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Basin-wide zonal wind stress and ocean thermal variations in the equatorial Pacific Ocean

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Abstract. Wind data, obtained from the model results of the European Centre for Medium Range Weather Forecasts (ECMWF), for the equatorial Pacific Ocean from 1985 to 1992, were employed to study the basin-wide distributions for various constituents of zonal wind stress. Forcing the ocean by the individual constituent of zonal wind stress, the upper ocean thermal variation was investigated using a linear analytical ocean model. The interannual variation of zonal wind stress is largest in amplitude in the west-central part of the basin and is dominated by an eastwardly propagating wave. The upper layer thickness perturbation $h$ shows an evolving eastward propagation over the region where the interannual wind stress variation occurs, while to the east and west the region changes nearly in phase. The evolution of $h$ agrees well with the evolution of upper ocean heat content anomaly, estimated from the Tropical Atmosphere-Ocean (TAO) moored array. The annual cycle of zonal wind stress can be described properly by a combination of annual and semiannual variations. The annual variation of zonal wind stress was found to propagate westward with relative maxima out of phase to the west and east of the west-central Pacific. The semiannual variation is stationary and is largest in amplitude in the western basin. Forced by the two components of the annual cycle of zonal wind stress, the evolution of $h$ shows a westward propagation. This does not correlate well with the pattern of the evolution of the annual cycle of upper ocean heat content which emanates from the east-central Pacific and shows both eastward and westward propagation. The seemingly anomalous eastward propagation is mainly related to the abrupt change during the relaxation/intensification period of the easterly wind stress. The abrupt change of the annual zonal wind stress variation in both time and space is, then, critical for the evolution of the pattern of the annual ocean response.

1. Introduction

With the Pacific Ocean occupying half the equatorial circumference of the Earth, it is not surprising that continuing research efforts have confirmed early hypotheses, such as those of Walker [1928] and Bjerknes [1966], that tropical Pacific ocean-atmosphere interactions play a major role in global climate variations. Indeed, the El Niño-Southern Oscillation (ENSO), a broadband, approximately twice per decade variation in the basin-scale surface pressure, sea surface temperature (SST), and wind fields of the tropical Pacific [e.g., Philander, 1990], is very well correlated with global rainfall anomalies [e.g., Ropelewski and Halpert, 1987]. ENSO tends to be phase locked to the annual cycle, with the largest SST anomalies generally occurring over the equatorial cold tongue region in the month of December. It follows that an improved understanding of the annual cycle is an important element toward an improved understanding of the interannual climate variability, and prerequisites to this are needed improvements in the description of the annual and interannual evolution of the upper ocean fields in response to surface wind forcing.

Previous studies using either sophisticated numerical models or simplified analytical models, e.g., Hurlburt et al. [1976], Busalacchi et al. [1983], Tang and Weisberg [1984], Weisberg and Tang [1990], Philander [1990], and Kessler and McPhaden [1995], have shown that the thermal evolution along the equator can be closely reproduced if the zonal wind stress (as the primary low-frequency driving force) is adequately specified. Wyrtki and Meyers [1976], analyzing voluntary observing ship (VOS) wind data, found that these low-frequency zonal wind variations in the tropical Pacific are dominated by both interannual and annual fluctuations, the interannual variations being related to ENSO. Barnett [1977], using VOS data, found the interannual variability to be largest in the western Pacific, in contrast to the mean state zonal winds that are largest in the central Pacific. Meridionally, these interannual variations in the zonal winds are maximum on, and decrease rapidly poleward from, the equator [Harrison and Luther, 1990]. Using a complex empirical orthogonal function (CEOF) analysis, Barnett [1983] found that the VOS data interannual zonal wind variations propagate eastward, and similar eastward propagation has been described by Kawamura [1991] for the European Centre for Medium Range Weather Forecasts (ECMWF) 850-mbar zonal winds in the western tropical Pacific. In contrast to the interannual variations, Lukas and Firing [1985], using the Wyrtki and Meyers [1976] winds, described the annual harmonic of the zonal wind stress as propagating westward with a phase...
Figure 1. Zonal wind stress ($\tau_x$) along the equatorial Pacific from 1985 to 1992. The $\tau_x$ values were calculated from ECMWF model winds. Contour interval is 0.1 dyn cm$^{-2}$. The shaded regions indicate the westerly wind stress, the unshaded regions indicate the easterly wind stress, and the bold line is zero zonal wind stress.

speed of about 40 cm s$^{-1}$, half that subsequently estimated by Halpern [1988] over the eastern Pacific using buoy-derived winds. Notwithstanding estimated speed differences, the zonal phase gradient is sufficiently large that the annual harmonic variations tend to be out of phase between the western and eastern sides of the Pacific Ocean basin. Along with westward propagation the amplitude of the zonal wind stress annual harmonic appears to be smallest in the west-central portion of the basin [Meyers, 1979], where the interannual variability tends to be the largest [e.g., Wakata and Sarachik, 1991]. In addition to the interannual and annual variations the zonal winds also have semiannual and intraseasonal variations. The semiannual variations have been described as mainly confined to the western Pacific Ocean and may be as large as the annual variation [Goldenberg and O'Brien, 1981; Hored, 1982]. On the equator, Meyers [1979] found that the semiannual variations had their largest amplitude in the central equatorial Pacific Ocean.

Numerical ocean circulation models, either forced by observed winds or allowed to evolve coupled to an atmospheric model, tend to show interannual variations in the upper ocean thermal field that propagates eastward with slow, zonally non-uniform speed [e.g., Chao and Philander, 1993]. These authors also argue that the associated pattern propagation speed is too slow to be attributable to any single Rossby or Kelvin wave. At annual timescales, Meyers [1979] examined the evolution of the thermal field through the vertical displacements of the 14$^\circ$C isotherm. Westward propagation was found with a speed of 50 cm s$^{-1}$ with maximum displacements located around 150$^\circ$W and east of 110$^\circ$W, and it was suggested that this westward propagating response to similar westward propagating surface wind forcing may be the result of a near-resonant equatorial...

Figure 2. Spatial distribution of (top) time average and (bottom) temporal evolution of annual cycle of the zonal wind stress ($\tau_x$) along the equatorial Pacific estimated from the ECMWF model zonal wind stress from 1985 to 1992. Contour interval is 0.1 dyn cm$^{-2}$. The shaded regions indicate the westerly wind stress, the unshaded regions indicate the easterly wind stress, and the bold line is zero zonal wind stress.
Rossby wave. The characteristics of the Rossby wave were further explored by Lukas and Firing [1985] and Kessler and McCreary [1993]. Rasmusson and Carpenter [1982] and Mitchum and Lukas [1990] also describe westward propagation at annual timescale using SST and island sea level data, respectively. Semiannual variations in the 14°C isothermal depth with largest amplitude on the eastern side of the basin were also introduced by Meyers [1979]. These were attributed to remote forcing over the central portion of the basin where the semiannual variations of zonal wind stress had their largest amplitude. A similar conclusion was obtained using a numerical model forced by the same data set [Kindle, 1979]. Additional large-magnitude, upper ocean thermal structure variations occur at intraseasonal timescales, particularly in boreal fall and winter when westerly wind bursts are observed over the western portion of the basin, generating ocean responses that propagate eastward [e.g., Knox and Halpern, 1982; Kessler et al., 1995]. Although there have been extensive previous studies describing the upper ocean fields and the surface winds, ambiguities remain; for example, does the annual cycle of the equatorial thermocline propagate eastward, westward, or in some combination to forcing occurring over different portions of the basin, and how is such forcing distributed zonally? To help resolve such issues, the present paper describes the basin-wide zonal wind and thermal fields over the period 1985-1992 (chosen to encompass the period between two successive El Niño events) with emphasis on the interannual, annual, and semiannual variations. Data on the ocean thermal structure were obtained from the Tropical Atmosphere-Ocean (TAO) moored array [Hayes et al., 1991] and from the National Oceanographic Data Center (NODC) historical expendable bathythermograph (XBT) file. Data on the winds were obtained from the model results of the ECMWF. These modeled winds include the TAO mooring winds by data assimilation [Böttger, 1989], thus augmenting a relatively sparse and unevenly sampled array over the chosen interval. Although the overall effects of TAO array data assimilation into the EC-

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**Figure 3.** Temporal evolution of the annual perturbations in zonal wind stress ($\tau^x$) along the equatorial Pacific obtained by subtracting the spatial distribution of time average from the temporal evolution of annual cycle. Contour interval is 0.05 dyn cm$^{-2}$. The shaded regions indicate the westerly wind stress, the unshaded regions indicate the easterly wind stress, and the bold line is zero zonal wind stress.

**Figure 4.** Temporal evolution of the interannual perturbations in zonal wind stress ($\tau^x$) along the equatorial Pacific obtained by subtracting the temporal evolution of annual cycle from the original data (Figure 1). Contour interval is 0.05 dyn cm$^{-2}$. The shaded regions indicate the westerly wind stress anomaly, the unshaded regions indicate the easterly wind stress anomaly, and the bold line is zero zonal wind stress anomaly.
Figure 5. First mode of CEOF of zonal wind stress. (a) The amplitude, phase, real, and imaginary parts of the spatial eigenfunction ($E_1$). A linear least squares fit between the two arrows is employed to estimate phase slope (dashed line). (b) The amplitude, phase, real, and imaginary parts of the temporal transfer function ($G_T$). A linear least squares fit is employed to estimate phase slope (dashed line). (c) The actual data represented as a function of longitude and time. The contour interval is 0.05 dyn cm$^{-2}$. The shaded regions indicate the westerly wind stress anomaly, the unshaded regions indicate the easterly wind stress anomaly, and the bold line is zero zonal wind stress anomaly.
Figure 6. Second mode of CEOF of zonal wind stress. (a) The amplitude, phase, real, and imaginary parts of the spatial eigenfunction ($E_2$). A linear least squares fit between the two arrows is employed to estimate phase slope (dashed line). (b) The amplitude, phase, real, and imaginary parts of the temporal transfer function ($G_2$). A linear least squares fit is employed to estimate phase slope (dashed line). (c) The actual data represented as a function of longitude and time. The contour interval is 0.05 dyn cm$^{-2}$. The shaded regions indicate the westerly wind stress anomaly, the unshaded regions indicate the easterly wind stress anomaly, and the bold line is zero zonal wind stress anomaly.
MWF model winds have been questioned [Anderson, 1994], both the TAO data and the model winds are highly coherent.

This paper proceeds as follows. The ECMWF model winds are employed in section 2 to explore the zonal wind stress variations in space and time along the equatorial Pacific Ocean using CEOF analysis to distinguish interannual, annual, and semiannual variations in zonal wind stress. Similarly, interannual and annual variations in the upper ocean heat content are introduced using TAO and XBT data, respectively, and CEOF analyses are applied to describe the semiannual and annual harmonics individually. The CEOF wind modes are then used in section 3 to develop the basin-wide thermocline response to interannual, annual, and semiannual zonal wind stress forcing using a linear, reduced-gravity (one active layer) ocean model. The ocean responses to such constituent forcing are described individually and the interannual variations are compared with the TAO array measurements. For the interannual variability the zonal wind stress CEOF mode forced wave response adequately accounts for the zonal evolution of the upper ocean heat content; however, this is not true of the annual variability. Adding the contribution from semiannual forcing provides an improvement but still fails to accurately describe the variations over the eastern half of the basin, due primarily to rapid relaxation/intensification of easterly wind stress that is not represented by annual and semiannual harmonics. The results are discussed and summarized in section 4.

2. Wind Stress and Heat Content Analyses

2.1. ECMWF Zonal Component of Wind Stress

Since the TAO array sampling was sparse during its initial buildup phase, surface winds (10 m above the sea level) from the ECMWF model are used to analyze the basin-wide zonal wind stress variations along the equator in the Pacific Ocean from 130°E to 82.5°W. The zonal resolution is 2.5° (64 locations along the equator), the time interval is 12 hours, and the analysis duration is the 8-year period from 1985 to 1992. Prior to analysis the data have been low-pass filtered to exclude variations on timescales shorter than 30 days. As expected, owing to the assimilation of TAO winds into the ECMWF model, the correlation between the model and the TAO array winds is very high.

The zonal component of surface wind stress $\tau'$ was computed by

$$\tau' = \rho_C C_{10} \sqrt{w_x^2 + w_y^2},$$  

where $\rho_C$ is air density, $w_x$ and $w_y$ are the zonal and meridional wind velocity components, respectively, and $C_{10}$ is

![Figure 7. First two CEOF modes of the annual cycle evolution shown in Figure 3. (a) The portion of annual cycle evolution represented by the first CEOF mode as a function of longitude and time. The mode represents 63.09% of the total variance. (b) The portion of annual cycle evolution represented by the second CEOF mode as a function of longitude and time. The mode represents 25.14% of the total variance. The contour interval is 0.05 dyn cm$^{-2}$. The shaded regions indicate the westerly wind stress anomaly, the unshaded regions indicate the easterly wind stress anomaly, and the bold line is zero zonal wind stress anomaly.](image)
Plate 1. Upper ocean (300 m) heat content (left) monthly mean and (right) anomalies along the equatorial Pacific from 1985 to 1992. The monthly mean was calculated as the average of the TAO array data from 2°S to 2°N. The anomalies are the monthly means minus the annual cycle evolution estimated from the historical XBT data. Contour interval is $0.1 \times 10^{10}$ J m$^{-2}$. The figure is provided by the TAO Project Office, Pacific Marine Environmental Laboratory, National Oceanic and Atmospheric Administration.

The resulting zonal wind stress distribution as a function of time and longitude along the equator is shown in Figure 1. The bold (zero zonal wind stress) line separates the positive westerly wind stress (shaded) from the negative easterly wind stress (unshaded) regions. Easterly wind stress appears over most of the basin with maximum values in the central Pacific. Westerly wind stress is confined to the west of the date line and to the eastern boundary.

Performing a climatological mean state analysis over the 1985–1992 interval yields the spatial distribution of the time average and the temporal evolution of the annual cycle for the zonal wind stress as shown in Figure 2. Easterly wind stress appears over the whole basin for the time-averaged mean state, with largest amplitude at 140°W, decreasing nearly linearly on either side. West of the date line, the time-averaged zonal wind stress is very small. Only near the eastern boundary is the mean state westerly wind stress large, but with very small fetch. The mean annual cycle (which includes the spatial distribution of the time average) does not show a clear sense of propagation. The easterly wind stress is weakest following the boreal spring equinox (April/May), it intensifies during the summer solstice, reaching a relative maximum in July/August, and it then reduces to a relative minimum during the autumn equinox and then reaches an absolute maximum during the winter solstice. The asymmetry in the annual cycle with respect to the spring and fall equinoxes must be related to ocean atmosphere coupling since the insolation cycle is symmetric. West of the date line, the annual cycle shows maximum westerly wind stress during the boreal winter solstice and maximum easterly wind stress during boreal summer. At other times the zonal wind stress over the western part of the basin appears to be small. It is during the boreal winter solstice that the zonal wind stress variation across the basin is largest with maximum westerlies west of the date line and maximum easterlies east of the date line.
2.2. CEOF Analysis of the ECMWF Zonal Wind Stress Component

To further investigate the spatial and temporal characteristics of zonal wind stress and to prepare a simplified version for driving a simplified analytical ocean model, a CEOF analysis is applied using the method described by Barnett [1983]. A zonal wind stress data matrix is formed with each row being a demeaned time series for each of the 64 locations along the equator. A Hilbert transform is then applied, converting this data matrix to a complex data matrix denoted by C which is then subjected to a conventional EOF analysis using the covariance matrix obtained from C. This results in a set of real eigenvalues $\lambda_n$ and complex set of dimensional (dynes per square centimeter) eigenfunctions $E_n$. A set of complex, orthonormal transfer functions $G_n$ are then obtained by projecting the data matrix C onto the eigenfunctions $E_n$. The data matrix C may then be represented as

$$ C = \sum_{n=1}^{N} E_n G_n $$

where $N$ is the total number of modes. Thus each CEOF mode contains three parts: $\lambda_n$, $E_n$, and $G_n$ with the $\lambda_n$ representing the relative amount of variance accounted for and $E_n$ and $G_n$ representing the spatial and temporal variations of each mode, respectively. Since $E_n$ and $G_n$ are complex, they can represent a progressive wave. The spatial gradient of the phase is indicative of a wavenumber, and the temporal rate of change of phase is indicative of a frequency.

Modes one and two, in sum accounting for 70% of the total variance, are shown in Figures 5 and 6, respectively. The left (Figures 5a and 6a), middle (Figures 5b and 6b), and right (Figures 5c and 6c) panels give the spatial eigenfunction, the temporal transfer function, and the portion of the actual data represented by each of the two modes, respectively. The first mode captures the interannual variability with largest amplitude in the west-central part of the basin. In comparison, the amplitude on either side of the basin is small. Within the large-amplitude region the phase increases linearly to the east, and a linear least squares fit to the phase slope between the two arrows gives an equivalent wavenumber of $1.71 \times 10^{-4}$ rad km$^{-1}$. The first mode transfer function (seen most clearly in the real and imaginary parts) primarily shows an interannual variation although it also contains some annual variation. The general progression is eastward with a speed of some 24 cm s$^{-1}$ averaged over the 8 years of record, as obtained from the wavenumber estimated above and the slope of the transfer stress anomalies. Periods of anomalous westerly wind stress during the 1986/1987 and 1991/1992 El Niño events are very clear, separated by a period of anomalous easterly wind stress. The overall sense of pattern propagation is eastward, but the anomaly amplitude, fetch, and pattern propagation speed are all variable. Also, the pattern of zonal wind stress anomaly differs from that for the zonal wind velocity component anomaly. In particular, the location of the maximum zonal wind stress anomaly is shifted eastward relative to the zonal wind velocity component anomaly. The origin for this is the region of large mean state winds in the central portion of the basin that affects the stress anomaly calculation through the square of the wind speed but does not affect the velocity component anomaly calculation.
function phase \((-3.52 \times 10^{-3} \text{ rad d}^{-1})\). The corresponding wavelength and period are \(3.67 \times 10^4 \text{ km}\) and 4.90 years, respectively. Multiplication of the first eigenfunction and transfer function shows the first-mode representation of the zonal wind stress variations. This representation contains the basin-wide-scale interannual variability observed in the anomaly field of Figure 4 (largely smoothed by the CEOF analysis), and it also shows an interannual modulation to the annual cycle originating in the west-central Pacific.

The second mode is distinguished from the first primarily by the representation of the annual cycle. The spatial distribution shows relative amplitude maxima west and east of the west-central part of the basin and a relative minimum at the west-central part. The amplitude is also small on both sides of the basin. The second mode's phase also varies linearly with longitude, but contrary to the first mode it decreases to the east. The estimated slope gives a wavenumber of \(-3.67 \times 10^{-4} \text{ rad km}^{-1}\) corresponding to a basin-scale wavelength of \(1.71 \times 10^4 \text{ km}\). A linear fit to the temporal phase slope gives a frequency of \(1.71 \times 10^{-2} \text{ rad d}^{-1}\) corresponding to a period of 1 year.

The second mode representation of the zonal wind stress variations clearly shows a westward propagating annual cycle with relative maxima being out of phase to the west and east of the west-central Pacific. This pattern differs from the annual cycle evolution shown in Figure 3 during boreal summer months where the data suggest that westward propagation stalls by the date line. These details are contained in higher but nonseparable modes. To address this, a simplified CEOF analysis was performed on just the annual cycle evolution of Figure 3 (exclusive of the interannual variability). The results shown in Figure 7 suggest that the annual cycle may be adequately described by two modes, one primarily capturing an annual harmonic and the other primarily capturing a semiannual harmonic. The annual harmonic is almost identical with the second mode from the previous analysis. The semiannual harmonic is largest over the western half of the basin where it appears to develop uniformly without propagation.

### 2.3. Analysis of the Upper Ocean Heat Content

The TAO array has evolved since 1985 to its present configuration of approximately 69 buoys distributed over the tropical Pacific Ocean [Hayes et al., 1991; McPhaden, 1993]. Plate 1 shows the monthly heat content from the TAO array, inte-
integrated from the surface to 300 m and averaged from 2°S to 2°N, as a function of time and longitude along with interannual anomalies (relative to a historical subsurface temperature climatology [see Kessler and McPhaden, 1995]) for the period of 1985-1992. The central portion of the basin separates regions of maximum and minimum heat content on the western and eastern sides of the basin, respectively. An annual cycle is evident but embedded within a larger interannual variability as noted in the anomaly field. The largest heat content anomalies are ENSO related with warming in the east during the El Niño phase and cooling in the east during the La Niña phase. The central portion of the basin is where the interannual response appears to slowly evolve from western to eastern basin conditions; however, this evolution in both the upper ocean heat and the surface winds differs between events.

For similar reasons the TAO array is sparsely sampled during its buildup phase; the historical XBT data from the NODC files are used to analyze the annual cycle and how both the annual and semiannual constituents sum in composing it. Figure 8 shows the temporal and spatial evolution of the climatological annual perturbation in upper ocean heat content, calculated from the historical NODC XBT file. The upper ocean heat content was obtained by vertically integrated the temperature from 0 to 300 m in between 2°N and 2°S. A trapezoid rule was applied to perform the vertical integration. Details of the calculation are given by Kessler [1990]. The pattern of the annual perturbation appears to evolve from the east-central portion of the basin as the zonal wind stress intensifies after the boreal spring equinox. The pattern then suggests propagation both to the east and the west, consistent with the findings of Meyers [1979]. The similar pattern, but with sparsely sampled data, is also seen in the climatological annual perturbation in upper ocean heat content which was calculated from the TAO array temperature. Decomposing the upper ocean heat content obtained from the XBT data by CEOF results in the two modes shown in Figure 9 that collectively account for 90% of the variance. The first and second modes primarily represent annual and semiannual harmonics, respectively. Comparing the annual harmonic for the zonal wind stress that propagates westward with the annual harmonic in the upper ocean heat content which propagates both eastward and westward, it becomes clear that the spatial modulation of the winds, resulting in largest values over the region from which the heat content response emanates, is a critical element in the pattern evolution for the annual cycle.

3. Ocean Response

The zonal wind stress perturbations, decomposed into CEOF modes, may be used to drive a simplified analytical
model to explore the upper ocean heat responses to particular aspects of the wind variability. The ocean model is a linear, reduced-gravity, equatorial $\beta$ plane model with one active layer, using the formalism of Cane and Sarachik [1976, 1977] and previously applied, for example, by Tang and Weisberg [1984] for the equatorial Pacific and Weisberg and Tang [1990] for the equatorial Atlantic. The nondimensionalized equations of motion are

$$\frac{\partial u}{\partial t} - yv + \frac{\partial h}{\partial x} = r^* - \epsilon u$$

(4)

$$\frac{\partial v}{\partial t} + yu + \frac{\partial h}{\partial y} = -\epsilon v$$

(5)

$$\frac{\partial h}{\partial t} + \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = -\epsilon h$$

(6)

where $u$ and $v$ are the velocity components in the $x$ and $y$ directions, $h$ is the upper layer thickness perturbation, $t$ is time, $r^*$ is the zonal wind stress, and $\epsilon$ is the damping parameter. The equations have been nondimensionalized using timescales and length scales $T = (c/\beta)^{1/2}$ and $L = (c/\beta)^{1/2}$, where $c = (g'D)^{1/2}$ is the reduced gravity wave speed. The $g'$ is the reduced gravity, $D$ is the undistributed layer water depth, and $\beta$ is the gradient of planetary vorticity. The model is forced from a state of rest by temporally and spatially varying zonal wind stress distributions. The oceanic response is obtained by Fourier transforming the equations of motion, projecting the forcing function onto the appropriate equatorial wave modes of the homogeneous equations, integrating in time, and then inverting the Fourier transforms using a longwave approximation. The timescale and length scale used here are 1.54 days and 333.33 km, respectively, which are obtained by assuming $c = 250 \text{ cm s}^{-1}$ and $D = 250 \text{ m}$. The damping coefficient is 0.01 (equivalent to a 154 day $e$-folding time).

The forcing function is assumed to have a Gaussian distribution in meridional direction. Its zonal distribution and time variation are composed of a series of linear functions, chosen to approximate the eigenfunctions and transfer functions of CEOF modes. Thus the forcing function is written as

$$\tau_m = \gamma_m e^{-(r^*)^{1/2}K_m(x)}F_m(t)$$

(7)

where the time variation $F_m(t)$ obtained from the CEOF transfer function is given by

$$F_m(t) = H(t) \frac{t}{T_m} + \sum_{i=1}^{n} H(t - T_{m,i})S_m \frac{t - T_{m,i}}{T_{m,i}}$$

(8)
Figure 12. The time evolution of the upper layer thickness perturbation $h$ along the equatorial Pacific forced by the forcing function shown in Figure 10. Contour interval is 3 m. The shaded area indicates the positive upper layer thickness perturbation, and the unshaded area indicates the negative upper layer thickness perturbation.

and the spatial variation $K_m(x)$ obtained from the CEOF eigenfunction is given by

$$K_m(x) = H(x) \frac{x}{X_{m,1}} + \sum_{j=1}^{J_m} H(x - X_{m,1}) Q_{m,j} \frac{x - X_{m,1}}{X_{m,j}}$$  \hspace{1cm} (9)$$

where $m = 1$ or $2$ denotes the particular CEOF mode; $H$ is the Heaviside step function; $\gamma_m$, $S_{m,1}$, and $Q_{m,j}$ set the magnitudes for the $\tau_m$ of variations; and $T_{m,1}$ and $X_{m,j}$ set the duration and fetch for these variations. For the interannual variability, Figure 10 shows the $K_1(x)$ and $F_1(t)$ approximations for the first eigenfunction ($E_1$) and transfer function ($G_1$) and the resulting forcing function to be used in driving the model. Similarly, Figure 11 develops the forcing function for the annual variability. In either case the $K_m(x)$ and $F_m(t)$ approximations to $E_n$ and $G_n$ appear to be satisfactory.

Applying the Figure 10 interannual forcing function to the model results in an upper layer thickness perturbation $h$ response as shown in Figure 12. Shaded (positive) regions denote an increase in upper layer thickness, while unshaded (negative) regions denote a decrease. The model response qualitatively reproduces the evolution of the upper ocean heat content anomalies between the 1986/1987 and 1991/1992 El Niños and the intervening La Niña. The pattern evolution shows nearly in-phase changes in $h$ on the eastern and western sides of the basin as contrasted with eastward propagation over the west-central part of the basin. This occurs for two reasons. First, it is only over the region of maximum forcing, the west-central part of the basin, that one observes propagation in response to slowly varying forcing. Second, outside of the region of maximum forcing the ocean response propagates sufficiently rapidly to render the pattern evolution as appearing stationary. It is noted that running the model with or without east and west boundaries (Figure 12 is the bounded result) makes little difference in the overall pattern development since the solution is dominated by the directly forced response. The apparent propagation speed in upper ocean heat content is simply determined by the timescale of the forcing function. Although the features of the two El Niño events are generally similar, there are still some notable differences. For example, the upper layer thickness response in the eastern basin has a shorter duration, but larger amplitude, in the later events than the earlier events. This difference is attributable to the westerly wind anomalies in the central basin having a shorter duration but more rapid change and larger amplitude in the later event than the earlier.

Figure 13. Comparison between the time series of upper layer thickness perturbations $h$ (bold line) and upper ocean heat content anomalies (thin lines) at 155°E, 165°E, 170°W, 140°W, and 110°W on the equator. The $h$ and upper ocean heat content anomalies are obtained from Figure 12 and Plate 1, respectively.
event. The spatial and temporal modulations of the winds are therefore both important, perhaps more so than the addition of more vertical modes to the model.

The model results may be compared in a more quantitative fashion by sampling both the TAO array heat content and the modeled upper layer thickness at 155°E, 165°E, 170°W, 140°W, and 110°W, normalizing the anomalies, and then overplotting them as in Figure 13. Despite the high-frequency fluctuations in the TAO data that are not represented in the CEOF rendition of the winds used to force the model, there exists a reasonable agreement at interannual timescales. For example, Figure 14 shows the coherence squared and phase between the model result and the TAO data at 165°E. For interannual variability the data and model results are highly coherent and in phase. Similar results (not shown) are also obtained at the other locations.

![Figure 14](image1.png)

**Figure 14.** The (top) coherence squared and (bottom) phase between the upper layer thickness perturbation $h$ and upper ocean heat content anomaly at 165°E, 0°N. The 90% significance level is 0.27.

Similarly, applying the Figure 11 annual forcing function to the model results in an upper layer thickness perturbation $h$ response as shown in Figure 15. Again, shaded (positive) regions denote an increase in upper layer thickness, while unshaded (negative) regions denote a decrease. Relative maxima in annual variability are noted in the east-central portion of the basin and on the eastern and western boundaries. Westward pattern propagation is also generally noted, and this differs from the climatological annual cycle that shows both westward and eastward propagation. This lack of agreement in the east requires explanation. To help explain this, we repeat the response analysis using the simplification to the CEOF analysis of the climatologically averaged annual cycle wind stress of Figure 7. The results are show in Figure 16. Figure 16a is the response to the annual harmonic. Primarily observed is westward propagation with some indication of eastward propagation in late fall. Eastward propagation does result from the semiannual harmonic (Figure 16b) since this harmonic has the largest amplitude on the eastern side of the basin. Adding these two constituent responses together (Figure 16c) does
Figure 16. (a) The time evolution of upper thickness perturbations $h$ along the equatorial Pacific, forced by the simplification of the first CEOF modes of the annual cycle evolution of zonal wind stress (Figure 7a). (b) The time evolution of upper thickness perturbations $h$ along the equatorial Pacific, forced by simplification of the second CEOF modes of the annual cycle evolution of zonal wind stress (Figure 7b). (c) The combination of Figures 16a and 16b the time evolution of upper thickness perturbations $h$ along the equatorial Pacific. Contour interval is 3 m. The shaded area indicates the positive upper layer thickness perturbation, and the unshaded area indicates the negative upper layer thickness perturbation.

Figure 17. The time series of zonal wind stress ($\tau^x$) at 150°W, 0°N for 1989 (thin line), 1990 (dashed line), 1991 (dash-dotted line), and annual cycle (bold line). The annual cycle at 150°W is a time slice taken from Figure 4.
show a combination of westward and eastward pattern propagation, but with the latter occurring too late in the year. Shifting the semiannual variation to the central Pacific where Meyers [1979] and Kindle [1979] found the largest semiannual forcing only slightly improves the result. The eastward propagation of $h$ perturbation emanating from the central basin may be a consequence of oversmoothing by the CEOF wind stress, which will be developed in the next section.

4. Discussion and Summary

Comparing the climatological average annual variation of zonal wind stress with that of the individual years, we found that either the relaxation or intensification of easterly wind stress is much more abrupt in the individual year than in the climatological average. This difference is nearly basin wide so that it should have a large impact on the annual cycle of ocean response. As an example, Figure 17 shows the temporal zonal wind stress variations at 150°W, 0°N for 1989, 1990, 1991, and the climatological average. Since the ocean response is integrated over the entire forcing region [Weisberg and Tang, 1987], the temporal variation of ocean response will not be as abrupt as the zonal wind stress. For example, an impulsive temporal forcing function which is zonally distributed in the interval of $l$ will generate a Kelvin wave with a timescale of $l/c$, not an impulsive function. Therefore in calculating the climatological average the abrupt change of zonal wind stress is smoothed, but its impact on the upper ocean heat content will remain. To examine the impact of an abrupt change of zonal wind stress on the ocean, the annual cycle of zonal wind stress is artificially enhanced by a patch of westerly/easterly wind stress anomaly. This additional wind patch does not change the pattern evolution of the annual cycle of zonal wind stress significantly. It only makes the relaxation/intensification of the annual cycle in the spring and summer become more rapid. Figure 18 shows the distribution of additional westerly/easterly wind stress anomaly, its ocean response, and the resulting annual cycle of ocean response. The eastward and westward propagation of $h$ perturbation emanating from the central basin starting in early summer is presented. This is mainly caused by the abrupt intensification of easterly wind stress which generates an upwelling Kelvin wave propagating to the east. Because of the timing the abrupt relaxation of easterly wind modulates the amplitude of $h$ but not the evolution of pattern. The complexities of the annual cycle of the upper
ocean heat content are highly possibly related with the details of the zonal wind stress. In addition to the abrupt change of zonal wind stress, higher vertical modes neglected herein may also add to the complexity of the upper ocean heat content annual cycle [e.g., Busalacchi and Cane, 1985]; however, we emphasize that the evolution of even a single vertical mode, owing to the spatial and temporal modulation of the forcing function, may be very complicated.

In summary, the model zonal wind stress of ECMWF in the equatorial Pacific Ocean from 1985 to 1992 was used to study the basin-wide variations of zonal wind stress. The mean state, climatological mean annual cycle and perturbations, and anomalies were examined. The mean state is dominated by the easterly wind stress. The amplitude is largest around 140°W, decreasing linearly to both sides of the basin. The westerly wind stress is confined west of the date line and within the eastern boundary of the basin. A westward propagation of annual wind fluctuations was observed in the climatological mean annual perturbations. This westward propagation of perturbation was modulated by semiannual fluctuations. The wind stress anomalies indicate that the interannual fluctuation of zonal wind stress propagates eastward. A CEOF analysis was employed to study the interannual, annual, and semiannual zonal wind stress variations individually. The interannual variation of zonal wind stress is dominated by an eastwardly propagating wave whose wave period, length, and phase speed are 4.9 years, $3.67 \times 10^4$ km, and 24 cm s$^{-1}$, respectively. The wave amplitude varies with time and space. It is largest in the west-central basin during the El Niño/La Niña years. The annual cycle of zonal wind stress is adequately described by the annual and semiannual harmonics. The annual harmonic is westward-propagating and has relative maxima being out of phase to the west and east of the west-central Pacific. The wave period, length, and phase speed are 1 year, $1.71 \times 10^4$ km, and 54 cm s$^{-1}$, respectively. The semiannual harmonic is largest over the western half of the basin where it appears to develop uniformly without propagation.

Using a linear analytical ocean model, the pattern evolution of ocean responses with respect to the various constituents of zonal wind forcing was studied individually. The interannual variability of ocean response is relatively straightforward. Pattern evolution is observed over the region where the interannual zonal wind stress variations occur, while to the east and west the pattern develops without propagation. This can be explained as a forced response to slowly varying winds. The model response of $h$ agreed well with the upper ocean heat content anomaly obtained from the TAO array data. The annual variations of ocean response are much more complicated since the pattern evolution is dependent upon the relatively rapid variations in both space and time. The annual variations in upper ocean heat content, obtained from the historical XBT data, show both eastward and westward propagation emanating from the east-central Pacific. The details are critically dependent upon the spatial and temporal modulation of the easterly wind stress forcing. The westward propagation of upper ocean heat content was mainly caused by the annual harmonic zonal wind stress which propagated westward. It was modulated by the eastward propagating Kelvin waves forced by the semiannual variation of zonal wind stress. Although the eastward propagation of upper ocean heat content can be explained partially by such an annual cycle of zonal wind stress, it is mainly related to the abrupt relaxation/intensification of easterly wind stress. In a climatological average this abrupt change of zonal wind stress is smoothed, but its impact on the ocean remains since the ocean response is an integrated response. Therefore the rapid change of zonal wind stress is critical for the evolution of annual upper ocean heat content. Considering the resonant forced Rossby wave only is not enough to properly describe such complex annual thermal evolution.

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