THE MIXED LAYER OF THE WESTERN EQUATORIAL PACIFIC OCEAN

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ABSTRACT

The mixed layer of the western equatorial Pacific and its dynamics are poorly known because of a general lack of data. The recent Western Equatorial Pacific Ocean Circulation Study (WEPOCS) conducted two expeditions to the near-equatorial region north of Papua New Guinea, performing high-resolution CTD profiling and other hydrographic observations.

The WEPOCS CTD profiles are analyzed for various measures of the upper layer or mixed layer thickness, using criteria which depend on vertical gradients of temperature, salinity, and density. From 243 profiles, the average mixed layer depth in the western equatorial Pacific during the two WEPOCS cruises was 29 m, which is about a factor of three shallower than had previously been thought. The depth of the top of the thermocline was found to be 51 m, so there is a nearly-isothermal layer which is deeper than the mixed layer. This discrepancy is attributable to salinity stratification. It is hypothesized that the waters in this layer between the bottom of the mixed layer and the top of the thermocline are formed to the east of the WEPOCS region, and subducted below the shallow and lighter mixed layer waters found in the west.

There was a significant, but weak, dependence of mixed layer depth on wind speed observed during the CTD stations. The scatter is partly due to the fact that spot wind measurements are not always representative of

the wind history that has determined the mixed layer depth, but another important factor is the strong stabilizing buoyancy forcing associated with heavy precipitation in the this region. Under light wind conditions, there was a tendency for warm and thin layers to form at the sea surface as a result of diurnal heating, however there did not appear to be any nighttime maximum to the mixed layer depth associated with convective overturn due to cooling. This contrast with the central Pacific is most likely due to the influence of salinity on the thermodynamics of the mixed layer.

A strong westerly wind burst was observed during WEPOCS II, and the mixed layer nearly doubled in depth, and cooled by more than 1°C. Evidence of downwelling near the equator, and upwelling off the equator, was seen in the distribution of temperature, salinity, and density in the meridional section along 143°E which was occupied immediately following the wind event. This event was apparently strong enough to erode through the salinity-stratified layer and into the thermocline, resulting in the observed cooling.

The results of this study suggest that, except during strong wind events, entrainment cooling may not be an important component of the heat budget of the western Pacific warm pool. Thus, a possible mechanism for interannual warming (and subsequent cooling) of the western Pacific has been identified as the intermittent wind forcing, and the switching on and off of entrainment cooling.

INTRODUCTION

The western equatorial Pacific has been identified as a region where the westerly wind anomalies associated with the onset of the El Nino/Southern Oscillation (ENSO) phenomenon first develop (Barnett, 1977). Our oceanographic database from this important region is relatively lean, and much of what we know about the hydrographic structure of this region is from discrete sampling with relatively poor vertical resolution, or from continuous temperature profiling using bathythermographs. As a direct result, many of our ideas about the structure and dynamics of the western equatorial Pacific are inaccurate and need to be improved. An excellent example of this problem is the almost universally accepted belief that the mixed layer in the western equatorial Pacific is very deep (100 m or more).

This notion has its roots in such observations as the trans-Pacific equatorial temperature section observed by Lemasson and Piton (1968) between 20 November 1964 and 8 March 1965 (Fig. 1, which is actually taken from Colin et al., 1971). In this often-cited section, the thermocline slopes upward from the central Pacific, intersecting the surface in the east. However, the thermocline is essentially flat between the central and western Pacific. The top of the thermocline is found at about 100 m, and from this section it appears as if the bottom of the mixed layer is found at the top of the thermocline. It should be noted that this section was occupied in two pieces, with a 2-month-long break near 150°W (Cremoux, 1981). Also, the 1965 ENSO event was beginning during these observations, and they may well be anomalous.

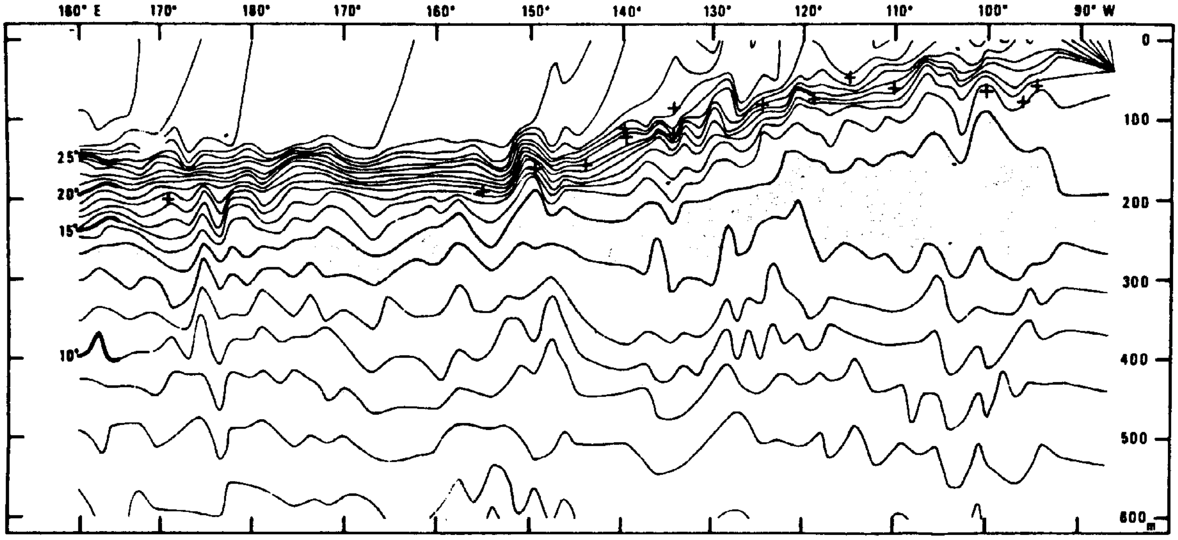

Figure 1. Thermal structure along the equator in the Pacific Ocean observed by Lemasson and Piton (1968) [After Colin et al., 1971].

An example of a deep mixed layer in the western equatorial Pacific is given in Fig. 2 which shows the temperature, salinity, and density observed at a near-equatorial station during the second cruise of the Western Equatorial Pacific Ocean Circulation Study (WEPOCS). All three properties are well-mixed down to a little more than 100 m. Figures 1 and 2 are consistent.

A very different situation is illustrated in Fig. 3, which shows a CTD profile from another near-equatorial station occupied during WEPOCS I. Here, the nearly isothermal layer is about 80 m deep. However, salinity is only well-mixed to 35 m, and thus the density profile has a sharp gradient near 35 m. A surprising result is that only 15-20% of the stations during WEPOCS I and II showed profiles like Fig. 2, and that, on average, the mixed layer in the western equatorial Pacific is quite shallow. In fact, the station shown in Fig. 2 was occupied immediately following a period of very strong winds associated with a westerly burst (cf. Luther et al. [1983] for a description of this phenomenon). This observation, as will be shown, has important implications for the ENSO phenomenon.

In the following, the WEPOCS observations will be presented in more detail, and the results of simple statistical analyses are discussed.

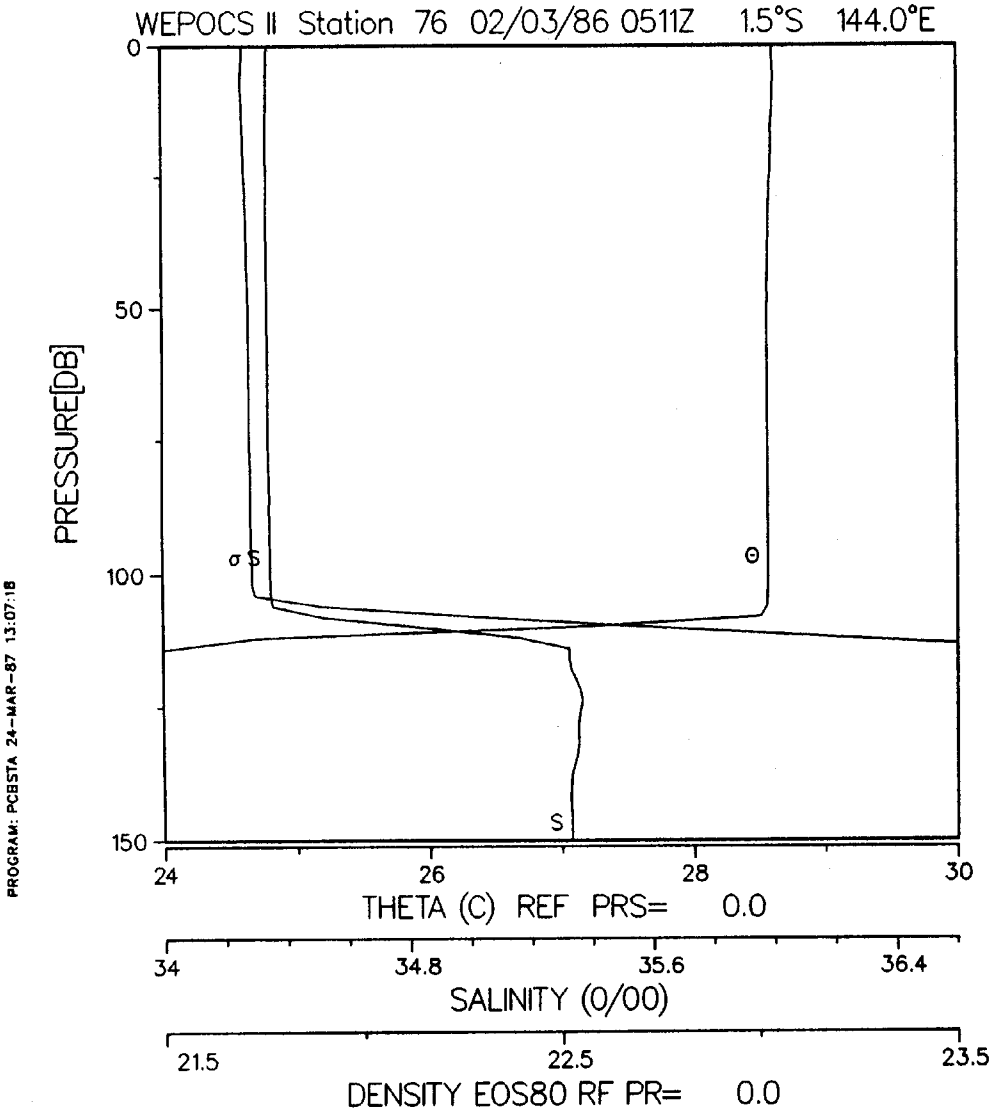

Figure 2. Potential temperature, salinity, and potential density from a CTD profile at 1.5°S, 144°E in February 1986.

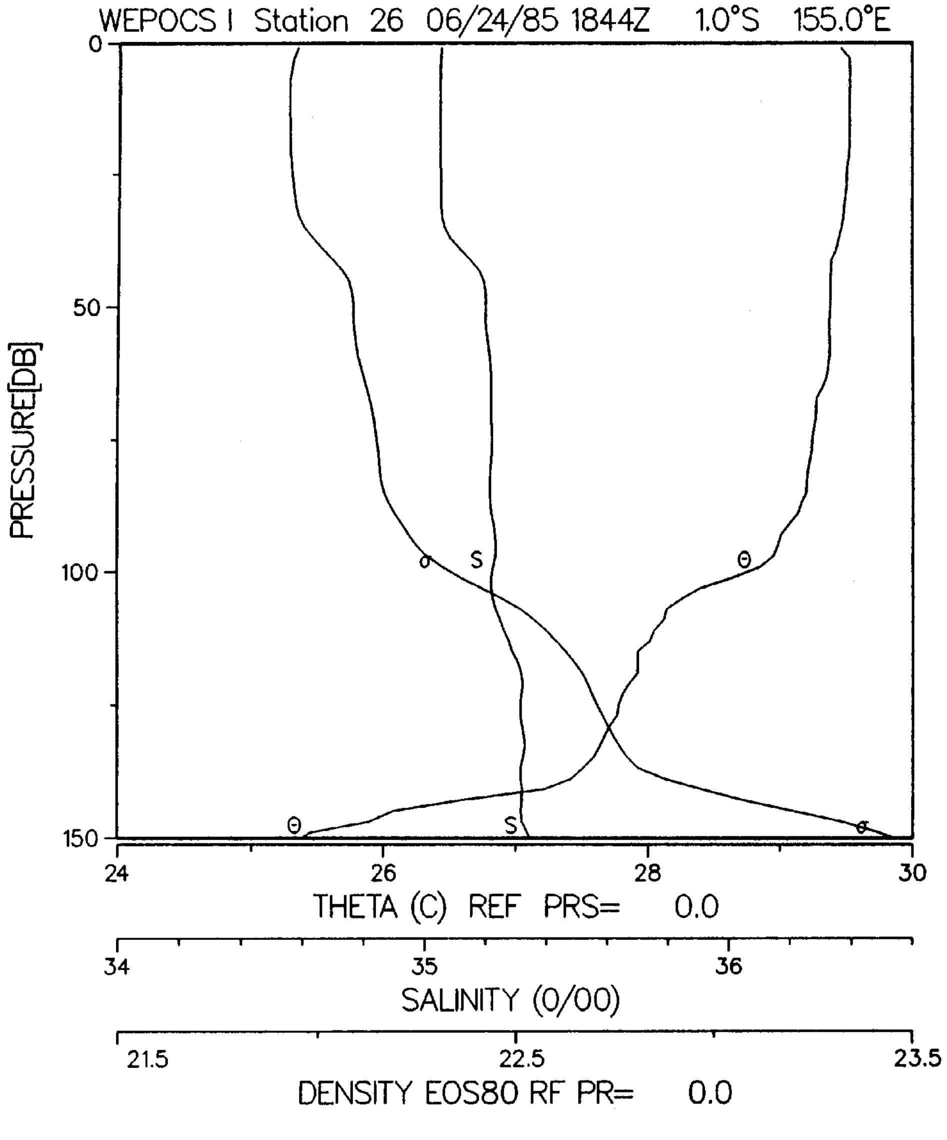

Figure 3. As in Fig. 2, except at 1°S, 155°E in June 1985.

The role of strong precipitation in the western equatorial Pacific and its influence on mixed layer dynamics is elaborated upon, and a hypothesis is outlined to explain the discrepancy between the nearly isothermal layer depth and the mixed layer depth.

DATA AND METHODS

WEPOCS

WEPOCS is a joint U.S.-Australian program to test a hypothesis concerning the source waters of the Equatorial Undercurrent, to improve our knowledge of the deep and intermediate circulation of the near-equatorial region north of Papua New Guinea, and to observe the oceanic response to the onset of the Northwest Monsoon (Lindstrom et al., 1987). Two expeditions were made (with U.S. and Australian vessels participating in each) at the two extremes of the seasonal cycle, the Southeast Trade season (June-August, 1985) and the Northwest Monsoon (January-February, 1986).

Hydrographic stations were occupied along the cruise tracks (Fig. 4), and moored current, temperature, and pressure measurements were made at key sites between the two expeditions. Acoustic current profiling measurements of the upper 150-450 m were made along the ships' tracks. Continuous temperature and conductivity observations were made at the depth of the ships' hulls, and wind speed and direction were recorded frequently during the cruises.

CTD Data

The temperature, salinity, and density data from CTD measurements are the primary data used in this study. Note that only data from the U.S. cruises have been included in the statistical analysis at this stage, but the statistics reported here will be updated soon. (However, upper ocean temperatures from a 24-hour sequence of CTD profiles on the equator at 150°E, performed from the Australian R/V Franklin, are included here.) The statistical results will not change significantly, because both ships were making observations nearly simultaneously, and the Australian cruise track filled in the center of the large polygon created by the U.S. cruise track (Fig. 4).

Neil Brown Mk IIIB CTD profilers were used during all cruises. Data acquisition and processing procedures are as developed at Woods Hole Oceanographic Institution and described by Fofonoff et al. (1974), with the exception of the U.S. WEPOCS II cruise, where the data were obtained and processed by the PACODF group of Scripps Institution of Oceanography. The modified processing procedures employed by PACODF are not yet documented, but the methods are not grossly different from those employed at Woods Hole. Details of the processing of the hydrographic observations made from R/V Thomas G. Thompson during WEPOCS I can be found in Lukas and Tsuchiya (1986).

The CTD data are processed to 2 decibar averages, starting from the surface. The methods used by Woods Hole result in downcast profiles with averages centered on odd pressure values. The Scripps group processes all data to bins centered on even pressures. This difference is not important for the present study, with the exception of defining sea surface temperature (SST) and sea surface salinity (SSS). Here, I have ignored the surface values ($p = 0$) from WEPOCS II, as there are questions concerning the reliability of the CTD data in this interval. Thus, "SST" and "SSS" refer to the average temperature and salinity between 0 and 2 decibars for WEPOCS I, and between 1 and 3 decibars for WEPOCS II. For the shallow depths under consideration, we will take 1 decibar equal to 1 m, and a mixed layer depth of 1 m (2 m) for WEPOCS I (WEPOCS II) indicates the presence of strong property gradients right to the surface.

Mixed Layer Depth Criteria

Historically, mixed layer and thermocline depths have been estimated using temperature gradient criteria or by specifying a net temperature decrease from the surface. Defant, according to Wyrtki (1964), used a critical temperature gradient of 0.02°C/m to define the thermocline depth, while Wyrtki used a 0.5°C change from the surface value. Levitus (1982) used a net temperature change of 0.5°C from the surface to define the mixed layer depth, but also used a density change of 0.125 sigma-T units, as he recognized the importance of salinity in stabilizing the upper ocean, especially in the subarctic.

In this study, we employ a density gradient criterion as the most reliable estimator of mixed layer depth. The vertical density gradient is a measure of the buoyancy force which must be overcome by the turbulent kinetic energy of the wind forcing to deepen the mixed layer. In addition, we use temperature and salinity gradient criteria to estimate the mixed layer depth, and to estimate the depth to the top of the main thermocline and halocline. The reason for using gradients for criteria is that they are related to the mixing process. The reason for using temperature and salinity is that they have different source functions, and thus offer some independent information on mixing.

For temperature, the critical gradients used in this study are 0.05°C/m (dT1) and 0.025°C/m (dT2). For salinity, the gradients used are 0.02 °/oo (dS1) and 0.01 °/oo (dS2). In density, a gradient of 1×10^{-2} kg m^{-4} (0.01 sigma-T units/m; dD) was used. The values chosen for dT1 and dS1 are roughly equivalent in their influence on density, as are dT2 and dS2. (At T = 29°C, S = 34 °/oo, the ratio of the haline to temperature expansion coefficients is 2.25). The value specified for dD is slightly greater than the density change corresponding to dT2 or dS2.

Starting from the "surface" values, the data were searched downward until the gradient criteria were exceeded for the interval between adjacent 2 m values. The layer depth was assigned the depth value of the shallower of the two data values.

The initial analysis was performed by hand during the WEPOCS II cruise. As each station's data was scanned for the different layer depths, data plots were inspected to see how the various estimates performed under different oceanic and atmospheric conditions. In general, the dT1 criterion produced the deepest layer estimates (Table 1). This depth did not usually correspond to subjective estimates of the mixed layer depth; most often this depth corresponded to the depth of the top of the thermocline. Table 1 shows that the dT2 criterion produced mixed layer depth estimates that were 10 m shallower on average, though the maximum and minimum depths were the same. The density gradient criterion discussed above agreed most closely with subjectively determined mixed layer depths, as found by Peters (1987, this volume) in his Tropic Heat mixed layer work.

Table 1. Basic statistics for wind speed (WSPD), sea surface temperature (SST), sea surface salinity (SSS), and for five different measures of upper layer depth. Data are from 243 CTD stations during WEPOCS I and II.

	WSPD (m/s)	SST (deg C)	SSS (o/oo)	DT1 (m)	dS1 (m)	dT2 (m)	dS2 (m)	dD (m)
mean	5.3	29.10	34.22	51	41	36	31	29
median	5.2	29.10	34.27	57	40	36	26	23
std. dev.	3.4	.65	.46	30	28	28	25	26
max	14.9	30.70	35.14	106	157	106	106	106
min	.0	27.20	31.70	1	1	1	1	1

Many stations were observed to have relatively deep isothermal layers which were not isohaline. In fact, this observation during the WEPOCS I cruise motivated the present study. Obviously, the mixing is not complete over the deeper layer if there are substantial salinity gradients within it. A priori, it was not clear what was "substantial". A variety of salinity gradient criteria could be tried until one is found that produces the same average mixed layer depth determined by one's favorite temperature gradient. This is not the most satisfactory way to proceed, but such a value (0.02 ^{o}/oo) was found (Table 1). However, because of the different surface forcing for temperature and salinity, such an estimate will still vary considerably from the temperature gradient criterion at times. The correlation between layer depths estimated from dT2 and dS2 was only 0.55.

RESULTS

Basic Statistics

Surface variables observed at CTD stations will be used in later sections as proxies for wind and buoyancy forcing. It is recognized that these are poor substitutes for the time integral of wind stress and buoyancy flux which is responsible for the observed mixed layer structure, however such time histories are unfortunately not available.

Table 1 shows basic statistics for wind speed, SST, and SSS observed during WEPOCS. In general, the winds are relatively light at 5.3 m/s, but they are quite variable. Average SST is 29.1 degrees, and does not vary strongly over the WEPOCS region in general; the large range of SST is a reflection of significant warming of the region between cruises, and the rapid cooling associated with a strong wind event encountered during WEPOCS II. As mentioned, the SSS is relatively low, and varies over a large range. Some of the extremely low salinity values are associated with river runoff, which is a significant source of fresh water in the coastal portions of the WEPOCS area, however substantial precipitation (1.7-2.6 m/yr) occurs over the oceanic portions of the WEPOCS region (Rao et al., 1976).

The mean depth to the top of the thermocline (as defined by the dT1 criterion) was 51 m. This depth varied considerably in time and space. The main halocline (dS1) was slightly shallower on average, but exhibited comparable variability.

The mixed layer depth estimated from temperature, salinity, and density gradients all showed comparable variability, and identical ranges. However, the mean depths varied systematically, with the density gradient yielding the shallowest depths, even though the 0.01 kg m^{-4} criterion was larger than the equivalent density change of the dT2 and dS2 criteria. The average mixed layer depth of 29 m is shallower by a factor of 3 than the commonly accepted value of 100 m.

The upper layer thickness (Fig. 5a) exhibits a bimodal distribution, so the average and standard deviation are inadequate descriptors. There is a minimum frequency near 20 m, which suggests that the dT1 criterion is picking the top of the thermocline most of the time, but it also picks out the mixed layer depth a substantial fraction of the time. The minimum in the frequency distribution illustrates the substantial separation between the bottom of the mixed layer and the top of the thermocline.

The three mixed layer depth estimators are also not normally distributed (Fig. 5b-d), with each having a mode in the 0-5 m class. The general shape of these distributions is exponential. Such a probability density function is characteristic of a Poisson process (cf. Mohanty, 1986). The implications of this are pursued in the discussion.

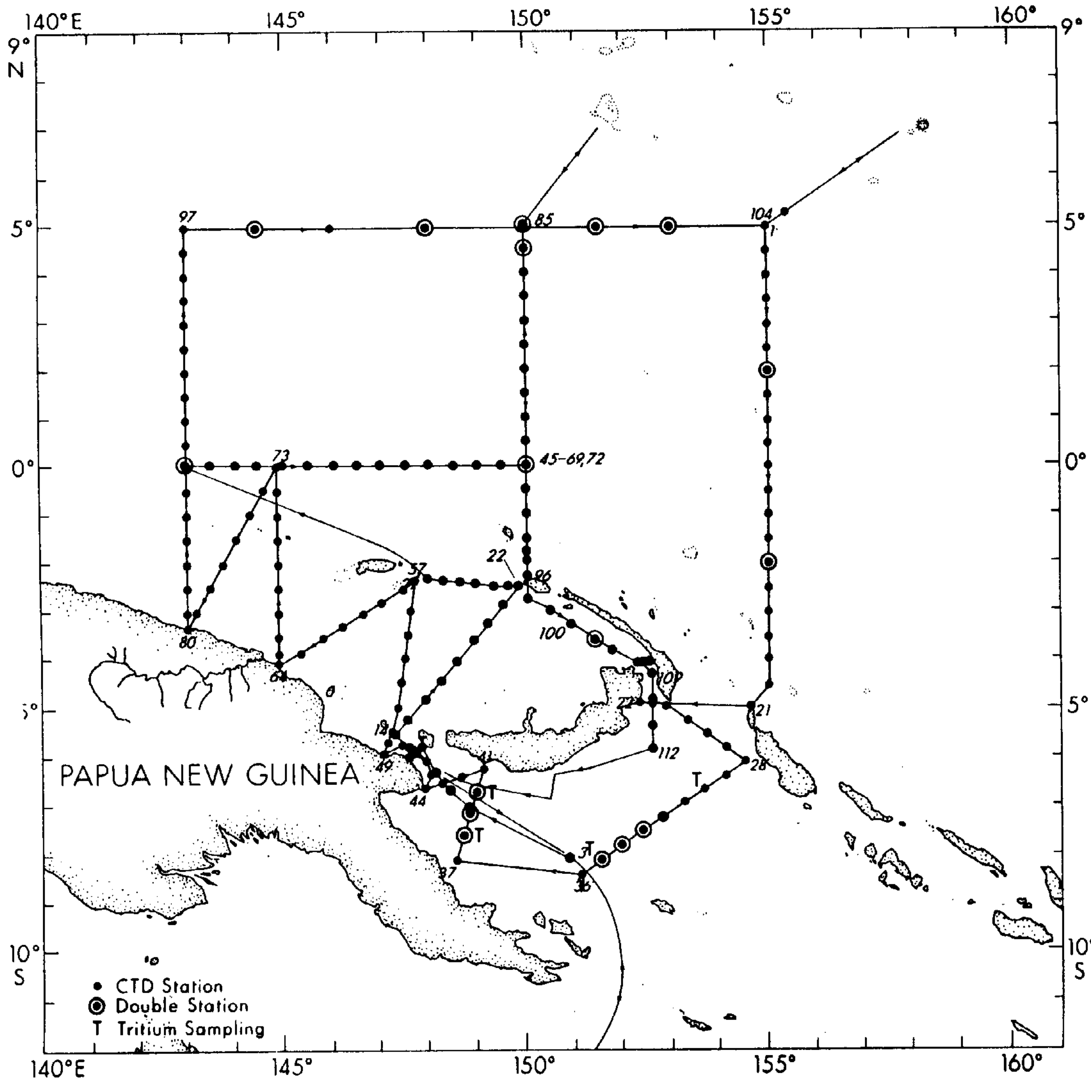

Figure 4. The cruise tracks and hydrographic station locations occupied by the R/V Moana Wave and the R/V Franklin during the WEPOCS II expedition, January-February 1986. The WEPOCS I cruise tracks and station positions were only slightly different.

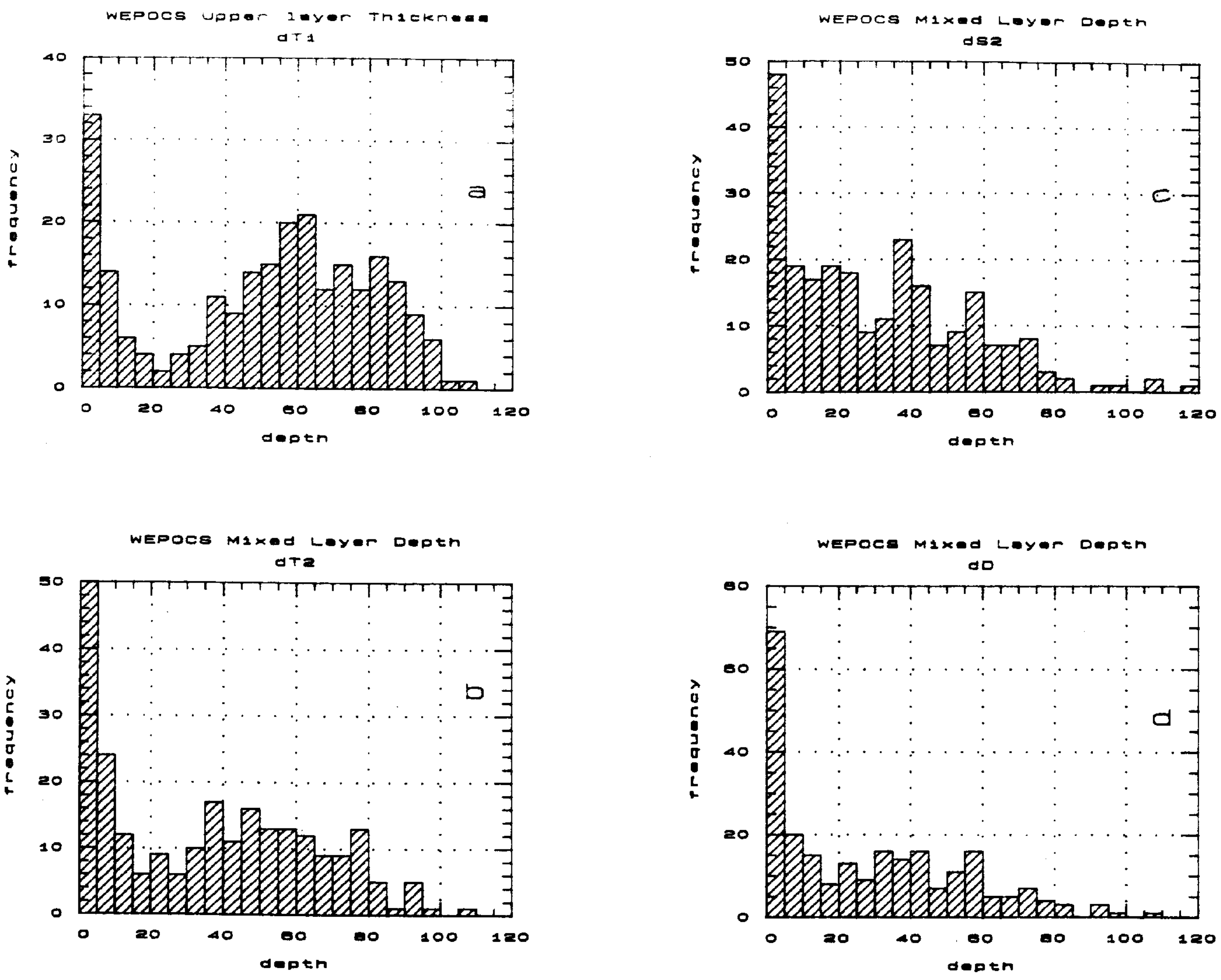

Figure 5. Histograms of the upper layer depth (a), and estimates of the mixed layer depth using temperature (b), salinity (c), and density (d) gradient criteria. See text for definitions and methods.

These mixed layer depth distributions suggest that shallow layers are fairly common in the western equatorial Pacific. In fact, these observations might be typical of summer in midlatitudes under light or no wind conditions, where diurnal warming is important. Pronounced effects of diurnal warming have been observed in the central equatorial Pacific mixed layer during Tropic Heat by Gregg et al. (1985). To investigate this possibility, the mixed layer depth observations were plotted against the time of day that the station was started (Fig. 6). It is obvious that there is no general correspondence between mixed layer depth and time of day. The observations were stratified into three different wind speed classes in the figure. Only during very light winds (speeds less than 3 m/s) is there a hint of a diurnal dependence, with these observations being shallow between 0600 and 1800. However, these very shallow mixed layers were found frequently during stronger winds as well (Fig. 7). Also, there is no apparent tendency for deep values to occur during the night hours, as is the case for the central Pacific (Gregg et al., 1985).

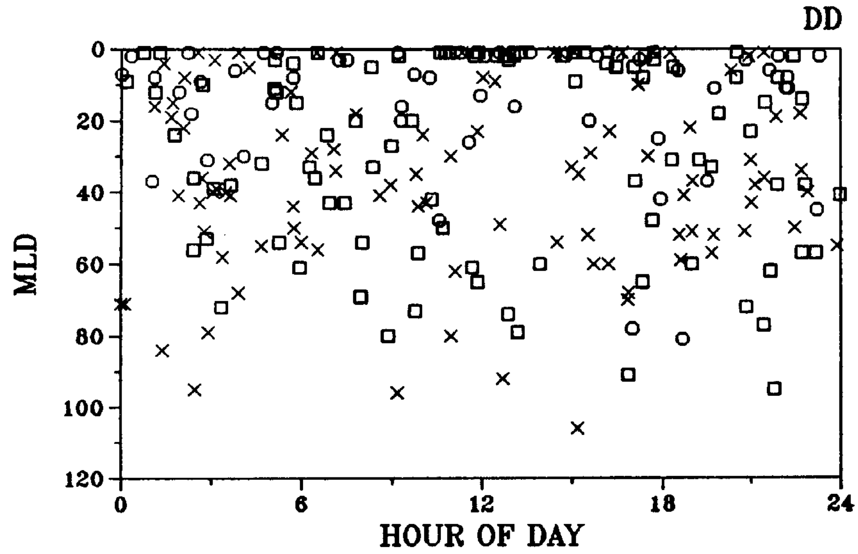

Figure 6. Mixed layer depth estimated from density gradient versus hour of the day (local time), for wind speed less than or equal to 3 m/s ([]), for wind speeds between 3 and 6 m/s (0), and for wind speeds greater than 6 m/s (X).

The average temperature and salinity of the mixed layer (as defined by density) were computed for each station, and the resulting frequency distributions are shown in Fig. 8. The mixed layer temperature shows a large peak in the 28.75–29.00°C class, for which an explanation does not exist. Further, there is a tendency towards a bimodal distribution which was quite pronounced when data from WEPOCS II were analyzed separately.

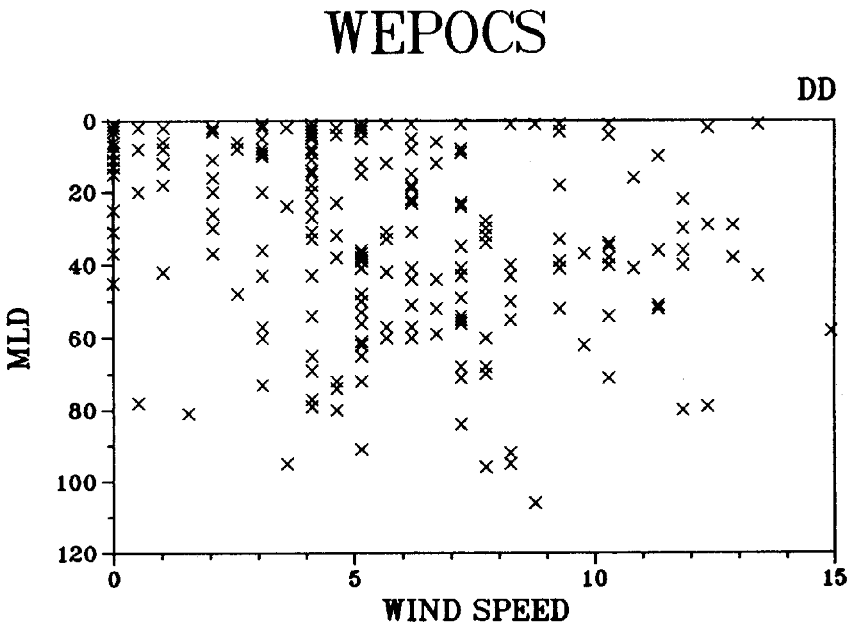

Figure 7. Mixed layer depth estimated from density gradient versus wind speed observed during CTD stations.

For those cases, there were sharp peaks in the 28.75-29°C and 29.75-30°C classes, with very few cases occurring in the 29-29.25°C class. The likely explanation for this situation is the strong burst of westerly winds that occurred in the middle of the WEPOCS II cruise, with wind speeds up to 15 m/s. The mixed layer temperatures during and after this wind event were cooler by one degree on average.

The mixed layer salinity distribution is highly skewed towards negative values, illustrating the influence of precipitation on the mixed layer of the western equatorial Pacific Ocean. The counter balancing influence on the high salinity side of the distribution is the high salinity layer of southern hemisphere subtropical water found in the thermocline over much of the WEPOCS region (Lindstrom et al., 1987).

A new variable was formed to estimate the thickness of the layer between the bottom of the mixed layer and the top of the thermocline. This is called the "barrier layer" by Stuart Godfrey (personal communication, 1987), and its existence in the western equatorial Pacific was previously unknown. The top of the thermocline is estimated using the dT1 temperature gradient criterion, but the first 10 m of the CTD profiles are ignored so that the shallow mixed layer is not distorting the signal. The difference between this depth and the mixed layer depth is then the thickness of the barrier layer. The frequency distribution for this variable is shown in Fig. 8c. Again, the distribution suggests an exponential density function. The mean thickness of this layer is 31 m, with half the observations deeper than 24 m. An explanation for the existence of this layer is suggested in the discussion.

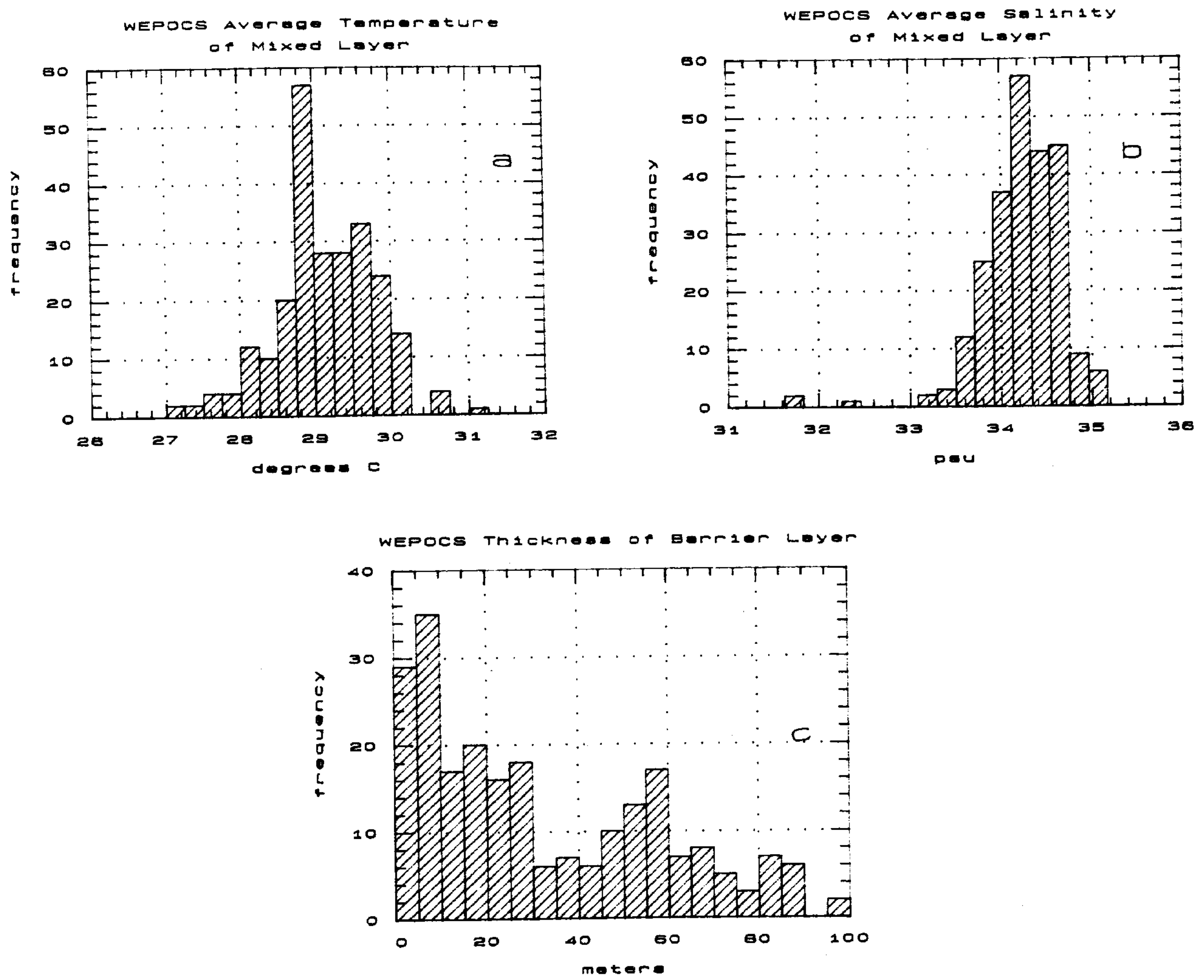

Figure 8. Histograms of average mixed layer temperature (a), average mixed layer salinity (b), and thickness of the layer between the bottom of the mixed layer and the top of the thermocline (c).

Figure 9 shows the temperature during the 24-hour time series CTD station on the equator at 150°E. The profiles were made at one hour intervals to a maximum depth of 400 m, though only the upper 100 m is shown. There are several features to point out in this figure. First, warmest SST occurs at 1600 local and coolest SST is found at 0800, consistent with the diurnal surface warming discussed earlier. Second, the mixed layer depth changes with the diurnal warming, but doesn't get deeper than about 20 m. Third, there is a large separation between the 29.4 and 29.3 isotherms relative to neighboring isotherms. Fourth, there is persistent temperature inversion of about 0.2°C near 70 m. This inversion is stable, being compensated by the salinity stratification. Finally, the top of the thermocline (as defined earlier) is found at about 75 m, and the layer between the top of the thermocline and the bottom of the mixed layer is about 50 m thick.

Correlation Results

It might be useful to examine the correlations between the surface variables and the mixed layer depths. Table 2 presents the correlations of the three different mixed layer depth estimators with the wind speed, SST, and SSS. These latter three variables are treated as proxies for the surface forcing from wind, heating, and precipitation. Also shown are the correlations of the discrepancy between the temperature and salinity gradient estimates of mixed layer depth with the three "forcing" variables.

Table 2. Correlation matrix of surface variables versus mixed layer depth estimated using density (dD), temperature (dT2), and salinity (dS2) gradient criteria from 243 CTD stations. Underlined values are considered statistically significant at the 95% confidence level, assuming 60 degrees of freedom (standard error of correlation coefficient = 0.13).

	WSPD	SST	SSS
dD	<u>.31</u>	-.19	.24
dT2	<u>.36</u>	-.22	.20
dS2	.20	-.10	<u>.33</u>
dT2-dS2	.20	-.15	-.11

The standard errors of the correlation coefficients were calculated using 60 degrees of freedom. This value assumes that each day's data during the cruise is independent from the other data. This is not strictly true, but perhaps the lack of perfect correlation within each day's observations offsets this factor. Future analysis will attempt to derive a better estimate of the degrees of freedom of these data.

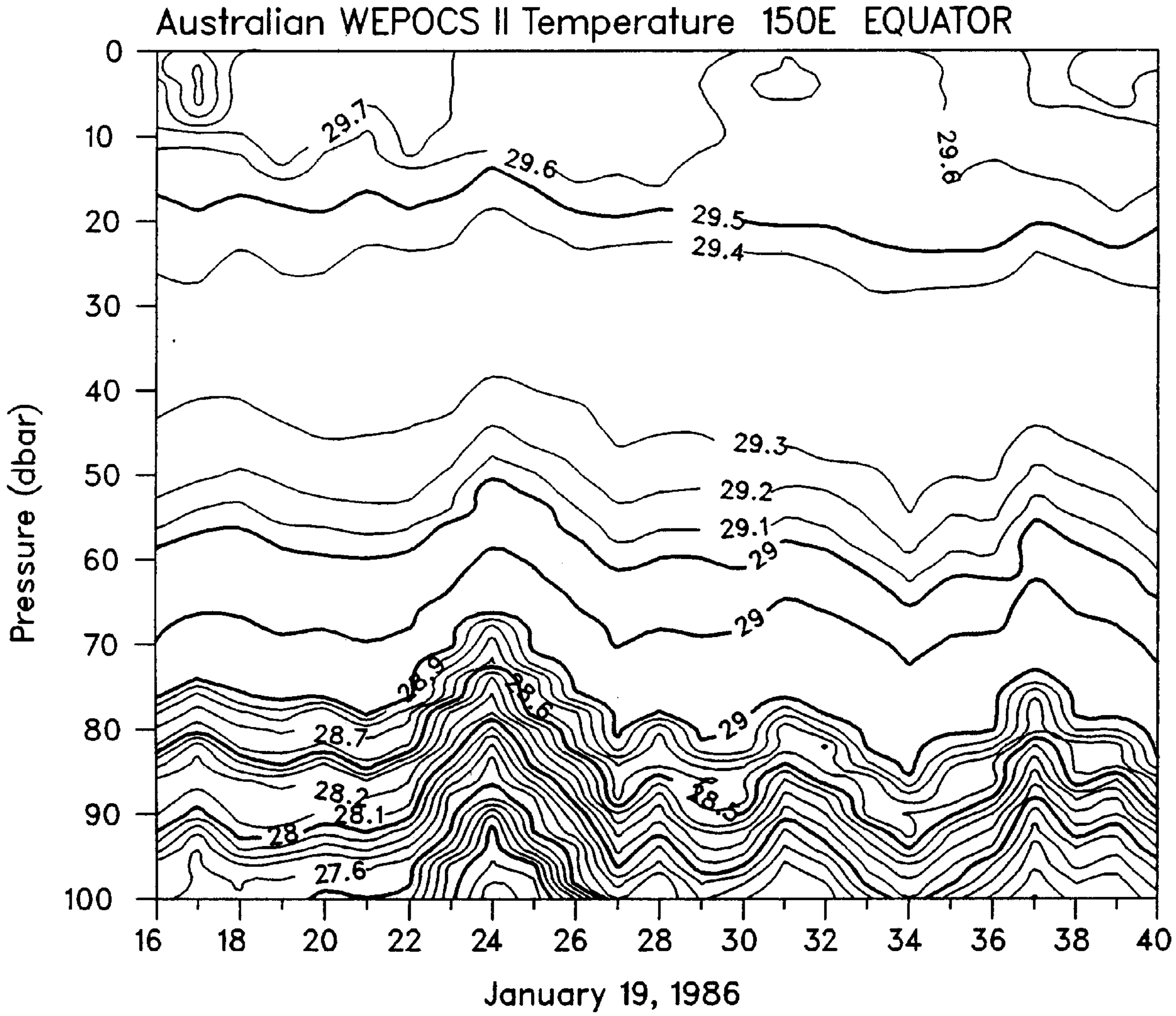

Figure 9. Time series of upper ocean temperature from hourly CTD casts on the equator at 150°E. The times are local.

At the 95% confidence level (r=0.26), there are three significant correlations. Both the density and temperature mixed layer depth estimates are positively correlated with the wind speed, suggesting a tendency for increased mixed layer depth with stronger wind. Figure 7 illustrates this correlation, though the percent of variance explained by the correlation is relatively small. The correspondence of wind speed and dS2 is not significant, but is of the same sense. The correlation of dS2 with SSS is significant though, which might explain the reduced dependence on wind speed. Note that these correlations were also run with wind speed squared (psuedostress), but the correlation coefficients actually were smaller.

At 90% confidence, dD shows the same correlation with SSS as dS2, which is consistent with the idea that surface buoyancy forcing associated with precipitation is an important factor in determining the mixed layer depth. Also, dT2 shows a negative correlation with SST, which indicates a tendency for the mixed layer depth to be shallower when SST is high, which might occur after maximum diurnal heating.

While there are no significant correlations between the surface variables and the difference of dT2 and dS2, the signs of the correlations are consistent with the idea that precipitation and heating are responsible for the mismatch in depths.

DISCUSSION

The Shallow Mixed Layer, and the Importance of Salinity

The western equatorial Pacific is a region where the warmest open-ocean sea surface temperatures in the world are found, and it is characterized by relatively low surface salinity due to frequent and heavy precipitation (Weare et al., 1981). In contrast to the central equatorial Pacific, the winds in the west are weak on average, and highly intermittent. The relative importance of wind and buoyancy forcing between these two regions is very important for understanding the distribution of mixed layer depth.

The new WEPOCS observations demonstrate that the mixed layer in the western equatorial Pacific is shallow in general, averaging 30 m. The top of the thermocline is found at an average depth of about 50 m. (DT1 in Table 1). The situation in Fig. 2 is quite exceptional. In the central Pacific, the observations from Tropic Heat show that the mixed layer is about 40-50 m deep, and there is a pronounced diurnal cycle of mixing associated with convective overturn due to nighttime cooling at the surface (Gregg et al., 1985). In the western Pacific, the stratification is almost always stable because of the strong buoyancy forcing associated with an excess of precipitation (P) over evaporation (E). This gradient in E-P can be seen in Fig. 10. The most negative values of E-P are centered in the WEPOCS region, and relatively small values are found on the equator near 140°W where Tropic Heat was conducted.

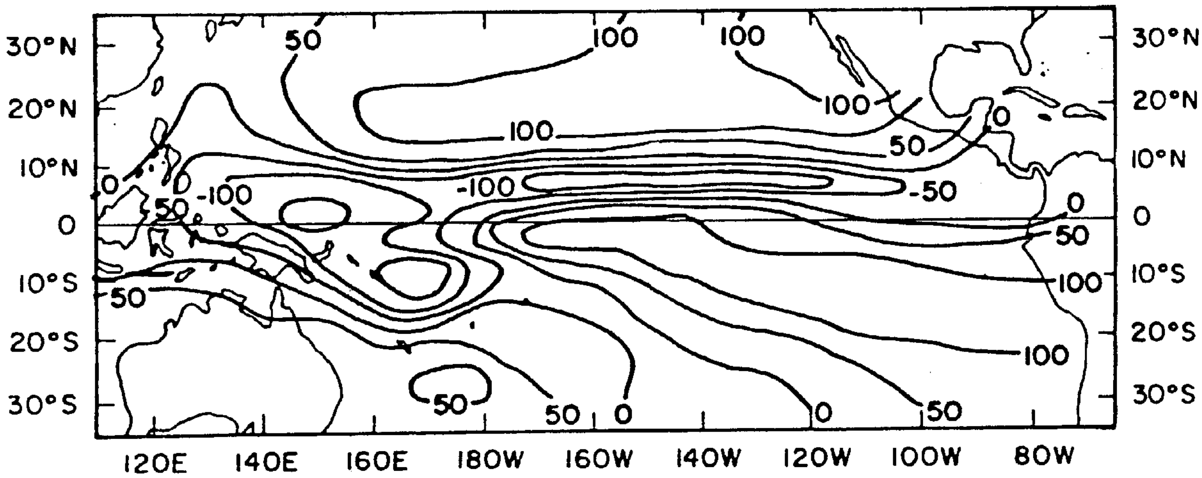

Figure 10. The annual average distribution of net evaporation over precipitation in heat flux units (W m^{-2}) [After Weare et al., 1981].

Salinity is usually neglected in models of the mixed layer. However, Miller (1976) investigated the effects of including salinity in the one-dimensional mixed layer model of Denman (1973). He found that the vertical salinity profile can play a critical role in determining the evolution of the depth of the mixed layer, and can even determine whether the mixed layer warms or cools, for given surface forcing. In one particular experiment, the addition of a stable salinity stratification with all other factors held constant resulted in an equilibrium mixed layer depth of 50 m versus 90 m without salinity.

This effect can be simply understood by comparing the available turbulent kinetic energy supplied by the wind versus the work that must be done to overcome potential energy when deepening the mixed layer. If the mixed layer is shallower as a result of the stabilizing effect of the salinity profile, then the net heat flux at the sea surface is distributed over a thinner layer, which will result in a warmer mixed layer. This assumes, however, that there are no external feedbacks between net heat flux, precipitation, and wind stress, and no feedbacks between any of these variables and SST. This assumption is of course absurd for the western equatorial Pacific, where the coupling between atmosphere and ocean is quite strong and complex.

The Subduction Hypothesis

The warm, nearly-isothermal upper layer of the western equatorial Pacific is deeper than in the central and eastern equatorial regions. Garwood et al. (1985) attribute this to greater mixing, but three-dimensional equatorial circulation models (eg., Busalacchi and O'Brien [1980]) clearly show that this situation is explainable in terms of the horizontal convergence of mass in the warm upper layer, driven by the mean wind stress distribution. This provides the backbone for a hypothesis to explain the mismatch in vertical scales between the mixed layer depth and the nearly-isothermal layer in the western Pacific. As in the model of Atlantic 18°C water formation by Woods and Barkmann (1986), we believe that mixed layer waters formed in the near-equatorial region in the vicinity of the dateline are subducted below the extremely light surface waters of the western equatorial Pacific, as they are moved westward in the South Equatorial Current. Meridional circulation in the western Equatorial Pacific may well contribute.

In much the same way that the large-scale meridional distribution of net E-P and heating determines the vertical T-S relationship (see Worthington [1981] for a thorough discussion), the horizontal distribution of net E-P and heating in the tropical Pacific determines the vertical T-S relationship of the upper ocean in the western equatorial Pacific.

The central equatorial Pacific is a region of net evaporation, while the western equatorial Pacific is a region of heavy net precipitation. The strong zonal gradient of E-P can easily be seen in Fig. 10. Therefore, the mixed layer waters formed near the dateline are saltier than those formed further to the west, but are of nearly the same temperature (Reynolds, 1982). As these denser waters are advected westward, they must be subducted below the mixed layer of the western region. In this way, the warm upper layer of the western Pacific can increase in thickness, while the layer where active mixing occurs can be quite a bit shallower.

Recall that temperature inversions of up to 0.5°C were observed frequently during WEPOCS I (Fig. 11), and Meyers (personal communication, 1987) observes these in XBT traces from this region at other times. Also, the mean temperature at 50 m was warmer than that at 15 m as measured from the WEPOCS equatorial mooring at 150°E during the period between the two expeditions. This could be a signature of the subduction process. If the warmest surface waters are found further to the east, as in an average July or August (Reynolds, 1982), then the subducted water can be even warmer than that of the mixed layer in the WEPOCS region. However, such a stable temperature inversion at the base of the mixed layer can also result from strictly one-dimensional processes as found by Miller (1976). Further observations must be made to determine the cause of these persistent temperature inversions.

Intermittent Westerly Bursts

The subduction hypothesis can explain the mean upper ocean structure, but there are seasonal and faster time scale variations of winds, currents, and heat fluxes which may mask this process. An example of this is seen in the response of the western equatorial Pacific upper ocean to a westerly wind burst such as observed during WEPOCS II. Between 27 January and 2 February 1986, westerly winds between 15 and 30 knots were blowing over the near-equatorial region north of New Guinea.

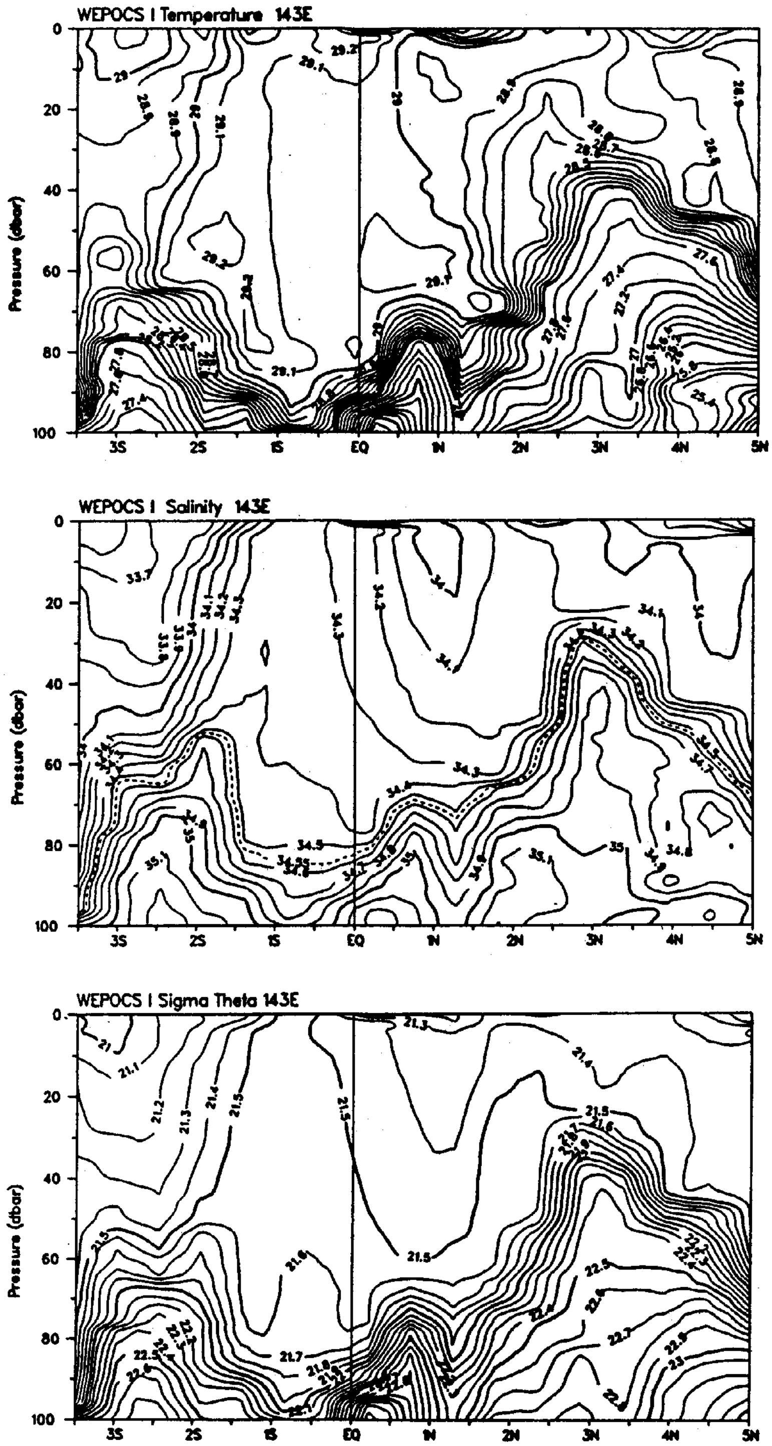

Figure 11 (a). Temperature, salinity, and density (from top to bottom) for WEPOCS I along 143°E. Data from CTD stations at 0.5° latitude spacing.

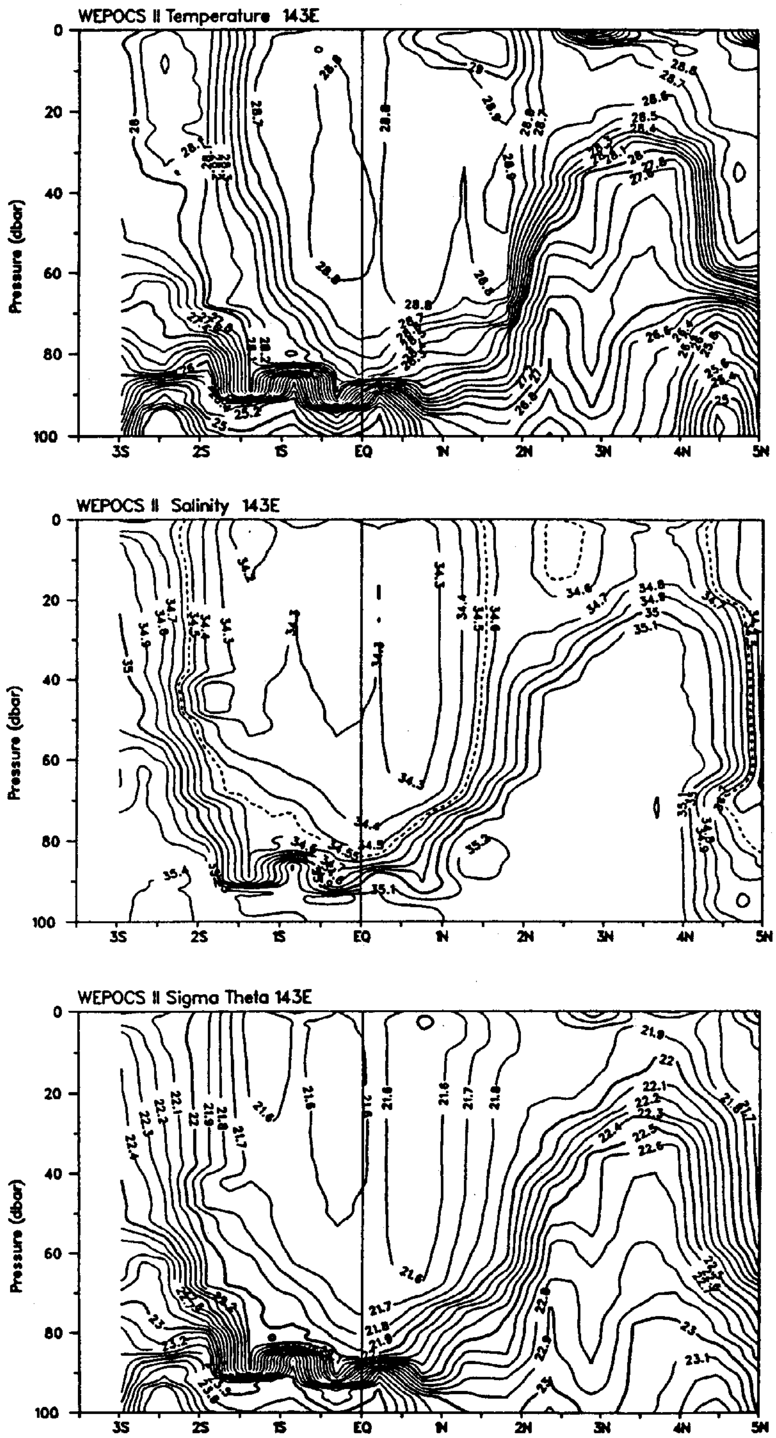

Figure 11 (b). Temperature, salinity, and density (from top to bottom) or WEPOCS II along 143°E. Data from CTD stations at 0.5° latitude spacing.

The climatological winds for January from the Australian Bureau of Meteorology Research Center (Fig. 12) show that, on average, the winds in the western equatorial Pacific are very light at this time of year, and the wind field is highly convergent. The winds veer from northeasterlies north of the equator to northwesterlies south of the equator. However, this smooth picture is the result of a highly variable wind field, both within the month and from year-to-year. Keen (1987) has developed a climatology of westerly wind bursts which shows that this area experiences such wind events more often than regions to the east, and that these events are far more frequent in the northern winter season. It is important to note that there are exceptionally few cases of easterly bursts, and the distribution of zonal wind in this region is highly skewed towards positive (eastward) values (Lukas et al., 1984).

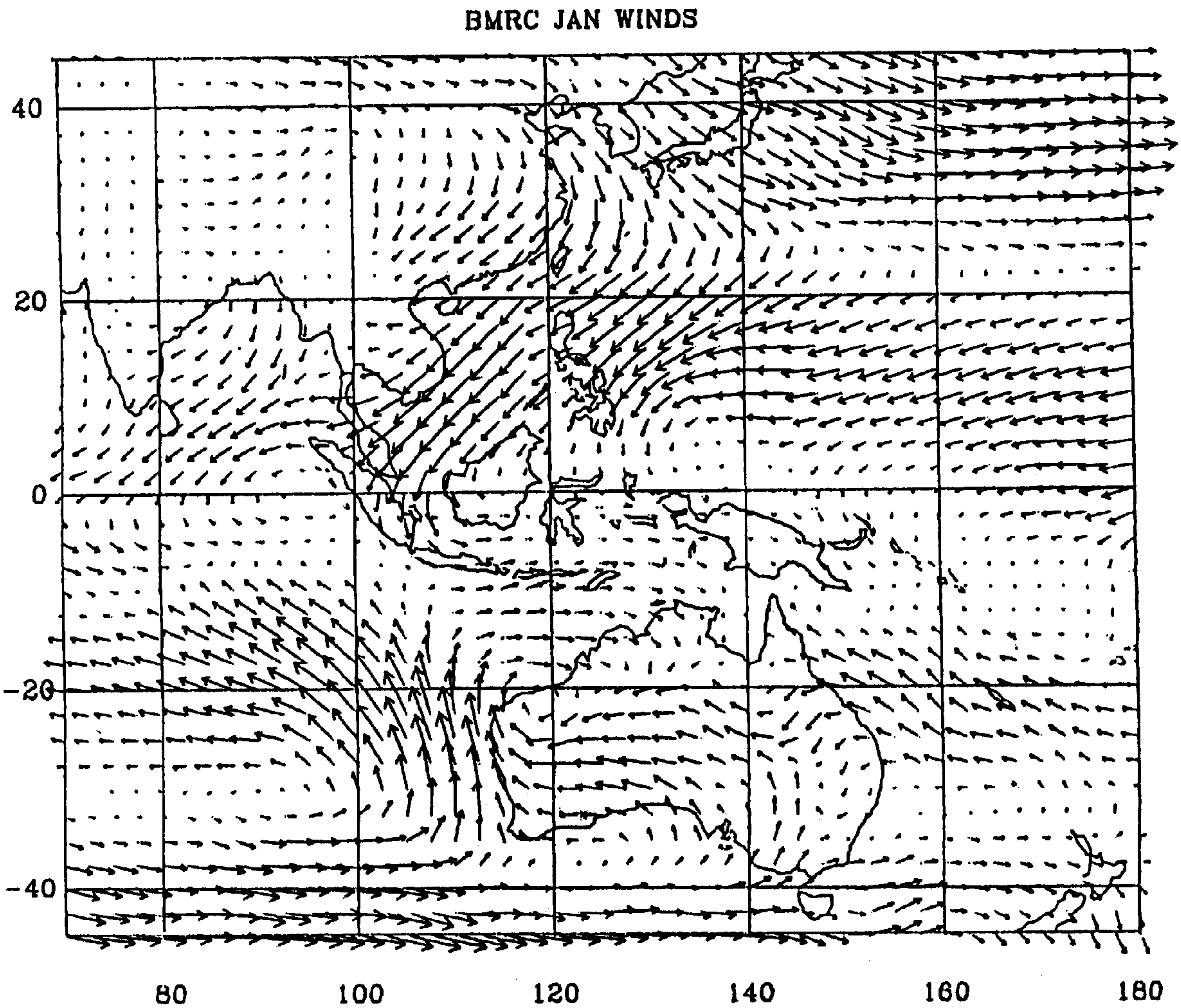

Figure 12. Climatological mean January wind vectors from the Australian Bureau of Meteorology Research Center (courtesy of G. Meyers).

Such intermittent and strong wind forcing events impulsively force the ocean, and they are responsible for the generation of equatorial Kelvin wave pulses (Lukas et al., 1984), among other effects. It is highly likely that these westerly bursts cause a substantial change from the usual fluxes of moisture and heat as well. (Evidence for such a change is found in the mean mixed layer temperatures observed before and after the burst.) There is certainly greater evaporative cooling, but the fresh water balance is not clear. Also, the cloud patterns associated with a westerly burst may affect the insolation in a way that is not obvious.

It was noted earlier that there was a bimodal distribution of mixed layer temperature during WEPOCS II, and that when these observations were separated into the two dominant classes, the observations fell into two time periods separated by the westerly wind event. As the cruise track did not repeat, it remains a possibility that spatial gradients were partly responsible. However, there were no strong spatial gradients of SST observed in the thermosalinograph records along the cruise track, as would be required to explain such a bimodal distribution. The mean SST before the event was 29.8°C, and 28.6°C after the event. The average mixed layer depth was 22 m before, and 44 m during and after the wind event. A portion of this difference may be due to real spatial variations, however. The change in SST is quite large by historical measures of western equatorial Pacific SST variability, and is associated with latent heat fluxes (Meyers et al., 1986), though entrainment of cooler thermocline waters into the mixed layer was probably a contributing factor.

The effect of such impulsive forcing on the western equatorial Pacific upper ocean can be seen in Fig. 11 from the contrast of the temperature, salinity, and density fields along 143°E from WEPOCS I (light winds) and WEPOCS II (just after the westerly burst). The near-equatorial temperatures are higher during WEPOCS I than during WEPOCS II, and the salinity is lower. This is consistent with a cooling of the upper ocean by the strong winds, and with enhanced evaporation. Also, the density field is markedly different. During WEPOCS II, density is symmetric with respect to the equator, and the mixed layer is deeper especially on the equator. Westerly winds straddling the equator cause an Ekman convergence at the equator, resulting in a downwelling response there. This circulation sets up in a matter of only a few days (Cane, 1980), and has the effect of concentrating mixed layer waters at the equator, which are then possibly detrained from the actively mixed layer. Upwelling occurs away from the equator as warm upper layer waters close to the equator are advected more rapidly than waters further away. The exceptionally large change in the hydrographic structure apparent near 3°S during WEPOCS II is due to upwelling along the Papua New Guinea coast induced by the westerly wind burst.

Recall that the distribution of mixed layer depths was found to be exponential which is characteristic of a Poisson process. This suggests physics which are intermittent or event-like. Either the boundary conditions (the winds stress and buoyancy flux) or the dynamics (e.g., entrainment) or both may be intermittent. The combined influence of intermittent forcing and the barrier layer is illustrated schematically in Fig. 13, where entrainment cooling switches on when the mixed layer deepens enough to reach the substantial temperature gradients found at the top of the thermocline. In a Poisson process, events occur randomly, but there is an average recurrence rate. It is appropriate to treat westerly wind bursts as discrete events because a) they are relatively rare, and b) there are no easterly bursts. For a nonstationary Poisson process, the rate parameter is a function of time, and for the westerly wind bursts, the rate parameter varies seasonally and interannually as evidenced by the climatology of such events constructed by Keen (1987).

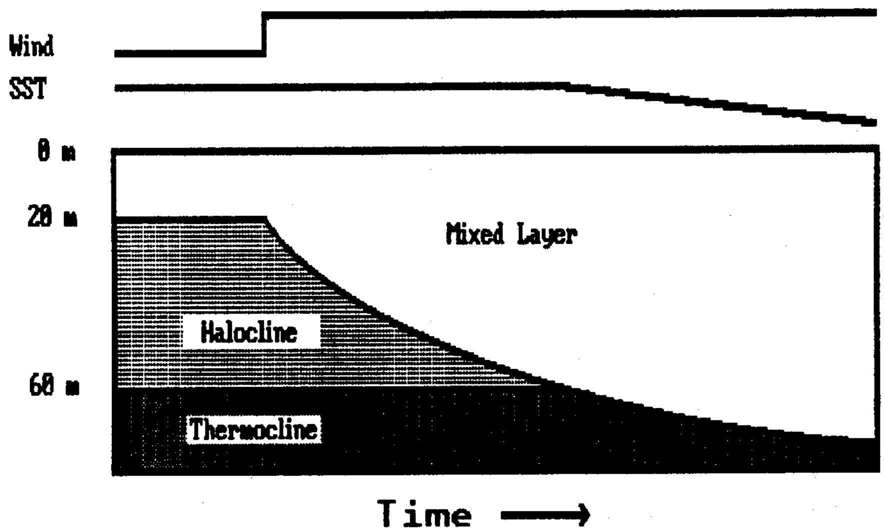

Figure 13. Schematic showing development of mixed layer depth and sea surface temperature when strong winds are turned on. Entrainment cooling of the mixed layer only begins when mixed layer has deepened through salt-stratified barrier layer.

CONCLUSIONS

The new observations from the Western Equatorial Pacific Ocean Circulation Study have shown that the mixed layer in this important region is much shallower than previously thought, averaging only about 30 m. This appears to be a result of strong positive buoyancy forcing associated with heavy precipitation, combined with highly intermittent wind forcing.

The vertical scale of temperature in the upper ocean of the western equatorial Pacific is larger than for salinity, which is probably a reflection of the larger zonal scale for SST along the equator than for precipitation. A hypothesis of subduction of saltier mixed layer waters from the east below the fresh mixed layer of the western equatorial Pacific appears to explain this aspect of the WEPOCS observations. New datasets are required to properly test this hypothesis.

The primary implication of the vertical mismatch in scales of the upper ocean thermal and haline structures in the western Pacific is that there is very little vertical heat flux out of the mixed layer since there is usually only a very weak vertical temperature gradient at the base of the mixed layer. There can be no entrainment of thermocline waters into the mixed layer except during very strong wind events, because of the strong stratification associated with the vertical salinity profile. Such entrainment was considered to be a continuous process by Niiler and Stevenson (1982), but these new observations suggest that entrainment might be very rare in the western equatorial Pacific. Thus, the mixed layer can easily warm in the absence of westerly wind bursts, which can erode the salinity-stratified warm upper layer and switch on entrainment cooling as a mechanism to modify the mixed layer heat budget (Fig. 13). This would imply that net cooling only occurs during the onset of ENSO events, when westerly bursts are more frequent (Luther et al., 1983; Keen, 1987). For the long term mean, there can be no net temperature increase of the mixed layer of course, but the ENSO phenomenon may exist solely because there is an inadequate removal of heat from the upper layer of the western tropical Pacific Ocean by the mean ocean circulation, as suggested by Wyrtki (1985).

We hypothesize that the switching on of the entrainment mechanism by the frequent and strong westerly bursts observed prior to the onset of ENSO conditions (Keen, 1987) is the essential process needed to allow the ocean warm pool/atmospheric convection supercluster to break away from the western boundary region and begin its eastward migration characteristic of the early phase of ENSO (Gill and Rasmussen, 1983). Further, we speculate that warming of the western equatorial Pacific sea surface increases the frequency and strength of such westerly bursts, effectively resulting in a negative feedback mechanism that is an essential part of the ENSO cycle. New observations are planned which will help to refine and test this hypothesis.

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