Variability in the Western Equatorial Pacific Ocean during the 1986–87
El Niño/Southern Oscillation Event*

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ABSTRACT

We describe variability in the western Pacific Ocean during the 1986–87 El Niño/Southern Oscillation (ENSO) event, with emphasis on time series measurements of currents, temperature, sea level and winds near the equator at 165°E. Zonal winds were anomalously westerly from mid-1986 to late 1987 and were punctuated by 2–10 m s⁻¹ episodes of westerlies lasting about 10 days to 2 months. Zonal currents in the upper 100-m surface layer responded to these wind variations typically within a week, in some cases with speeds exceeding 100 cm s⁻¹ to the east. Zonal current variations in the thermocline below 100 m were generally less coherent with the local winds than currents near the surface. They were also generally less variable, although the Equatorial Undercurrent disappeared for 3–4 weeks in October-November 1987 at a time when the normal eastward directed zonal pressure gradient force reversed along the equator. Periods of intense and prolonged eastward flow in the surface layer were associated with a decrease in sea level by 10–20 cm at the end of 1986 and in May–August, 1987. Similarly, significant westward flow near the surface and in the thermocline in September-November 1987 was accompanied by rising sea level and a westward migration from the date line of surface waters > 30°C. These results suggest that wind-driven zonal currents at the equator were important in the evolution of the mass and heat balance of the western Pacific during the 1986–87 ENSO. Conversely, meridional wind stress and meridional velocity energy levels at periods longer than 100 days on the equator were 5–10 times weaker than in the zonal direction and less obviously related to the evolution of the 1986–87 ENSO.

1. Introduction

Our ability to accurately model and predict the evolution of El Niño/Southern Oscillation (ENSO) events is limited in part because the oceanic processes that redistribute mass and heat in tropical Pacific are poorly understood, particularly west of the dateline. Wyrtki (1985), for example, suggested that the warm surface layer of the western tropical Pacific was advected eastwards during the 1982–83 ENSO by anomalous wind driven currents at a rate of 40 Sv (1 Sv = 10⁶ m³ s⁻¹). About 10 Sv of this anomalous eastward mass flux occurred in the North Equatorial Countercurrent in late 1982 (Meyers and Donguy 1984). Much of the remaining 30 Sv may have occurred near the equator where model studies (e.g., Gill and Rasmusson 1983; Philander 1983) indicate large eastward mass and heat fluxes occur during ENSO.

The importance of ENSO-related variations in the Equatorial Undercurrent and South Equatorial Current near the equator west of the dateline has been difficult to assess from observations however, because until recently most studies of upper ocean current variability in the equatorial western Pacific have relied on ship drift data and profiling current meter surveys. Geostrophy is of limited value near the equator because it lacks validity for meridional currents and accuracy for zonal currents (e.g., Picaut et al. 1989); moreover, it is not relevant to motion on time scales less than about a week. Thus, the full spectrum of variability of relevance to ENSO (daily to interannual time scales) is inadequately described by the present database. Specifically, direct current measurements indicate the existence of an Equatorial Undercurrent in the western Pacific (Montgomery 1962; Philander 1973), but almost nothing about its seasonal or interannual variability. Ship drift data analyses (Meehl 1982) provide a measure of the mean seasonal cycle of surface currents, but with coarse horizontal resolution (5° x 5°) due to the sparsity of data. Furthermore, shipboard measurements are badly aliased due to the existence of a broad spectrum of energetic equatorial internal
waves (Wunsch and Webb 1979; Eriksen 1982) and to short time scale wind-forced current variations. Hisard et al. (1970), for instance, found that the South Equatorial Current within 2° of the equator could reverse within a week in response to westerly wind burst forcing. Resolution of this short time-scale variability may be necessary for a complete description of the ENSO phenomenon since repeated wind burst forcing often occurs prior to and during ENSO events (Keen 1982; Luther et al. 1983). The cumulative effect of such forcing may be to displace warm water eastward via anomalous advection both in the forced region and in the eastern Pacific via Kelvin wave radiation.

The purpose of this study therefore is to describe variability in the western equatorial Pacific Ocean during 1986–88, with emphasis on time series measurements near 165°E. The data span the 1986–87 ENSO, so for the first time it is possible to describe ENSO related current variations west of the date line from daily to interannual time scales. The relationship of these variations to temperature, sea level, zonal pressure gradient and wind variations is a central focus of this paper. The longest previous time series of upper ocean equatorial currents in this region were from a 6-month mooring deployment at 0°, 150°E during August 1985–January 1986 just before the ENSO developed (Lindstrom et al. 1987).

The basin scale evolution of the 1986–87 ENSO has been described in Kousky and Leetmaa (1989) using operational atmospheric and oceanic models and data analyses. Anomalous sustained low level (850 mb) westerly winds appeared west of the date line in mid-1986 and were followed by widespread sea surface temperature anomalies > 1°C warmer than climatology in the eastern and central Pacific. Anomalous deep convection and associated atmospheric heating developed over the warmest waters in the tropical Pacific as they migrated eastward towards the dateline in late 1986 (see also Fig. 1). The Southern Oscillation index, a measure of the strength of the tradewinds, decreased to a minimum in March–May 1987 (reaching its lowest values in over 20 years, excluding the 1982–83 ENSO). These anomalous conditions, which resemble the mature phase of the Rasmussen and Carpenter (1982) composite ENSO, persisted until near the end of 1987. By early 1988, the Southern Oscillation index rebounded to near normal and anomalously cool SSTs and strong tradewinds appeared across the basin.

The remainder of this paper is outlined as follows. Mooring and hydrographic data sources are described in section 2. This is followed by a description of wind, temperature, surface height and current time series during January 1986 to May 1988 in sections 3–6, respectively. Section 7 is a discussion of the relationship

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**Fig. 1.** SST charts for February and December 1986, October 1987 and May 1988. SST contours are 1°C except dashed line which is for 29.5°C; data are from NOAA’s Climate Analysis Center SST analyses. Superimposed are the locations of the US/PRC current meter mooring (●), thermistor chain moorings (○), Nauru and Kapingamarangi (●). The zonal extent of CTD sections used in this study is also indicated. Note that prior to the ENSO (February 1986) and after the ENSO (May 1988), the warmest surface waters in the western tropical Pacific tend to be concentrated west of 165°E, whereas during the ENSO (December 1986, October 1987) the warmest waters tend to be concentrated east of 165°E.
between zonal current variations based on mooring data and zonal pressure gradient variations based on CTD data. The paper concludes in section 8 with a discussion of the most important results and their relevance for understanding the ENSO phenomenon.

2. Data

The moored time series used in this study consist of current, temperature and wind data from the locations shown in Figs. 1 and 2. Measurements from taut-wire surface moorings were made at the nominal positions of 2°N, 0° and 2°S along 165°E in water depths of approximately 4.4 km. The lengths of the moored time series and depths instrumented are shown in Fig. 3. The current meter mooring at 0°, 165°E was equipped with EG&G Model 610 Vector Averaging Current Meters (VACMs) at 5–6 depths in the upper 300 m. Seven additional depths were instrumented with SeaData temperature recorders (TRs). Temperature and velocity data were recorded at 15-minute intervals and then processed to daily averages. The VACM measures current speed from a unidirectional 16 × 18 cm Savonius rotor and direction from a 9 × 17 cm

Fig. 2. As in Fig. 1, except for the 3-month sequence, September–November 1987. Note westward migration of water > 30°C.

Fig. 3. Record lengths of moored current, temperature and wind time series; and island sea level and wind time series used in this study.
pivoted vane with a time constant of about 1.5 sec. The error in velocity components due to high frequency current rectification and mooring motion is expected to be <5 cm s\(^{-1}\) for daily averages based on comparisons with Vector Measuring Current Meters (VMCMs) which are generally considered to be more accurate in variable flow regimes (Halpern 1987a). Instrumental accuracies of the VACM and TR temperature sensors is approximately 0.01° and 0.05°C, respectively.

Winds from the equatorial current meter mooring were sampled with a Vector Averaging Wind Recorder (VAWR) mounted 4 m above the mean waterline on the surface toroid. The VAWR is an inverted VACM equipped with Climet cup model 011-2B 3-cup anemometer and pivoted vane. Laboratory calibration of the anemometer leads to expected instrumental errors in wind speed of about 0.1 m s\(^{-1}\). Differential rates of cup acceleration and deceleration in variable wind regimes can lead to overspeeding which biases estimates of mean wind speed high by 3–10% (e.g., Hayashi 1987). This appears not to be a problem for the type of cups we use in the equatorial Pacific where wind speeds are typically <10 m s\(^{-1}\). Freitag et al. (1989) performed a 115-day field intercomparison of the VAWR Climet cup and vane with an R. M. Young model 05103 propeller and vane mounted on surface toroids separated by 11 km near 0°, 140°W. Mean cup determined speeds were only 0.02 m s\(^{-1}\) larger (0.3% of the mean) which was not significantly different from zero. Moreover, 2-hourly time series of speed, direction and vector wind components were correlated at 0.97 or better. Halpern (1987b) has noted from field experiments that VAWR wind speeds estimated from a surface toroid tend to be higher than those estimated from a more stable spar buoy. However the relative difference is only about 2% for 8-hour averages (e.g., 0.1 m s\(^{-1}\) in 5 m s\(^{-1}\) flow).

SST is measured 1 m below the surface using either a Yellow Springs Instrument (YSI) model 44032 temperature sensor interfaced to the VAWR (calibrated accuracy of 0.01°C); or a YSI model 44204 temperatures sensor attached to a Telonics ARGO transceiver (calibrated accuracy of 0.05°C). Additional information on the processing of data from equatorial current meter moorings can be found in Freitag et al. (1987).

The 2°N and 2°S moored data were collected with ATLAS (Autonomous Temperature Line Acquisition System) thermistor chains. ATLAS (Milburn and McLain 1986) is a taut wire surface mooring that measures winds, air temperature, SST (i.e., temperature from 1 m depth), and 10 subsurface temperatures to a maximum depth of 500 m. Thermistor resistance is digitized in each measurement pod and this data is time-multiplexed to the surface electronics package using three conductor double armored cable. Thermistors are calibrated prior to deployments to an accuracy of about 0.005°C; in situ comparisons with nearby CTD casts indicate a long-term accuracy of better than 0.1°C. Data are telemetered to shore via Service ARGOS as 2-hour averages (or in some cases 1-hour averages). Normally five unique data transmissions are received each day. The basic time series is taken to be daily averages of these data.

Sea level data are from the TOGA island sea level stations of Nauru (0°32'S, 166°54'E) and Kapingamarangi (1°00'N, 154°50'E). These data have been detided and averaged to daily values. The instrumental accuracy of tide gauge measurements is usually quoted in terms of a monthly mean uncertainty of about 0.1 cm (e.g., Wyrski and Leslie 1980). However, McPhaden et al. (1988) have found that sea level at both island stations can be biased upward by as much as 30 cm during periods or pronounced westerly winds as for example occurred during May 1986. Sea level data have not been corrected for the inverted barometer effect which is negligible in the equatorial Pacific (Enfield 1987).

Wind data are available from Nauru from an R. M. Young model 05103 propeller and vane anemometer mounted a tower 10 m above the ground. Anemometers are replaced every 6 months to 1 year and are calibrated prior to deployment to within 0.2 m s\(^{-1}\). Data are vector averaged for 40 minutes of each hour, then three individual hourly samples are transmitted to shore via GOES geostationary satellite. Data are processed to daily means for this study.

Nauru winds tend to underestimate the amplitude of variations at the 0°, 165°E mooring site, especially during periods of westerlies. The wind sensor is located on the northeast side of the island in the lee of a 40-m-high hill during periods of westerlies. Nonetheless, there is a high correlation between daily time series at the two locations (0.88 for zonal winds and 0.85 for meridional winds for the period December 1986–October 1987) so that Nauru winds can be used as an index for winds at the mooring site.

CTD data were collected along the four equatorial transects indicated in Fig. 1 during 1986–1988 as part of the US/PRC bilateral air–sea interaction program. Stations were occupied approximately every 2.5° of longitude except during the January–February 1986 section which had approximately 5° spacing. In each case, casts were made to at least 1000 db and processed to 2-db resolution. Mangum et al. (1989) describe the acquisition and processing of these data in greater detail.

Figure 4 shows time series of temperature along the equator in the upper 500 m from January 1986 to May 1988. Data from 2°N are also presented for the period July–December 1986 when no equatorial data were available. These measurements have been used to generate continuous 2½-year daily time series of SST, 20°C isotherm depth and surface dynamic height relative to 500 db. The time series were then smoothed with an
11-day Hanning filter (Bendat and Piersol 1971) to highlight variability on time scales of a week and longer (Fig. 5). Details about data interpolation, extrapolation, compositing and use of a mean T–S for dynamic height in generating these time series are discussed in the Appendix.

3. Winds

Figures 5a and 5b show zonal and meridional wind variations at 0°, 165°E and at Nauru. Emphasis in the following discussion is given to the zonal wind component because monthly mean meridional winds seldom deviate from climatology by more than 2 m s⁻¹ and 1986–1987 annual means are within 0.3 m s⁻¹ of climatology (Table 1). Monthly mean zonal winds on the other hand deviate from climatology by up to 6 m s⁻¹ as is evident from Fig. 5a. Also, the yearly averaged zonal wind component measured at 0°, 165°E during 1987 was 0.8 m s⁻¹ to the east compared to a climatological mean of 1.6 m s⁻¹ to the west, implying an interannual anomaly of 2.4 m s⁻¹ (Table 1).

Easterly winds were stronger than normal by 2–3 m s⁻¹ for the first 4 months of 1986. In May, a 10-day westerly wind burst triggered an equatorial Kelvin wave which propagated into the eastern Pacific at speeds near 3 m s⁻¹ (McPhaden et al. 1988; Miller et al. 1988).
At the time it was thought that this wind burst might be a trigger for the El Niño (Kerr 1987). However, subsequent analysis (McPhaden et al. 1988) indicated that the ensuing Kelvin pulse had little discernable effect on SST in the eastern Pacific. More persistent westerly anomalies appeared at the end of June 1986 and were followed in August and September by persistent 1–2°C SST anomalies in the eastern Pacific that marked the beginning of the El Niño (Kousky and Leetmaa 1989).

The strongest westerly winds that were measured at 0°, 165°E during the El Niño occurred in November and December of 1986, with peak daily averaged zonal wind speeds of 10 m s⁻¹. These westerlies were associated with a local 50-m elevation of 20°C isotherm depth and 20-cm drop in dynamic height (Figs. 5d, e).
They also excited an equatorial Kelvin wave train, which propagated into the eastern Pacific, depressing sea level there by 20 cm (e.g., Miller et al. 1988). After a brief period of stronger than normal easterlies in January 1987, anomalous 10-day to 2-month outbreaks of westerly winds were observed from February to October with peak speeds of 2–7 m s\(^{-1}\). November and December 1987 were much closer to climatology than the same period in 1986, and by March–April 1988 easterly anomalies of 2–3 m s\(^{-1}\) had developed.

4. Temperature

a. Temporal variations

Temporal variations in near equatorial SST at 165°E are summarized in Fig. 5c. Daily SST ranged between 28.5°C and nearly 31°C and temperatures were often warmer than climatology by 1°C. Periods during which SST was near to or slightly cooler than climatology (March–April and December 1986; July–September 1987; April 1988) were associated with month-long episodes of high easterly or westerly wind speeds. The warmest temperatures in both 1986 and 1987 occurred in October–November. The maximum monthly mean temperature in November 1987 was 30.4°C, which was reached as the core of the warm pool (>30°C) migrated westward across the moored array (Fig. 2).

The 20°C isotherm (Fig. 5d) was near its climatological mean depth in January and February 1986, after which it deepened by 40 m in April following a period of stronger than normal easterlies. It remained deeper than climatology until November and December 1986 when it shoaled rapidly by 50 m during a westerly wind event. The 20°C isotherm then remained shallower than normal by about 20–40 m for over a year. Not until April 1988 following, and during, a period of enhanced easterlies did it return to its normal depth. Thus, the transitions from a deep to a shallow thermocline and vice versa occurred on time scales short compared to the duration of the El Niño event, and were associated with brief periods of anomalous forcing.

It is interesting to note that the annual range in 20°C depth based on XBT data is only 10 m, similar to Meyers' (1979) result for the 14°C isotherm. The annual range of SST is only 0.3°C based on Reynolds (1988). Month-to-month variations evident in the time series shown in Fig. 5 are generally larger than these ranges. Also, yearly averaged anomalies inferred from Table 1 are comparable to or larger than these ranges.

SSST variations do not appear to be consistently related to thermocline variations (as indicated by 20°C). For example, relatively cool surface temperatures during July–September 1987 are associated with a relatively shallow thermocline. However, similarly cool SST is found in March–April 1986 when the thermocline is 50 m deeper. Table 1 also shows that yearly averaged SSTs were nearly identical in 1986 and 1987, whereas the yearly averaged 20°C isotherm was 35 m shallower in 1987 than 1986. A coherence spectrum shows the relationship between these two variables more quantitatively (Fig. 6a). Except for peaks at periods near 3–4 days and 30–40 days, there is no statistically significant coherence between SST and 20°C. We note in particular that the range of periods below 100 days, which includes annual and interannual fluctuations, is incoherent. Moreover, phase in the 30–40 day band indicates that cold SST precedes shallow 20°C isotherm depth by one-quarter of a cycle.

These results suggest that entrainment from the thermocline is generally not a dominant process affecting SST. Meyers et al. (1986) arrived at a similar conclusion from a surface layer heat budget analysis using XBT data west of the dateline during the 1982–83 ENSO. This may in part be due to the fact that entrainment cooling is inversely proportional to the depth of the surface layer (McPhaden 1982a) which was 50–100 m deep during 1986–88 (Fig. 4). This is at least as deep as the 50-m surface layer in the central equatorial Indian Ocean where comparable wind speeds are ineffective at changing SST via entrainment (McPhaden 1982a). Moreover, in the western Pacific the surface layer is not always well mixed with regard to temperature (e.g., Fig. 11c) or salinity (Mangum et al. 1989) which would further insulate SST from thermocline variability.

b. Meridional temperature sections

Simultaneous mooring data are available for 14 months between December 1986 and May 1988 at 2°S, 0° and 2°N during 1986–88, from which it is possible to construct monthly mean meridional temperature sections. These sections have a coarser meridional and vertical resolution than those based on hydrographic data presented in Toole et al. (1989). However, they are less affected by temporal aliasing due to short period
fluctuations, and they are more numerous than hydrographic sections taken during the same time period.

Figure 7 illustrates the meridional structure for December 1986, March and October 1987 and March 1988. Except for October 1987, these sections show significant weakening of the thermocline near the equator relative to 2°N and 2°S, indicating the presence of a geostrophically balanced Undercurrent (Delcroix et al. 1987; Picaut et al. 1989). As will be discussed below, the Undercurrent was absent from the thermocline in October 1987 (section 6). Also, the December 1986 section shows a pinching of isotherms between 100-150 m which corresponds to a region of eastward geostrophic flow near the surface and westward flow in the thermocline above the Undercurrent (Picaut et al. 1989; Fig. 11a below).

The December 1986, March 1987 and October 1987 plots show very little meridional SST gradient between 2°N and 2°S along 165°E (see also Fig. 1). This was characteristic of all months (8) from December 1986–December 1987 for which moored SST data were available at all three sites. The difference between 2° and the equator generally did not exceed 0.3°C; moreover, the only month during this period to exhibit an equatorial SST minimum was October 1987 when the equator was warmer than 2°S by 0.2°C and cooler than at 2°N by 0.1°C. The lack of a significant equatorial SST minimum during late 1986 and 1987 is consistent with the predominance of westerly winds which are not upwelling favorable. Indeed, strong westerlies in December 1986 appear to have led to significant downwelling along 165°E as indicated by the depression of the 26°–28.5°C isotherms at the equator relative to 2°N and 2°S.

The December 1986, March 1987 and October 1987 sections can be contrasted with the March 1988 section during which a very pronounced equatorial SST minimum is observed. Nauru zonal winds were upwelling favorable in March 1988 at nearly 5 m s⁻¹ from the east, over 3 m s⁻¹ stronger than climatology. Compared to the previous March, equatorial temperatures were 0.5°C cooler and the 24°–28°C isotherms tended to dome more near the equator. This strong upwelling diminished in April and May 1988 as the easterlies diminished in intensity.

5. Surface height

Surface dynamic height variations are shown in Fig. 5e. Surface height undergoes a slow 10 dyn cm decrease punctuated by 5–10 cm peak-to-peak monthly fluctuations throughout much of 1986. Towards the end of 1986, surface height drops abruptly by 20 cm to a minimum in mid-January 1987 in association with the strong westerly wind event shown in Fig. 5a. It rebounds by 10 cm during February, then undergoes another slow 10 cm decrease beginning in April. The minimum surface height of the ENSO event is reached in late August, after which there is a rise of 10–15 cm by the end of 1987. In early 1988, surface height fluctuates with 10 cm peak-to-peak monthly oscillations, but is on average about 10–15 cm lower than the corresponding pre-ENSO period in 1986. The annual mean dynamic height relative to 500 db in 1987 was

![Fig. 6. Coherence and phase between daily values of 20°C isotherm depth and (a) SST and (b) dynamic height 0/500 db from moored time series data at 0°–2°N. In phase relationship implies shallow 20°C isotherm corresponds to cold SST in (a) and low dynamic height in (b); 95% confidence limits for rejection of the null hypothesis based on 5 and 15 frequency band averages are indicated.](image-url)
1.31 dyn m, 0.14 dyn m lower than in 1986 and 0.09 dyn m lower than climatology (Table 1).

Qualitatively, changes in surface dynamic height (Fig. 5e) tend to mirror changes in thermocline depth as measured by the 20°C isotherm (Fig. 5d). This suggests that the western Pacific is behaving like a “two-layer” fluid (Wyrtki 1985; Rebert et al. 1985). The two-layer analogy is not perfect, however. A weakening of the upper thermocline in mid-1987 indicates that more than one internal mode of variability is important during the ENSO at this location. Also, although by May 1988 the 20°C isotherm had returned to within a few meters of its normal depth, dynamic height was 10–15 cm lower than in early 1986 as noted above.

The two-layer approximation is quantified in Fig. 6b, which shows nearly in-phase coherence (≥0.7) between surface height and 20°C isotherm depth over a broad range of periods between 2–12 days and >30 days. The coherence is marginally insignificant with 95% confidence at periods longer than 60 days. However, with broader frequency band averaging the coherence rises above 95% confidence limits because of the stability of the phase and coherence levels in the lowest frequency bands. Least squares linear regression analysis of daily data indicates that the ratio of dynamic height to 20°C isotherm depth variations is about 3 × 10⁻³. This is comparable to the value of 5 × 10⁻³ that Wyrtki (1985) used to convert sea level to isotherm displacement in his mass budget study of the upper tropical Pacific.

To examine the representativeness of the dynamic height time series, sea level at Kapingamarangi and Nauru is shown in Fig. 8. For daily data, correlation of dynamic height with Nauru sea level over the 849-day span in Fig. 8 is 0.59 as compared to 0.76 with Kapingamarangi sea level. Nauru is closer than Kapingamarangi to the mooring site, but is biased during periods of strong westerly winds because the tide gauge is located in a small confined harbor on the west side of the island (McPhaden et al. 1988). This leads to large 10–30 cm spikes in sea level due to westerly wind setup, most evident in January, May, November–December 1986 and July 1987. Kapingamarangi is less affected by local setup, though spikes still may be seen in May and November 1986 and July 1987.

Variations in sea level at Kapingamarangi tend to be larger in amplitude than at 165°E and there appears to be a higher frequency component in the sea level time series. However, the month-to-month changes agree reasonably well in phase at the two sites. Both show a slow decrease in surface height in 1986 which is accentuated by a 20 dyn cm (30 cm) drop in dynamic height (sea level) in November–December; also, a broad minimum occurs in August–September 1987 after which the surface height begins to recover. Thus, the dynamic height time series at 165°E is representative of a region extending at least 10° farther to the west. Indeed, the depression in surface height seen in 1987 between 155°–165°E along the equator is evident to a greater or lesser extent to the western boundary between 20°N and 20°S (Climate Analysis Center 1987).

6. Currents

a. Zonal velocity variations

Figures 9 and 10 show time series of currents recorded at the equator. Line plots are presented in Fig. 9 for the period January–July 1986. Only four depths were instrumented at this time, one of which (200 m) was in the thermocline. The data are presented in Fig. 10 as a contour plot for December 1986–May 1988 when 5–6 levels were instrumented with a vertical resolution of 50 m. Wind stress is also plotted for each time period in Figs. 9 and 10. Stress has been calculated assuming a drag coefficient of 1.2 × 10⁻³ and a neutral stability boundary layer to correct 4-m winds to 10-m anemometer heights.

Three major zonal currents can be identified in Figs. 9a and 10a. The westward flowing South Equatorial Current is found in the upper 100 m during periods of easterly winds. The eastward Equatorial Undercurrent...
is found in the upper thermocline with core speeds typically between 150 and 200 m. The Equatorial Intermediate Current (Hisard et al. 1970; Delcroix and Henin 1988) is situated below the Undercurrent and flows westward at 300 m with typical speeds of about 25 cm s\(^{-1}\). Its variations tend to be less dramatic than those in the South Equatorial Current and Equatorial Undercurrent which are the focus of the following discussion.

Figure 9a shows that the South Equatorial Current is present in upper 100 m (coincident with the surface warm pool) from January to April and again in June 1986. It is strongest in late March and early April (about 75 cm s\(^{-1}\)) following a month of unusually strong easterlies. In mid-May, currents in the upper 100 m accelerated rapidly towards the east in response to a 10-day westerly burst, peaking at over 100 cm s\(^{-1}\) at 10 and 50 m four days after the maximum westerly winds (McPhaden et al. 1988). Westward flow was later reestablished in the upper 100 m at speeds \(\geq 25\) cm s\(^{-1}\) in June. The record length mean 10-m current in Fig. 9a is 14 cm s\(^{-1}\) to the west, comparable to the average surface current based on monthly mean ship drift data (Meehl 1982). The ship drift climatology however suggests that flow is westward throughout the year at 165°E.

Measurements at 200 m are located in the thermocline near the speed core of the Equatorial Undercurrent (Delcroix et al. 1987). The average current at 200 m is 60 cm s\(^{-1}\) for January–July 1986, with relatively little variation compared to the surface. In particular, 1 m s\(^{-1}\) speed changes like those observed in the surface layer are not observed at 200 m during the May wind burst. This is consistent with the observations of Hisard et al. (1970) who found little change in the undercurrent speed in response to a westerly wind burst lasting 8 days in April 1967.

Figure 10a shows contours of zonal velocity for December 1986–May 1988. No data were available at 10 m during this time period because of instrumental problems. However, the correlation between 10 and 50 m zonal currents for the 164-day record in Fig. 9a is 0.87. The 164-day average 10-m currents were also 12 cm s\(^{-1}\) stronger to the west than the 50-m currents, consistent with the fact that the mean winds were easterly over the same time period. Thus, we will interpret the 50-m record in Fig. 10a as representative of flow closer to the surface, with the qualifier that currents are probably vertically sheared in the direction of the surface winds.

The response of the zonal currents to the strong westerlies at the end of 1986 was dramatic. Eastward currents at 50–100 m exceeded 1 m s\(^{-1}\) and a strong positive shear stress penetrated the thermocline to 150-m depth (cf. Figs. 11a and 11c). Pinched isotherms in the upper thermocline at the equator (Fig. 7a) indicate the geostrophic signature of this jet (Picaut et al. 1989). At the same time, the Undercurrent at 200 m weakened to speeds of 30 cm s\(^{-1}\). For comparison, during the previous winter season (January–February 1986) when winds were on average easterly, flow in the upper 100 m was 30–50 cm s\(^{-1}\) to the west and the undercurrent flowed at >50 cm s\(^{-1}\) to the east (Fig. 11). The difference in zonal flow in the upper 100 m for these two periods separated by approximately one year is about 150 cm s\(^{-1}\).

The surface winds reverted to easterly in early January 1987 and within a week currents in the upper 100 m switched to westward. For the remainder of January through March 1987, the South Equatorial Current flowed to the west at about 40 cm s\(^{-1}\) and the Under-

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**Fig. 8.** Dynamic height time series as in Fig. 3, plus sea level at Nauru and Kapingamarangi. Daily time series have been smoothed with an 11-day Hanning filter.

**Fig. 9.** Daily time series of (a) zonal velocity and wind stress and (b) meridional velocity and wind stress at 0°, 165°E for January 20–July 2, 1986. Nauru wind stress (dotted line) is also plotted.
current flowed to the east at about 50 cm s$^{-1}$. As westerly wind anomalies developed in April and May, the currents at 50 m and 100 m reversed and became eastward at speeds of 40–60 cm s$^{-1}$. Continued intensification of the westerly wind anomalies in June and July was followed by a gradual reduction and reversal of flow in the undercurrent, first at 250 m in late July, then at 200 and 150 m in mid-September. The strongest eastward flow during August-early October was at 100 m in the upper thermocline (Fig. 11a), but this too reversed by mid-October. For 3–4 weeks in October and early November 1987, flow between 50 m and 250–300 m was westward, with peak daily speeds greater than 40 cm s$^{-1}$ in the thermocline. We note that westward migration of surface water warmer than 30°C shown in Fig. 2 coincided with the appearance of westward flow in the upper 200 m in September–November 1987.

The Undercurrent was reestablished in early November nearly simultaneously between 150–250 m. However, until April 1988 it seldom reached speeds as high as of 40 cm s$^{-1}$. From January to March 1988 for example, the EUC was weaker than in 1986 by about 30 cm s$^{-1}$ and weaker than in 1987 by about 20 cm s$^{-1}$. Then, as the easterlies strengthened in March–May 1988, the Undercurrent accelerated to a maximum daily speed of 98 cm s$^{-1}$, which was the highest Undercurrent speed observed over the record length (see also Fig. 11a). At the same time flow in the thermocline was accelerating, the South Equatorial Current, which was well developed in February–March, began to decrease in strength at 50 and 100 m. This inverse relationship between Undercurrent and South Equatorial Current speeds is similar to that predicted by nonlinear theories of the undercurrent (e.g., Charney 1960; Philander and Pacanowski 1980). These
Theories indicate that strong easterlies drive a vigorous meridional circulation that 1) advects eastward flow from the thermocline toward the surface and 2) advects westward momentum put into the ocean by the surface winds to higher latitudes. The net effect is to reduce near-surface westward flow at the equator and to amplify and vertically stretch the undercurrent. Wind-driven upwelling during spring 1988 consistent with these arguments has already been noted in the discussion of Fig. 7.

b. Meridional velocity variations

Figures 9b and 10b show meridional current variations for 1986–88. Meridional currents are dominated by oscillations with periods < 1 month and peak-to-peak amplitudes of O(10 cm s⁻¹). Two- to four-week average profiles of meridional velocity (Fig. 11b) show weaker currents in general than in the zonal direction and vertical structure that is not easily interpreted.

Weak meridional temperature gradients in 1987 of 0.1–0.2°C per 100 km (Fig. 7), combined with relatively short period meridional velocity fluctuations suggest sea surface temperature changes of O(0.1°C) due to meridional advection. Thus the large (i.e., >1°C) changes in SST in Fig. 5a during 1987 were probably due to other processes. Also, the appearance of westward flow in the thermocline above 250 m during September–November 1987 was most likely not due to a meridional displacement of the undercurrent.

Meridional currents change sign 2–3 times during September–November without changing the sign of zonal velocity.

c. Spectral characteristics

Zonal and meridional velocity variability is summarized in the spectra plotted in Fig. 12. Four depths (50 m, 100 m, 150 m and 250 m) are shown for illustrative purposes. Spectral levels at periods shorter than about 10 days are comparable in both velocity components and generally decrease with increasing frequency up to the Nyquist period of 2 days. There is evidence for enhanced meridional velocity variability near 4 days at 50 m and 6–7 days at 100 m which may be related to inertia–gravity wave energy (e.g., Wunsch and Gill 1976). However, the most striking differences in spectral energy densities between zonal and meridional velocity occur at periods longer than 10 days. Meridional velocity has a spectral peak at periods between about 10–20 days (at 50 m), 10–30 days (100 m) and 20–30 days (150 m) where it tends to be more energetic than zonal velocity. Conversely, at periods longer than about 30 days, the meridional velocity spectra tend to flatten whereas the zonal velocity spectra continue to rise. Spectral levels in the two velocity components are separated by 1 to 2 orders of magnitude in lowest frequency band which corresponds to periods of 104 to 522 days. There is also a tendency for energy...
levels to decrease from the surface layer (50–100 m) to the thermocline (150–250 m). The most significant decreases occur at periods longer than 10 days in zonal velocity, e.g., a factor of 10 decrease in the lowest frequency bands between 50 and 250 m. In meridional velocity on the other hand, the most pronounced decrease in energy is at periods between 10–30 days.

d. Coherence with local winds

Spectra of zonal and meridional wind stress at the equatorial mooring site are shown in Fig. 13 for a 10-month period from 13 December 1986 to 14 October 1987. Daily moored wind time series are gappy before and after this period and are therefore excluded in the calculation. The two stress components have comparable magnitudes and decrease in energy with increasing frequency at periods shorter than 10 days. At periods longer than 10 days, the meridional wind stress spectrum tends to flatten whereas the zonal wind stress spectrum continues to rise. In the lowest period band (periods of 61 days to 304 days), there is five times more energy in the zonal component than in the meridional component.

Coherence and phase between wind stress calculated from moored measurements and velocity at two depths (50 m in the velocity surface layer and 150 m in the thermocline) are shown in Fig. 14. Fluctuating zonal currents at 50 m are coherent with the zonal winds at periods near 3 days and 10 days and at periods longer than 20 days (Fig. 14a). The 3-day peak may be related to inertia–gravity wave resonances whereas the 10-day coherence peak may be related to short duration westerly wind bursts. The high coherence (≥0.8) for periods longer than 20 days is remarkable but not unexpected based on visual inspection of the time series in Fig. 10a. The phase relationship for coherent fluctuations at periods 10 days and longer is such that winds lead currents, consistent with the atmosphere forcing the ocean. The implied lag for the oceanic response is 2–3 days at 10-day periods and about one week in the lowest period band centered near 100 days.

Local zonal wind stress variations are much less coherent with zonal current variations in the thermocline than in the surface layer (Fig. 14b). At 150 m, in fact, there is no significant coherence between the two variables in any resolved frequency band. This does not necessarily imply that local wind forcing is unimportant in the dynamics of the Undercurrent. Instead, it suggests that remotely generated, wind-forced equatorial waves account for a large percentage of its variability at periods shorter than 1 year. The high spectral ratio of zonal and meridional velocity in Fig. 12 is consistent with this inference, since for low frequency equatorial Kelvin and Rossby waves, zonal current variations are much larger than meridional current variations (Eriksen 1980; McPhaden 1982b).

Meridional winds are significantly coherent with 50-m meridional currents near 10-day periods, with winds leading by one-eighth of a cycle (Fig. 14c). Coherence is marginally insignificant, but peaked near 10-day periods at 150 m as well. Hayes et al. (1989), using the same data, have shown that meridional velocity fluctuations near 10-day periods are vertically coherent between 50–250 m and associated with an antisymmetric dynamic height fluctuation across 2°N–2°S. They interpreted this variability in terms of first vertical mode, wind-forced mixed Rossby–gravity waves. Note that the frequency band over which coherence with the local winds is significant is narrower than the meridional velocity spectral peaks between 10–30 days shown in Fig. 12. This could be due to the importance of remote wind forcing at 10–30-day periods or possibly to nonstationarity of the 10-day mixed Rossby–gravity wave (Hayes et al. 1989). It is interesting to note though that the zonal and meridional velocity spectra shown in Fig. 12 are qualitatively similar to those in the upper few hundred meters of the eastern equatorial Pacific (Philander et al. 1985) where meridional velocity variations are dominated by 20–30 day barotropic instabilities of the large-scale zonal flow field. A similar oscillation may also contribute to relatively high energy levels at periods of 20–30 days in the western Pacific if the meridional shear of the large scale zonal currents is sufficiently intense to be unstable.

7. Zonal pressure gradient and the Undercurrent

Theory indicates that the Equatorial Undercurrent is driven by an eastward baroclinic zonal pressure gradient force in the thermocline and that the magnitude of the pressure gradient is determined by the zonal wind stress (McCreary 1981; McPhaden 1981). In this section we compare zonal currents from the equatorial mooring (Fig. 11) with zonal pressure gradient determined along the equatorial CTD sections. Figure 15 shows surface dynamic height relative to 1000 db along the 4 CTD transects which extend from the mooring site at 165°E westward to 141°E (except for the January–February 1986 section which spans 175°–141°E). These surface dynamic heights, and dynamic heights at 25 db intervals, were least squares fit to a mean and a trend. The trend, expressed as a pressure gradient in N m⁻³, is plotted as a function of depth in Fig. 16.

The January–February 1986 CTD section was occupied prior to the ENSO. Shipboard winds measured on this section were easterly east of 150°E and westerly west of 150°E. The surface height sloped upwards in the direction of the wind stress, but geopotential surfaces in the thermocline sloped more uniformly upward to the west. The May 1988 CTD section was occupied after the ENSO when easterlies prevailed along the entire transect. Geopotential surfaces sloped upward to the west more sharply at the surface and in the thermocline than in January–February 1986. The Undercurrent was also stronger in May 1988 with a maximum
speed greater than 90 cm s\(^{-1}\). The December 1986 and October 1987 CTD sections were made during the El Niño. Intense westerly winds prevailed on the first of these, and the surface sloped upward toward the east by about 10 dyn cm from 141\(^\circ\) to 160\(^\circ\)E. Significant positive pressure gradients were confined to the upper 100 m, while at 200 m near the core of the Undercurrent the zonal pressure gradient was near zero. The undercurrent flowed eastward with a reduced speed of about 30 cm s\(^{-1}\) at this time and may have been supported at 165\(^\circ\)E by a negative pressure gradient to the east of 160\(^\circ\)E (Fig. 15). Westerly winds and pressure gradients in the upper 100 m were weaker in October 1987 than during the previous December. However, the October measurements are distinguished by a significant positive pressure gradient in the thermocline between 150–200 m, and the disappearance of the Undercurrent (Fig. 11a).

Thus, an eastward pressure gradient force in the thermocline prior to and after the El Niño was associated with a well-developed Undercurrent, whereas during the El Niño when this pressure gradient reversed, the Undercurrent disappeared. Analysis of the mean seasonal cycle in zonal pressure gradient from historical hydrographic data in the western Pacific (Mangum et al. 1989) suggests that the adjustment time of the Undercurrent to fluctuating winds is longer than 3 months. The zonal pressure gradient force across
133°–162°E is eastward between 100–200 m throughout the year, including in December–February during the northwest monsoon. This explains why, prior to our measurements under El Niño conditions, the Undercurrent had been observed during all seasons in the western Pacific east of about 140°E (Montgomery 1962; Lindstrom et al. 1987). From this we would conclude that the episode of strong westerly winds in November–December 1986 was too short to cause a reversal of the Undercurrent. In contrast, the disappearance of the Undercurrent at all depths in the thermocline in October 1987 presumably resulted from the cumulative effect of nearly a year of prevailing westerly winds in the western Pacific.

The precise role of equatorial wave transients in reversing the Undercurrent in October 1987 is unclear from our data. However, flow in the thermocline is not in equilibrium with the local winds which were near zero on average in October 1987 when the Undercurrent disappeared (cf. Fig. 5 and Fig. 10a). Also, westerly winds prevailed for most of December 1986–September 1987 at a time when an eastward flowing Undercurrent was found in the thermocline. Mean zonal winds for this time period were westerly at 1.3 m s⁻¹ compared to easterly at 1.8 m s⁻¹ based on Wyrtki and Meyers’ (1975) climatology. Thus, a complete dynamical description of the current variations observed along 165°E during 1986–87 requires information on wind, current and temperature variations across the basin.

8. Discussion and conclusions

We have described wind, temperature, surface height, zonal pressure gradient and current variability in the western equatorial Pacific during the 1986–87 ENSO event using time series and CTD data. Our description has spanned daily to interannual time scales in the ocean and the atmosphere. Major results are summarized and discussed below.

Zonal winds were characterized not only by year-to-year changes > 2 m s⁻¹, but also by 10-day to 2-month westerly wind episodes with peak amplitudes approaching 10 m s⁻¹. Locally, these wind fluctuations had a dramatic effect on ocean currents, temperatures and dynamic heights, particularly in the upper 100 m. In this surface layer, for example, zonal currents responded to wind variations, typically within a week or less, and sometimes reached speeds > 100 cm s⁻¹ as occurred during episodes of strong westerly winds in May and November–December 1986. Variations in the thermocline were less coherent with the local winds; they were also generally of smaller amplitude, though notably the Equatorial Undercurrent disappeared for 3–4 weeks in October–November 1987 when the normal eastward directed zonal pressure gradient force in the thermocline reversed. Thus, the Undercurrent is not a permanent feature of the general circulation in the western Pacific as had been previously assumed (Philander 1973). The Undercurrent also disappeared at 159°W in September 1982 (Firing et al. 1983) and January 1983 at 110° and 95°W (Halpern 1987c) during the 1982–83 ENSO. In both instances the normal sea level slope along the equator and presumably of the eastward pressure gradient force in the thermocline reversed. As with our observations, the disappearance of the Undercurrent during the 1982–83 ENSO lasted several weeks and could not be accounted for in terms of local equilibrium dynamics.

Our data suggest that zonal current variations along the equator are important in the mass balance of the western tropical Pacific. For example, the 100 cm s⁻¹ eastward jet in the upper 100 m in late 1986 (Fig. 11a), if representative of flow within 2° of the equator, implies 40 Sv of mass flux across 165°E. The 50 m rise in the 20°C isotherm depth and 20 dyn cm drop in dynamic height from late November 1986 to January 1987, if representative of variations within 2° of the equator and 25° to the west of the mooring site, could supply about 10 Sv of this mass flux. The difference of 30 Sv could come from westerly wind driven meridional Ekman convergence: 0.07 N m⁻² eastward stress acting over 25° of longitude implies about 15 Sv equatorward at both 5°N and 5°S. Note that Ekman convergence is a local response to westerly wind forcing that would tend to depress the thermocline. Evidence for this can be seen in Fig. 7 which shows pinched isotherms between 100–150 m at the equator in December 1986. However, superimposed on this local response is a net rise in thermocline depth and drop in dynamic height (Fig. 5) consistent with a nonlocal equatorial Rossby wave excitation by large scale transient westerly wind forcing as occurs in simple models of El Niño (e.g., McCreary 1977).

Similar arguments pertain to the period of sustained westerly wind forcing and eastward near surface flow.
in May–August 1987. An eastward flow of 25 cm s$^{-1}$ in the upper 100 m within 2° of the equator implies a transport of 10 Sv. About 25% of this would account for the 10 dyn cm drop in dynamic height between May and August when dynamic height reached its lowest values of the event. Westerly wind-driven Ekman convergence could supply additional mass to feed the eastward jet. Note also that sea level rose from September to November 1987 coincident with the appearance of westward flow at Undercurrent depths in the thermocline; it is possible that mass convergence across 165°E was at least in part responsible for this rise.

A variety of processes are involved in controlling SST variability in the western Pacific warm pool. High wind speeds, which in principle favor evaporative cooling and entrainment from the thermocline, corresponded to falling SST. Entrainment may also be enhanced during periods of upwelling favorable easterlies (e.g., as in March 1988). We infer, however, that in general entrainment is weak since the depth of the thermocline is poorly correlated with SST. This is in contrast to the eastern Pacific during 1986–88 where along the equator on interannual time scales a deep (shallow) thermocline corresponds to warm (cool) SST (McPhaden and Hayes 1990). The difference is due
to the fact that the thermocline is on average deeper and the surface layer thicker in the western Pacific. Surface temperatures are buffered by the large thermal capacity of the warm pool so that entrainment mixing does not readily translate into SST signals (Meyers et al. 1986). Lukas and Lindstrom (1987) have noted that shallow haloclines in the surface layer may further insulate SST from thermocline variations. Evaporative cooling on the other hand may be important in affecting SST, an issue which is examined in greater detail in McPhaden (1990).

Meridional advection was probably not responsible for the largest O (1°C) SST changes observed during the ENSO at 0°, 165°E; meridional temperature gradients near the mooring site were weak, and O (10 cm s⁻¹) velocities at dominant 10–30 day periods would lead to changes in SST ≪ 1°C. On the other hand, intense eastward flow may have been responsible in part for the eastward displacement of the near equatorial SST maximum towards the dateline in late 1986. Also, zonal advection is implicated in the westward migration of water warmer than 30°C from the date line during September–November 1987 (Fig. 2). For part of this time at 0°, 165°E, rising SST was associated with westward flow near the surface (and coincidently with westward flow in the thermocline when the Undercurrent disappeared). As the warmest waters retreated westward from near the dateline, anomalous atmospheric heating diminished, the tradewinds strengthened and the ocean–atmosphere system returned to near-normal conditions (Kousky and Leetmaa 1989).

The 1986–87 ENSO event evolved differently than the Rasmusson and Carpenter (1982) composite (Kousky and Leetmaa 1989). For example, warm SST anomalies in the central Pacific appeared first in August–October of 1986 before (rather than after) South American coastal warming; and the event lasted about 18 months rather than 12 months as in the composite. Also, persistent westerly anomalies first appeared in July–August 1986 rather than during the November–January season. Differences from the composite, as also occurred for example during the 1982–83 ENSO (Cane, 1983), are not unexpected since Rasmusson and Carpenter based their analysis on only six events from 1950 to 1973. Within this ensemble are significant event-to-event variations. Hence, as with SST and winds, current and subsurface temperature variations observed in the western equatorial Pacific during 1986–87 may not be representative of other ENSO events.

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APPENDIX

In this section we discuss details of data interpolation, extrapolation, compositing and use of a mean T-S relationship in generating the time series in Fig. 5.

When SST data at the current meter mooring site were unavailable, we substituted 10-m temperatures from the equatorial mooring (19 January–4 July 1986) or SST from the 2ºN mooring (5 July–11 December 1986). For the 214-day period 13 December 1986–16 July 1987, correlation between 10-m temperature and SST was 0.85 and the mean difference was 0.17ºC (SST warmer). For 267 days in 1987 when SST data were available from both 2ºN and the equator, correlation of SST at the 2 sites was 0.91 and the mean difference was 0.12ºC (2ºN colder).

Figure 4 shows the depth of the 20ºC isotherm is representative of variations in the middle of thermocline. The thermocline tends to be stronger at 2ºN than at the equator on average. However, correlation of overlapping data in 1987 (which for 20ºC isotherm depths was 0.53 for 265 daily averages and 0.62 for 11-day Hanning filtered data) indicates that variations are coherent over 2º of latitude.

The temperature data in Fig. 4, combined with salinity from a mean temperature–salinity (T-S) profile, were used to calculate surface dynamic height relative to 500 db (Fig. 5e). Wyrski (1975) noted that the 500 db surface relative to 1000 db is flat to within ±1 dynamic cm in the equatorial Pacific. A similar result was found by Delcroix et al. (1987) for 400/1000 db dynamic height variations within a few degrees of the equator along 165ºE from SURTROPAC cruises. Using all the available CTD data between 2ºN and 2ºS along 165ºE for January 1984–May 1988 (100 casts from the US/PRC and SURTROPAC programs), we found that the standard deviation of dynamic height 0/500 db was 7.3 dyn cm compared to a 500/1000 db standard deviation of only 0.9 dyn cm. Thus, 500 db is a reasonable reference level for discussing the time variability of dynamic height as inferred from the mooring temperature data.

We computed dynamic height 0/500 db during July–October 1987 by filling the 3-month temperature record gap (Fig. 4) with the mean vertical temperature gradient. This method, discussed in Kessler et al. (1985), assumes that temperature changes at depth are due to vertical displacements of the mean thermal structure. Sensitivity experiments in which we substituted a mean gradient for existing mooring data below 300 m indicate that the method leads to daily averaged surface dynamic height errors of <1 dyn cm.

The mean T-S profile we used was based on 65 US/PRC and SURTROPAC CTD casts from 2ºN and 2ºS along 165ºE for the pre-ENSO period January 1984–June 1986. Other choices of mean T-S are possible, e.g., a T-S based on all CTD data 1984–1988 or on CTD data for the period 1986–1988 coincident with the mooring time series. The principal effect of using these latter T-S curves is to offset the dynamic height time series from the one shown in Fig. 5e by about +2 dyn cm. However, temporal variability, which is what we are primarily interested in, is little affected by the choice of mean T-S profile. Cross-correlation of daily time series based on the different T-S curves is >0.99 and the standard deviation of differences is <1 dyn cm.

Figure 5e shows individual dynamic height estimates based on 12 CTD casts superimposed on the 0/500 db dynamic height time series. The standard deviation of the differences between daily averaged dynamic heights from the moorings and dynamic heights from the CTD data is 4 dyn cm, comparable to the value found by Emery and Dewar (1982) from historical data. The negative bias in the mooring time series in 1987 is due to an ENSO-related freshening of the surface layer by over 1 psu which is not contained in the mean T-S.

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