \int_{-h}^{0} \frac{\overline{(w'^2)}}{2} dz = 0 = -\frac{B_0 h}{2} + \Pi_{zz} - \frac{\Omega_y h \tau_x}{\rho} - \frac{D}{3}. \tag{3.4}$$

The shear production is G; Π_{zz} is the pressure–rate of strain, which redistributes turbulent kinetic energy from the horizontal to the vertical; D is the viscous dissipation rate. These equations are derived assuming steady-state, zero entrainment, local isotropy (dissipation partitioned equally among the components), linear flux profiles, and one dimensionality for the turbulent kinetic energy budget (a good approximation even if the density and momentum fields are not horizontally homogeneous).

Although $\Omega_y \tau_x$ is not present in the total turbulent kinetic energy equation,

$$\frac{\partial}{\partial t} \int_{-h}^{0} \frac{\overline{(u'^2 + v'^2 + w'^2)}}{2} dz = 0 = G - \frac{B_0 h}{2} - D, \tag{3.5}$$

it will influence the total turbulent kinetic energy by modulating the partitioning of energy between the turbulent kinetic energy and the potential energy. From the standpoint of the total mechanical energy budget (Fig. 6), an east wind ($\tau_x < 0$) will cause increased vertical mixing of buoyancy, and increase the system potential energy at the expense of dissipation. Thus the equilibrium state of the mixed layer is a function of wind direction. The reader is referred to Garwood *et al.* (1985) for the derivation and a more detailed explanation of (3.1–4).

Using (3.1), a revised $h(x)$ (Fig. 8) is calculated from the computed $\Phi(x)$ (Fig. 7) and the previous $L(x)$ (Fig. 5). To illustrate the sensitivity of the function (3.1) to variability in the surface buoyancy flux, the prescribed net surface heat flux, $Q_0(x)$, was altered by ± 3 W m^{-2} for all values of x, and the vertical bars in Figs. 5 and 8 represent the resulting fluctuations in L and h, respectively. These bars are not to be construed as error estimates. Rather, their purpose is to show that seemingly small fluctuations in the surface forcing may cause important changes in the equilibrium mixed layer depth when Φ is strongly negative, as it is in the central and western equatorial Pacific. This sensitivity to minor changes in the surface fluxes of both buoyancy and momentum suggests that the western Pacific equilibrium state may be rather delicate. This characteristic may be important for large-scale air–sea interactions. Notice that these perturbations in surface forcing have a much reduced effect in the eastern region, where B_0 is large and Φ is small. In the east, Φ is negligible because the winds are primarily meridional, and h converges to L. Garwood *et al.* (1985) suggest a theoretical limit of $\phi = -1/2$ for the equilibrium solution (3.1) to apply. If $\Phi < -1/2$, then no equilibrium is possible, and the mixed layer will continually entrain. This would result in continual deepening, unless countered by advection. Figure 7 suggests that the central equatorial Pacific may be close to this limit.

Considering the possible errors in computing the boundary conditions and in the neglect of advection and seasonality, the general agreement between this new theoretical limit to vertical mixing and the mixed layer depth that one would objectively analyze from a section such as Fig. 1 is rather remarkable. It is important to recognize that it is insufficient to simply adjust the proportionality coefficient between h and L to achieve a comparable fit. As Pacanowski and Philander (1981) found, the alternative mixing models cannot be tuned to give a deeper equatorial mixed layer without causing excessive mixing at other locations. The zonal distribution of the calculated $h(x)$ (Fig. 8) is particularly sensitive to the zonal variation in the wind direction, and this factor does not influence $L(x)$, which is the key parameter for both Hughes (1980) and Schopf and Cane (1983).

4. Conclusions and remarks

A limiting depth of mixing is computed for the near-equatorial tropical Pacific which is dependent upon

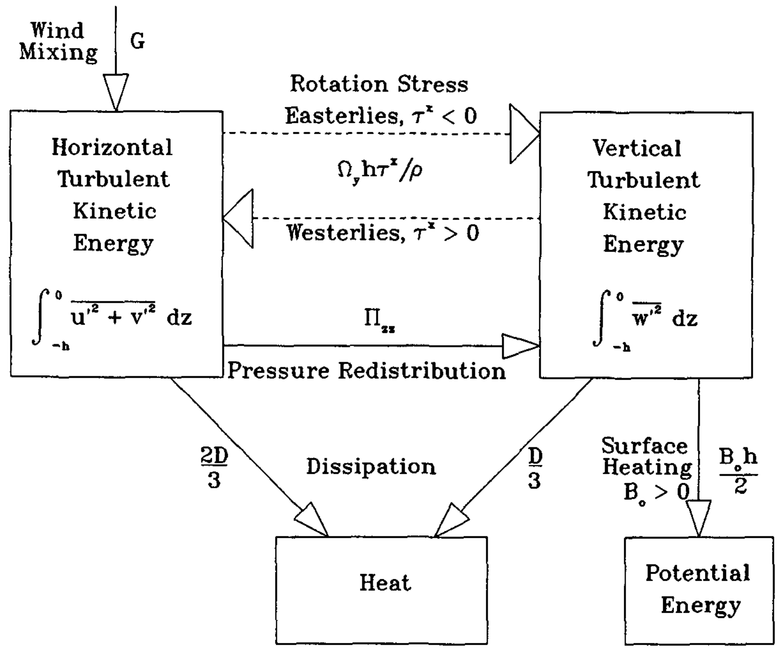

FIG. 6. Schematic of the equilibrium turbulent kinetic energy budget.

both the classical Obukhov length scale and a parameter which depends upon the magnitude and direction of the wind stress. This new limit compares much more favorably with observations, and it seems to be a plausible explanation for the very deep mixed layer in much of the tropical Pacific.

Principal limitations to this application of the theory are the accurate specification of a steady surface forcing, the neglect of advection of mixed layer depth, and the assumption that the depth of the climatological near-equatorial mixed layer should be comparable to this theoretical limit. Right on the equator (±50 km), the turbulent kinetic energy budget is almost certainly influenced by both the upwelling and the additional shear production associated with the undercurrent. Eq. (3.1) does not include the effects of upwelling and entrainment shear production that must exist if there is upwelling at the equator. However, this narrow zone is as likely to be influenced by the rotation–stress mechanism as is the wider surrounding tropical region, and (3.1) is an asymptotic limit when upwelling is reduced. Unless the estimates of the surface forcing (buoyancy flux and wind stress) are grossly in error, these limitations do not appear to detract from the comparison between the computed zonal structure of the vertical limit to mixing and observations of the tropical Pacific mixed layer.

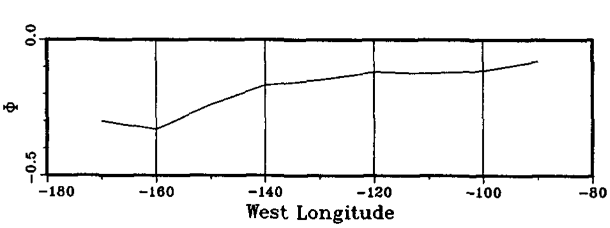

FIG. 7. The dimensionless rotational-stress parameter, $\Phi(x)$. Large negative values of Φ approaching -0.5 indicate extra vertical mixing due to the conversion of horizontal to vertical turbulent kinetic energy by the interaction of east-west wind stress and the north component of planetary rotation.

Future equatorial investigations will include not only the roles of advection, but nonstationarity, and interaction with the atmosphere as well. The sensitivity to small variations in surface heat flux shows that the equilibrium state of the central and western tropical Pacific mixed layer may be a delicate one. Small fluctuations in the surface forcing may cause much greater changes in mixed layer depth, and hence surface temperature, then would be expected in the eastern tropical Pacific, or at higher latitudes where the winds are not predominantly from the east. An anomalous reduction in mixed layer depth may increase considerably the local oceanic transfer of heat to the atmosphere by reducing the local capacity of the ocean to store heat generated by solar radiation.

This study represents a part of a larger investigation of the role of the rotation–stress mechanism upon mixed layers in general. Because the magnitude of Ω_y is greatest at the equator, it was logical to consider the possible effects at low latitude at an early stage of the investigation. The equatorial Pacific was chosen as an initial test location because of the variation in wind direction which would provide a suitably complete test. Although more extensive verification is in order, we consider this first test of the possible significance of the

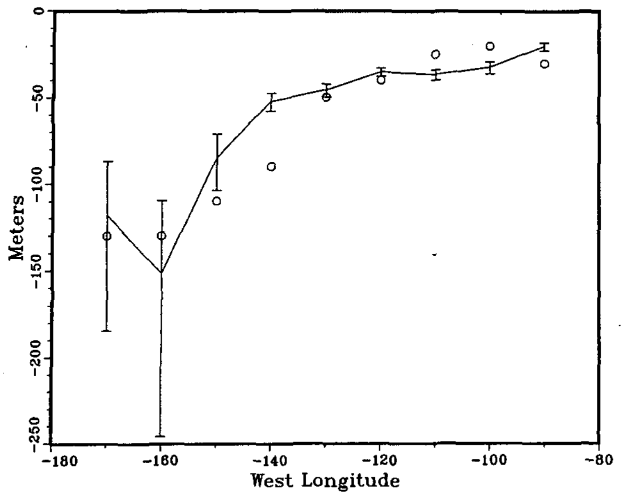

FIG. 8. The computed equilibrium mixed layer depth using Eq. (3.1), showing the importance of the rotation–stress mechanism in maintaining a deep mixed layer in the central and western Pacific. The vertical bars are for the variations in h due to fluctuations in Q_0 of ±3 W m^{-2}. Circles approximate the depth of the mixed layer in Fig. 1.

rotation–stress term for vertical mixing in the upper ocean turbulent boundary layer to be positive.

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REFERENCES

Adamec, D., and R. W. Garwood, Jr., 1985: The simulated response of an upper ocean density front to local atmospheric forcing. *J. Geophys. Res.*, **90**, 917–928.

Colin, C., C. Henin, P. Hisard and C. Oudet, 1971: Le Courant de Cromwell dans le Pacifique central en Fevrier 1970. *Cah. ORSTOM*, **19**, 167–186.

Crawford, W. R., and T. R. Osborn: Control of equatorial ocean currents by turbulent dissipation. *Science*, **212**, 439–540.

Elsberry, R. L., and N. T. Camp, 1978: Oceanic thermal response to strong atmospheric forcing. 1, Characteristics of forcing events. *J. Phys. Oceanogr.*, **8**, 206–214.

Firing, E., C. Fenander and J. Miller, 1981: Profiling current meter measurements from the NORPAX Hawaii to Tahiti Shuttle Experiment. Hawaii Institute of Geophysics Data Rep. 39, HIG-81-2, 146 pp.

Garvine, R. W., 1984: Radial spreading of buoyant, surface plumes in coastal waters. *J. Geophys. Res.*, **89**, 1989–1996.

Garwood, R. W., Jr., P. C. Gallacher and P. Muller, 1985: Wind direction and equilibrium mixed layer depth: General theory. *J. Phys. Oceanogr.*, **15**, 1325–1331.

Gill, A. E., 1975: Models of equatorial currents. *Numerical Models of Ocean Circulation*, Natl. Acad. Sci., 181–203.

Gregg, M. C., 1976: Temperature and salinity microstructure in the Pacific equatorial undercurrent. *J. Geophys. Res.*, **81**, 1180–1196.

Halpern, D., 1980: A Pacific equatorial temperature section from 172°E to 110°W during winter and spring 1979. *Deep Sea Res.*, **27A**, 931–940.

Haney, R. L., and M. A. Rennick, 1985: Stable and unstable air-sea interactions in the equatorial region. *J. Atmos. Sci.*, (submitted).

Hastenrath, S., 1980: Heat budget of tropical ocean and atmosphere. *J. Phys. Oceanogr.*, **10**, 159–170.

Hughes, R. L., 1980: On the equatorial mixed layer. *Deep-Sea Res.*, **27A**, 1067–1078.

Jerlov, N. G., 1976: *Marine Optics.* Elsevier, 231 pp.

Kraus, E. B., and J. S. Turner, 1967: A one-dimensional model of the seasonal thermocline: II. The general theory and its consequences. *Tellus*, **19**, 98–106.

Leetma, A., J. P. McCreary, Jr. and D. W. Moore, 1981: Equatorial currents: Observations and theory. *Evolution of Physical Oceanography.* The MIT Press, 184–196.

Lemasson, L., and B. Piton, 1968: Anomalie dynamique de la surface de la mer le long de l'equateur dans l'ocean Pacifique. *Cah. ORSTOM., Ser. Oceanogr.*, **6**, 39–45.

Miller, J. R., 1976: The salinity effect in a mixed layer ocean model. *J. Phys. Oceanogr.*, **6**, 29–35.

Moore, D. W., and S. G. H. Philander, 1977: Modeling of the tropical oceanic circulation. *The Sea: Ideas and Observations on Progress in the Study of the Seas*, E. D. Goldberg, Ed., 319–361.

Muller, P., R. W. Garwood, Jr. and J. P. Garner, 1984: Effect of vertical advection on the dynamics of the oceanic surface mixed layer. *Ann. Geophys.*, in press.

Niiler, P. P., 1982: Tropic Heat—A study of the tropical Pacific upper ocean heat, mass and momentum budgets. Oregon State University, 30 pp.

Osborn, T. R., 1980: Estimates of the local rate of vertical diffusion from dissipation measurements. *J. Phys. Oceanogr.*, **10**, 83–89.

Pacanowski, R. C., and S. G. H. Philander, 1981: Parameterization of vertical mixing in numerical models of tropical oceans. *J. Phys. Oceanogr.*, **11**, 1443–1451.

Philander, S. G. H., T. Yamagata and R. C. Pacanowski, 1984: Unstable air-sea interactions in the tropics. *J. Atmos. Sci.*, **41**, 604–613.

Robinson, M. K., 1976: *Atlas of North Pacific Ocean Monthly Mean Temperatures and Mean Salinities of the Surface Layer.* Naval Oceanographic Office, 173 pp.

Schopf, P. S., and M. A. Cane, 1983: On equatorial dynamics, mixed layer physics and sea surface temperature. *J. Phys. Oceanogr.*, **13**, 917–935.

Simpson, J. J., and T. D. Dickey, 1981: The relationship between downward irradiance and upper ocean structure. *J. Phys. Oceanogr.*, **11**, 309–323.

Weare, B. C., P. T. Strub, and M. D. Samuel, 1981: Annual mean surface heat fluxes in the tropical Pacific Ocean. *J. Phys. Oceanogr.*, **11**, 705–717.

Wyrtki, K., 1975: El Niño—the dynamic response of the equatorial Pacific Ocean to atmospheric forcing. *J. Phys. Oceanogr.*, **5**, 572–584.

——, 1981: An estimate of equatorial upwelling in the Pacific. *J. Phys. Oceanogr.*, **11**, 1205–1214.

——, and G. Meyers, 1976: The trade wind field over the Pacific Ocean. *J. Appl. Meteor.*, **15**, 698–704.

——, and B. Kilonsky, 1984: Mean water and current structure during the Hawaii-to-Tahiti shuttle experiment. *J. Phys. Oceanogr.*, **14**, 242–254.