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Upper ocean variability on the equator in the Pacific at 170°W

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Abstract. The monitoring of upper ocean velocity, temperature, and surface winds at 0°, 170°W was initiated in May 1988 as part of the Tropical Ocean–Global Atmosphere program. Located between regions of warm (cold) sea surface temperature in the western (eastern) equatorial Pacific, this west central region exhibits large interannual variations in surface wind stress associated with the El Niño–Southern Oscillation. Here we report on the first 3 years of data collected at 0°, 170°W, exclusive of an El Niño event. The west central Pacific is a region through which the Equatorial Undercurrent (EUC) accelerates downstream. The vertical position of the EUC core, the vertical penetration of the South Equatorial Current (SEC), and the vertically integrated zonal volume transport all show large annual cycles. Annual and higher-frequency transport variations are due primarily to fluctuations occurring above the thermocline, as opposed to within the EUC itself. Unlike the EUC position, the EUC core speed remains relatively steady annually, except during boreal fall and winter, when short duration, zonal momentum pulses generated as Kelvin waves to the west reduce the EUC to minimal values. Interannually, the EUC core speed was highest in 1988, and it has decreased, on average, since then. Also during 1988, tropical instability waves, generally observed farther east, were well developed at 170°W, while not in subsequent years. This suggests the same role interannually as annually for these waves: to smooth out heat and momentum gradients that form in response to the wind stress. (The winds in 1988, following the 1986–1987 El Niño, were stronger than in subsequent years.) At higher frequencies, evidence exists for an inertial-gravity wave mode previously identified in nearby Canton Island sea level records.

1. Introduction

As part of the Tropical Ocean–Global Atmosphere (TOGA) program's Pacific monitoring array, upper ocean velocity, temperature, and surface wind measurements were initiated at 0°, 170°W in May 1988 with tandem deployments of a subsurface acoustic Doppler current profiling (ADCP) mooring for velocity and a surface ATLAS mooring for temperature and winds. These moorings have been serviced on an annual basis since then. The mooring location relative to the sea surface temperature (SST) pattern for July 1990 and the other current meter and ATLAS moorings composing the TOGA monitoring array at that time are shown in Figure 1. This west central Pacific site was chosen to fill a gap between measurements within regions of warm SST to the west and cold SST to the east. Since understanding the evolution of the coupled ocean-atmosphere system and the related interannual wind stress and SST variability, collectively called the El Niño–Southern Oscillation (ENSO), is a primary objective of the TOGA program, this transitional region is an important one for developing long time series on upper ocean variability.

This paper reports on findings from the first 3 years of data collected at 0°, 170°W. The background for initiating these measurements is given in section 2, along with a review of

self-contained ADCPs for profiling upper ocean currents which led to our selection of this instrument for equatorial application. Section 3 describes the initial results, and section 4 provides a discussion and summary.

2. Background

0°, 170°W as an ENSO Monitoring Site

One outcome of the first half of the TOGA program is an improved understanding of ENSO as an oscillation of the tropical Pacific's coupled ocean-atmosphere system. Models of this coupled system (e.g., as reviewed by *McCreary and Anderson* [1991]) that build upon *Bjerknes's* [1969] hypothesis and *Gill's* [1980] formalism relating SST and wind anomalies point to processes occurring along the equatorial Pacific waveguide as being important. The delayed oscillator mechanism of ENSO, for example, depends on the generation of wind anomalies in the west central part of the basin by SST anomalies occurring farther east. The wind anomalies strengthen the SST anomalies through downwelling Kelvin waves, providing positive feedback for anomaly growth. At the same time, these wind anomalies generate Rossby waves which, upon reflection from the western boundary as upwelling Kelvin waves, provide negative feedback and thus the possibility for oscillations. Such oscillations depend upon ocean dynamics for redistributing mass and heat and upon thermodynamics for controlling SST and surface heat fluxes. The west central Pacific is important both dynamically and thermodynamically, since there the easterly wind stress and its interannual

¹Deceased July 30, 1992.

Figure 1. The locations of the 0°, 170°W subsurface ADCP and surface ATLAS moorings relative to the SST pattern for July 1990 and the other elements of the Tropical Ocean–Global Atmosphere/Tropical Atmosphere Ocean (TOGA-TAO) array at that time.

variations are large [e.g., Wakata and Sarachik, 1991], and SST and thermocline depth are not simply related.

Previous upper ocean velocity time series from the equatorial central Pacific have been located east of 152°W. Limited vertical resolution time series are available at 152°W from April 1979 to May 1980 at 50 and 100 m [Lukas, 1987] during the Hawaii to Tahiti shuttle cruises [e.g., Wyrki et al., 1981]. Firing et al. [1983] report on meridional sections of currents obtained over the entire water column from 3°N to 3°S along 159°W during the 1982–1983 El Niño. A particularly notable finding was the disappearance of the Equatorial Undercurrent (EUC) during that event. Using a general circulation model, Philander et al. [1986] described the seasonal cycle of the entire tropical Pacific. The model showed the EUC and the South Equatorial Current (SEC) varying out of phase annually in the central Pacific, with the EUC (SEC) being maximum (minimum) in boreal spring and conversely in boreal fall. The EUC speeds during the spring maximum were 120 cm/s, compared with 80 cm/s during the fall minimum, and the westward flowing SEC was always confined above 25 m. Using the same model, Philander and Seigel [1985] simulated the 1982–1983 El Niño. The observations at 0°, 170°W have similarities and dissimilarities with these studies, as will be shown in section 3.

Acoustic Doppler Current Profilers for Upper Ocean Currents

Introduced to the ocean sciences for measuring velocity relative to a moving ship, oceanographic Doppler sonars follow the same principles as meteorological Doppler radars. Pettigrew and Irish [1983] review these developments and compare data from a bottom-mounted, RD Instruments 300-kHz ADCP with data from conventional EG&G, Inc., vector-averaging and vector-measuring current meters (VACM and VMCM, respectively), deployed together on the California continental shelf. Mean speed offsets and rms speed deviations between the ADCP and the mechanical current meters were generally less than 0.5 and 2 cm/s, respectively, and correlation coefficients exceeded 0.97, except near the surface at 10 m. Based on laboratory calibration studies, Pettigrew et al. [1986]

argued that the ADCP more closely measures water velocity than either the VACM or the VMCM does, since these mechanical devices overspin and underspin, respectively, when subjected to high-frequency, oscillatory flows.

The incorporation of firmware for in situ coordinate transformations facilitated the use of ADCPs on moorings. Schott [1986] reported on a proof-of-concept experiment in the Florida Current, showing that subsurface-moored ADCPs can provide accurate velocity profiles if the mooring tilts are small. Using a 150-kHz ADCP with a 20° transducer orientation, Schott and Johns [1987] obtained near-surface velocity profiles in the Arabian Sea spanning the monsoon seasons with currents of 150 cm/s. Combined with subsequent applications by Pettigrew et al. [1987] and Johns [1988], these results demonstrated the utility of using ADCPs on subsurface moorings for profiling large-amplitude, near-surface currents, which led to our application of a subsurface ADCP mooring in the equatorial region at 0°, 170°W in May 1988.

Argos satellite telemetry was then added to a surface-moored, downward looking ADCP by McPhaden et al. [1991]. Comparing ADCP data at 14 m with VMCM data at 10 m on an adjacent (17 km away) mooring, these authors found (after correcting for shear between 10 and 14 m) that the two velocity records were indistinguishable to within the RD Instruments' specifications of 1.4 cm/s, given the number of pings per ensemble used. Consistent with previous findings, the ADCP appears to be as accurate for upper ocean equatorial applications as any other commonly used moored current meter.

3. Description of the Measurements at 0°, 170°W

Moorings Performance and Sampling

The ATLAS system [Hayes et al., 1991] measured surface wind velocity (at 3.5 m), air temperature, SST (at 1 m), and subsurface temperatures at 25, 50, 75, 100, 125, 150, 200, 250, 300, and 500 m. Figure 2 shows these depths in relation to temperature and salinity profiles obtained on the deployment

Figure 2. Distribution of ATLAS mooring temperature sensors relative to temperature and salinity profiles obtained during the deployment cruises.

cruises. The near-surface temperature is nearly constant to a depth of 75–100 m, and the thermocline extends from this point to about 200 m. Salinity gradients are generally weak within the surface layer; salinity increases near the top of the thermocline and relative salinity extrema are observed within the thermocline. Despite these salinity variations, density is primarily determined by temperature, except in a few conductivity-temperature-depth (CTD) casts which show a thin surface layer (5 m) with relatively low temperature and salinity, presumably owing to rainfall [e.g., *Lukas and Lindstrom, 1991*].

The ADCP mooring employed an RD Instruments, 150-kHz unit with a 20° transducer orientation (nominal bin size of 8.68 m) and a targeted instrument depth of 300 m. In each of the three deployments the ADCP settled to a modal depth from which it made downward vertical excursions owing to mooring drag. The modal depths for the three deployments were 250, 300, and 310 m, respectively. The depth difference between the first and subsequent deployments may have resulted from an underestimate of Dacron rope (used below 1000 m) stretch; the other deployments used wire rope throughout. Small depth discrepancies between, or vertical excursions during, deployments are irrelevant with the ADCP, since the profiles can be resampled by linear interpolation. In this case the moorings were stable, with typical vertical excursions of only a few meters. Maximum vertical excursions measured during the three deployments were 30, 70, and 25 m, respectively, and these occurred as events, generally in boreal summer when the EUC was most strongly developed and in boreal winter when the upper ocean exhibited a sequence of velocity reversals. Apparently, these upper ocean, eventlike processes have deeper signatures as well. Pressure records are available for the first and third deployments; however, the pressure sensor failed after 2 months of the second deployment, so navigating the second deployment profiles required a relationship between the profiles of echo return amplitudes and the instrument depth. This was obtained using data from the first de-

ployment and the portion of the second deployment for which a pressure record was available. The results show that the ranges from the transducer to the positions of maximum and minimum echo amplitude, defining the principal beam and initial side lobe echo returns from the surface, respectively, are capable of determining the transducer depth to within about one half of a sampling bin. Thus all of the profiles were vertically navigable despite the failed pressure sensor.

The processing steps were to correct each profile for magnetic deviation and sound speed at the transducer, and then to linearly interpolate the data between bins, resampling the profiles at fixed 10-m intervals using a depth-averaged sound speed of 1531 m/s (corresponding to a bin size of 9.04 m). Since the sound speed is slightly less (greater) than the depth average below (above) the thermocline, the cumulative depth errors tend to cancel with range after reaching a maximum of about 1.0 m within the thermocline. The instrument calculates Cartesian velocity components in firmware, using the housing orientation determined by a compass and tilt sensors. Predeployment and postdeployment compass calibrations showed that they were within the 2° accuracy (after read-only memory (ROM) corrections) as claimed by the manufacturer. Mooring tilts on the first and third deployments were small (less than 1°); however, one of the two tilt sensors failed during the second deployment. While the mooring did not tilt, one sensor indicated tilts systematically increasing from zero to 6° over the first two thirds of the record and then reaching as high as 19° during the last 2 months. The resulting errors (uncorrectable, since ensemble averages were stored, not the individual pings composing these averages), are of two forms, both proportional to the cosine of the tilt angle. The first is an underestimate of the horizontal velocity vector by mapping it onto a tilted coordinate system, and the second is a smearing of depth bins because of the differing slant ranges among the four different beams. The first error is negligible over the first two thirds of the record, but it then increases to about 6% over the

Figure 3. Representative time series from the ADCP and ATLAS moorings. (a) Hourly sampled u component at 30, 130, and 200 m, (b) hourly sampled v component at 30, 130, and 200 m, and (c) daily averaged temperature at 25, 125, and 200 m.

Figure 3. (continued)

last 2 months. Similarly, over the last 2 months the vertical navigation of the profiles may be smeared from between zero near the transducers to about two bins near the surface. These problems are comparable to those encountered using conventional current meters for which mooring line tilts and vertical excursions also result in cosine response errors and depth smearing. These problems only occurred during the second deployment; during the first and third deployments, there are no indications that the rms velocity errors should differ from the manufacturer's specifications of about 1.2–1.1 cm/s, given the 240–280 pings per ensemble (at a 3-s ping rate) that was used.

The Data

Representative time series for the hourly sampled east (u) and north (v) velocity components at 30, 130, and 200 m and the daily-averaged temperature (T) at 25, 125, and 200 m are shown in Figure 3. The geometry of the ADCP transducer array results in contamination of the near-surface bins by reflections perpendicular to the mean sea surface. Sound propagating vertically and reflecting from the sea surface will return to the transducer at the same time as sound propagating along the transducer normal direction and reflecting within the water column at the same range. The nominal distance that is contaminated is $D(1 - \cos \phi)$, where D is the transducer depth and ϕ is the transducer beam angle. With $D = 300$ m and $\phi = 20^\circ$, such side lobe reflections become problematic within 18 m of the surface. Thus 30 m is the closest 10-m interval to the surface that should be uncontaminated, and this is supported by the echo amplitude distribution, wherein data composing the 30-m sample generally fall below the bin of minimum echo amplitude. The 130-m time series coincides approximately with the mean depth of the EUC core and the 200-m depth is the deepest with coincident temperature data. These three time series near the surface, within the EUC core and within the low-speed region beneath the core, are typical of observations made elsewhere within the equatorial waveguide using conventional current meters.

The mean currents are highly anisotropic (Figure 4), as

expected (e.g., Weisberg *et al.* [1987] and Halpern *et al.* [1988] for the equatorial Atlantic and Pacific, respectively). Contrasted with the small v component means, the u component means are large, showing a well-developed EUC positioned within the thermocline, with the largest variability occurring above the EUC core. Comparing the mean vertical profiles here with those observed farther east [e.g., Halpern *et al.*, 1988] or west [e.g., McPhaden *et al.*, 1990] shows that 0°, 170°W lies within the region through which the EUC accelerates downstream.

Evolution of the Low-Frequency Variations in T , u , and v : May 1988–March 1991

The evolution of temperature as a function of depth and time is shown in Figure 5. Gaps in the record are due to ATLAS mooring sensor or cable failures. The thermocline appears to be shallowest in boreal spring and summer and deepest in boreal fall and winter (boreal is implicit hereafter in regard to seasons). During fall and winter, large oscillations are observed in the isotherm depths with periods of about 2 months. These oscillations, however, do not show any clear relationship with either the SST or the local zonal component of surface wind stress (τ^x). Along with annual variability, there is an indication of interannual variability, with the thermocline being shallower during the spring and summer of 1988 than during comparable times in 1989 and 1990, as evidenced in the vertically integrated temperature. SST was also about 1°C cooler in summer 1988 than in summer 1989. Taking the depth of the 27°C isotherm as proxy for the depth of the mixed layer, the mixed layer is shallower in spring and summer than it is in fall and winter, consistent with the vertically integrated temperature, but again without any obvious relationship with SST. During February 1990, when the mixed layer and thermocline were deep, there occurred a short-lived increase in SST, coincident with the relaxation of the local τ^x . This temperature response of 1.0–0.5°C between the surface and about 100 m occurred over an approximate 5-day interval as the vertically integrated temperature was decreasing and the near-surface flow was small. The subsequent cooling occurred equally rap-

Figure 4. The vertical distributions of the record length means (solid lines) and the means plus and minus the standard deviations (dashed lines) for the u and v components and temperature at all sampled depths (10-m intervals for the ADCP mooring and at 1, 25, 50, 75, 100, 125, 150, 200, 250, 300, and 500 m for the ATLAS mooring). Note the scale difference between the velocity components. The change at 220 m for v is due to the record length difference below that depth.

Figure 5. Isotherms as a function of time and depth in relation to the zonal component of surface wind stress τ^x , sea surface temperature (SST), and temperature vertically integrated over the upper 300 m of the water column. The temperature time series are from daily averages; the τ^x time series has been low-pass filtered to remove oscillations at timescales shorter than 10 days.

Figure 6. Isotachs of the u component as a function of time and depth in relation to the zonal component of surface wind stress τ^x , the u component at 30 m depth, and the u component vertically integrated over the upper 220 m of the water column. All time series have been low-pass filtered to exclude oscillations at timescales shorter than 10 days. Shaded regions denote westward flow, and the contour interval is 20 cm/s.

idly, coincident with the increase in τ^x and a further decrease in vertically integrated temperature. The decrease in SST is therefore consistent with upwelling. Comparing the v component with SST in Figure 3 shows the warming to be consistent with meridional advection, since v peaked at 60 cm/s when SST was increasing, whereas u was nil, and a surface flux contribution alone would have to have been unrealistically large.

The evolution of the u component as a function of depth and time is shown in Figure 6 after low-pass filtering to remove oscillations at timescales shorter than 10 days. The differences in depth ranges here and in Figure 8 are due to differences in the three subsurface buoy deployment depths, and the values above 30 m are drawn by linearly extrapolating the 30-m values, using the shear between 40 and 30 m. Annual cycles are observed in the near-surface SEC, the vertical position of the EUC core, and the vertically integrated transport. During spring and summer the SEC is weak or absent, while the EUC is strong and shallow. Conversely, during fall and winter the SEC is more intense and penetrates as deeply as 150 m, while the EUC is weak. In all 3 years the fall and winter seasons are characterized by a sequence of reversing zonal momentum pulses which largely alter the zonal volume transports and cause the surface flows to switch direction. These pulses appear to result from propagating disturbances, in that their

associated local accelerations are large (order 10^{-4} cm/s²), their vertical shears over the surface mixed layer are small, and there is no consistent correlation with the local τ^x . When combined with the annual cycle, these zonal momentum pulses result in the EUC being strongest, with eastward flow appearing at the surface at two different times during the year: the transition between spring and summer and the transition between fall and winter. June 1988, May 1989, and July 1990 all show a strongly developed EUC existing over the entire depth range, with a high-speed core at about 120 m. This has the appearance of an annual cycle. In contrast, January 1989, December 1989, and November 1990 also show a strongly developed EUC, but associated with the annually occurring momentum pulses.

Winter 1990 provides an exception to the lack of correlation between the surface currents and the local τ^x . In particular, during February 1990, westerly winds accelerated the near-surface currents eastward, sandwiching a layer of westward flow between eastward flow above and below. *Hisard et al.* [1970] reported a similar finding farther west at 0°, 170°E in April 1967, *Philander et al.* [1986] commented upon this in their numerical model simulation, and *McPhaden et al.* [1988] have also observed this behavior at 0°, 165°E. While this flow regime, with eastward flow at the surface separated from the

Figure 7. The depth of penetration for westward flow, the depth of the EUC core, and the speed of the EUC at the core depth. All time series have been low-pass filtered to remove oscillations at timescales shorter than 10 days.

EUC, does occur, it is unique to this period of eastward wind stress. At other times when eastward flow is observed near the surface, it is contiguous with the EUC.

The u component at 30 m depth is correlated with both the vertically integrated zonal transport and the vertically integrated temperature. Coincident with the fall and winter momentum pulses, when the surface flow is strongly westward and the EUC is diminished in magnitude, the vertically integrated transport may be westward. This occurred in 1989, despite the fact that the local τ^x remained relatively steady that winter, in contrast to 1990 when the local τ^x relaxed. Both remote and direct forcing are clearly influential at this location.

The depth to which surface westward flow penetrates, the depth of the EUC core, and the u component speed at the EUC core are shown in Figure 7. The westward flowing SEC is generally confined to the upper 50 m, except during the winter events, when it penetrates as deeply as 150 m. In contrast to the very distinct annual cycle in the EUC core depth, ranging from around 120 m in spring/summer to 180 m in fall/winter, the speed at the core is relatively steady over the year. The exceptions are the fall and winter pulses, the westward phase of which reduces the EUC core speeds to as low as 25 cm/s. Thus, over the annual cycle, the EUC core speed is relatively steady, except when strong westward current bursts extend down with the mixed layer. An interannual trend is also observed, with EUC core depth rising and EUC core speed decreasing steadily between 1988 and 1991. This latter finding will be of particular interest as the record lengths become longer.

The evolution of the (similarly low-pass filtered) v component as a function of depth and time is shown in Figure 8. Unlike the u component, the v component oscillates at relatively high frequency, and these oscillations are modulated annually and interannually. During fall 1988 and winter 1989 a set of regular, 3-week-period oscillations are observed in v and its vertically integrated transport. These oscillations are confined primarily to the mixed layer, and they are reminiscent of the instability waves that have been observed in both the Atlantic and Pacific Oceans [e.g., Halpern and Weisberg, 1989]. However, these waves are not prominent at this location during the succeeding 2 years of record, nor are they visually correlated with the local τ^y , which itself varies interannually. During 1988, τ^y , with the exception of high-frequency fluctuations, was zero, while during the succeeding 2 years it was southerly, indicative of southeast trade winds, except during fall and winter. The vertically integrated meridional transport also shows interannual variability, being southward, northward, and then southward again during the 3 years.

To further examine the relationship between the local wind stress and the currents, a coherency analysis was performed between the wind stress components and the velocity components observed at 30 m depth, and the velocity component shears observed between 40 and 30 m. This was done separately for the different years of record owing to the breaks in the wind data, with the results being consistent from year to year. No prominent bands of coherence were found between τ^x with either u or u_z , or between τ^y with either v or v_z , similar

Figure 8. Isotachs of the v component as a function of time and depth in relation to the meridional component of surface wind stress τ^v , the v component at 30 m depth, and the v component vertically integrated over the upper 220 m of the water column. All time series have been low-pass filtered to exclude oscillations at timescales shorter than 10 days. Shaded regions denote southward flow, and the contour interval is 20 cm/s.

to the finding in the equatorial Atlantic Ocean by Weingartner and Weisberg [1991]. This lack of coherence may arise in several ways. For example, the direct coupling between the local wind stress and the upper ocean velocity may occur over shorter vertical scales, or local coupling, in a spectral sense, may be masked by a broad wavenumber bandwidth imposed by propagating equatorial waves. At a midlatitude location, Price *et al.* [1987] found the momentum transfer between the local wind stress and the currents to occur at depths shallower than 25 m. At the equator in the central Pacific, Santiago-Mandujano and Firing [1990] found that while shear and wind stress directions were correlated, the shear and wind stress magnitudes were not. This provides another possible explanation, since it implies (by stress continuity across the air-sea interface) variable vertical eddy diffusivity proportional to wind speed squared. Understanding the local coupling between the wind stress and the upper ocean currents, therefore, remains an important research topic.

Higher-Frequency Variability

Variance density spectra for the velocity component fluctuations (Figure 9) are similar to those found in other regions of the equatorial oceans, with u and v component anisotropy following from equatorial waveguide dynamics [Matsuno, 1966]. The peak occurring in the v component spectrum at periodicity around 500 hours relates to seasonally modulated waves generated by surface current instability [Philander,

1978]. Using a complex demodulation analysis over a bandwidth containing most of the wave variance, Figure 10 shows the amplitude modulations of the u and v components as functions of time and depth. The instability waves at this location are surface confined, as elsewhere, and both seasonally and interannually modulated. The years 1988 to early 1989, when the trade winds were stronger than normal following an El Niño, show well-developed instability waves, at a time when the eastern Pacific surface cold tongue penetrated into the west central Pacific. In contrast, the subsequent 2 years, when the trade winds were less intense, do not show well-developed instability waves. Given this interannual modulation, the amplitude and phase distribution with depth is described over the first deployment interval May 1988–June 1989 using a complex empirical orthogonal function (EOF) analysis performed over the same bandwidth. The first mode accounts for 70 and 47% of the variance in v and u , respectively. As shown in Figure 11, the amplitudes for v and u decrease to background levels below 160 m, and the v component phase is nearly uniform over this domain, while the u component phase decreases monotonically with depth across the mixed layer and the upper thermocline. Since the EOF phase differences are relative to the 30-m time series, the absolute phase difference between u and v requires a measure of the phase difference at 30 m, which from the cross spectrum has u lagging v by about -0.55 rad. Thus, near the surface, u and v are nearly in phase, while at the top of the thermocline, u and v are nearly in quadrature.

Figure 9. Variance density spectra for the u and v components of velocity sampled at 50 m depth.

These amplitude and phase distributions are consistent with generation by barotropic instability, since the horizontal Reynolds stress $\langle u'v' \rangle$ is a maximum where u' and v' are maximum and in phase and zero where u' and v' are in quadrature. Since the tropical instability waves are important for SST and air-sea interaction [Hayes *et al.*, 1989], their appearance at this west central Pacific location, the transition region between the warm-pool waters to the west and the cold-tongue waters to the east, suggests that they serve a similar role on interannual timescales as they do on annual timescales. Consistent with instability, this role is to reduce the upper ocean heat and momentum gradients. Using Geosat altimetry data at 5°N, Perigaud [1990] also reported oscillations with instability wave scales extending into the west central Pacific during 1988.

Prominent spectral peaks are also observed in the v component at timescales associated with equatorial inertial-gravity waves. One of the findings of Wunsch and Gill [1976], in which tropical Pacific sea level fluctuations were compared with equatorial wave theory, was a spectral peak centered on 3.93 days' periodicity having properties consistent with an inertial-gravity wave of first vertical and second meridional mode. This peak was most pronounced at Canton Island (located at 2.75°S, 171.75°W), which is close to the ADCP mooring. The spectral peak in Figure 9 with the most coherent vertical structure is the one at 3.93 days, and the associated velocity fluctuations are rectilinear, with north-south orientation and high ellipse stability. A complex EOF analysis over the bandwidth encompassing this peak shows that 73% of the variance is accounted for by one mode with nearly uniform amplitude and phase distri-

butions over the measurement domain (Figure 12). Independent analyses performed over each of the 3 years also show these results to be stationary over this interval. For a first baroclinic, second meridional inertial-gravity wave mode, sea level should peak near the latitude of Canton Island, while the v component should peak on the equator. The ratio of the dimensional rms amplitudes observed for v on the equator (3.0 cm/s) and for sea level at Canton Island ($\eta = 0.94$ cm) agrees with theory to within about 20% ($v_{\text{rms}}/\eta_{\text{rms}} = 1.05$ g/c, where g is gravity, c (=2.8 m/s) is the first baroclinic mode phase speed, and 1.05 follows from the eigenfunctions), and this agreement is statistically significant at the 90% level (using an F test on variance ratio). With the caveats that the data sets are separated in time and the baroclinic mode phase speed is uncertain, this velocity finding at 0°, 170°W is consistent with the sea level result of Wunsch and Gill [1976].

4. Discussion and Summary

The initial 3 years of upper ocean velocity and temperature data collected at the TOGA Tropical Atmosphere Ocean (TAO) array, 0°, 170°W, site show variability over a broad range of timescales. Descriptions focused upon the annual cycle, interannual variations, including the interannual modulation of the higher-frequency tropical instability waves, and what appear to be stationary oscillations at inertial-gravity wave timescales. Interannual variations in currents and temperature are expected because of the large interannual wind stress variations that occur over the west central equatorial

Figure 10. Complex demodulation amplitudes as a function of time and depth for the v and u component computed over the frequency band 0.00125–0.00275 cph, encompassing the oscillations associated with tropical instability waves. Amplitudes exceeding 20 cm/s are highlighted by shading.

Pacific. Annual variability arises from responses integrated over the entire equatorial waveguide, despite the fact that the annual wind stress variability at this location is relatively small. As predicted in the simulation by *Philander et al.* [1986], the EUC and the SEC vary out of phase on the annual cycle, with the EUC being strongest when the SEC is weakest, and conversely. The EUC core speeds are similar in magnitude to those modeled; however, the SEC penetrates deeper than predicted by the model. In the model the thermocline is thicker, enabling the modeled EUC to be closer to the surface. In nature the thermocline is sharper beneath a deeper mixed layer, hence the deeper penetration of the westward flowing SEC.

The annual cycle is not uniform among all quantities. While the EUC core depth and the vertically integrated zonal transport both show clear annual cycles, the annual cycles in the EUC core speed and the thermocline depth (or vertically in-

tegrated temperature) are not as well defined. The EUC core speed is relatively steady except in fall and winter, when reversing momentum pulses occur, presumably owing to Kelvin waves caused by westerly wind bursts over the western part of the basin [e.g., *McPhaden et al.*, 1988]. Thus, during each of the 3 years, there are intervals when the EUC speed diminishes. Since this occurs annually, the observation by *Firing et al.* [1983] that the EUC nearly disappeared during the 1982–1983 El Niño may not have been unique; rather, it may have reflected a response of larger magnitude than during a more typical year. In terms of the ocean circulation then, the question arises regarding the conditions for which positive feedback may occur in the west central Pacific, facilitating the development of an ENSO warming farther east. The fall 1989 through winter 1990 seasons provide an example of a sequence of events which looked like a developing El Niño, yet nothing happened. The Southern Oscillation index was dropping and

Figure 11. Amplitude (rms) and phase as a function of depth for the first mode complex EOF computed for the u (dashed line) and the v (solid line) components over the frequency band 0.00125–0.00275 cph, encompassing the oscillations associated with tropical instability waves.

SST was warming, reaching a peak at the end of February 1990 following a burst in westerly winds, yet these SST and wind events were not sustained, and conditions shortly thereafter returned to normal.

In their comparison between equatorial Pacific (140°W) and Atlantic (28°W) variability, *Halpern and Weisberg* [1989] found that the annual variations in the EUC core speed at 140°W were much more pronounced than those at 28°W, and conversely for the vertical position of the EUC core and the thermocline. The present 170°W data, in this regard, resemble

the Atlantic (28°W) data more than they do the Pacific (140°W) data. Such contrasting observations should provide tests for numerical models and for hypothetical momentum balances such as reported by *Wacogne* [1990].

While the EUC core speed generally changes little over the annual cycle, a correlation does exist between the volume transport per unit width and the near-surface flow, both of which do have large annual cycles. This raises a question regarding the role of the EUC in the zonal transport variability at annual and shorter timescales. Using a time domain EOF analysis, 75% of the u component variance and 95% of the vertically integrated transport variance reside in the first mode (distinguishable from the second mode following *North et al.* [1982] with 16 degrees of freedom). This transport mode (Figure 13) has maximum amplitude at the base of the mixed layer, and it decreases rapidly across the thermocline, in contrast to the mean u component profile, which has a maximum in the thermocline. The implication is that the zonal transport fluctuations, on annual and shorter timescales, are primarily associated with fluctuations in the near-surface currents, as opposed to fluctuations of the EUC as a separate entity. Similar results were found for the equatorial Atlantic at 0°, 28°W by *Tang and Weisberg* [1993], who hypothesized that linear dynamics control the transport fluctuations on seasonal timescales, while nonlinear dynamics are important on longer timescales. The interannual decrease observed in the EUC core speed at 0°, 170°W may allow for a test of this hypothesis as the record length increases.

Ascribing the annual and shorter timescale transport fluctuations to forced equatorial wave dynamics would account for the lack of coherence between these fluctuations and the local wind stress. Still, one might expect the upper ocean shears to be correlated with local winds over some timescale and depth scale. For example, the mean shear within the mixed layer (Figure 13) for the first mode may reflect the mean easterly wind stress. Just how the local winds couple with the upper ocean in the tropics remains an important research question,

Figure 12. Amplitude and phase as a function of depth for the first mode complex EOF computed for the v component over the frequency band 0.01017–0.01167 cph, encompassing the inertial gravity wave peak centered upon 3.93 days periodicity.

Figure 13. The first-mode eigenfunction from a time domain EOF description of the u component variability (dashed line) compared with the mean u component profile (solid line).

especially in the west central Pacific where the interannual wind stress variations are largest.

The observations at inertial-gravity wave frequency of what appears to be a stationary, baroclinic wave mode raises another question. If the 0°, 170°W velocity measurements support the inertial-gravity wave mode explanation by Wunsch and Gill [1976] for sea level fluctuations at Canton Island, why haven't similar steady, inertial gravity wave vertical modes been observed in equatorial velocity measurements elsewhere?

In summary, subsurface-moored ADCPs provide a reliable way of monitoring upper ocean currents along the equator. The west central Pacific, 0°, 170°W monitoring site lies within the region through which the EUC accelerates downstream. An annual cycle is observed in the speed and depth range of the surface SEC and in the vertical positions of the thermocline and the subsurface EUC embedded therein. Along with the annual cycle are large, reversing zonal momentum pulses occurring in fall and winter. These fluctuations are remotely generated, consistent with Kelvin waves originating by westerly wind bursts over the western equatorial Pacific. On interannual timescales, a decreasing trend is observed in the EUC core speed, paralleling a decreasing trend in the zonal wind stress component. The year 1988, one of strong easterlies following an El Niño year, showed instability waves extending into the west central Pacific, as contrasted to subsequent years, when these waves were confined to points farther east. Inertial-gravity wave oscillations, consistent with previous sea level findings at nearby Canton Island, were found in all 3 years.

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