Wind Stress Variations and Interannual Sea Surface Temperature Anomalies in the Eastern Equatorial Pacific

XUEBIN ZHANG
School of Oceanography, University of Washington, Seattle, Washington

MICHAEL J. McPHADEN
NOAA/Pacific Marine Environmental Laboratory, Seattle, Washington

(Manuscript received 16 February 2005, in final form 2 August 2005)

ABSTRACT

Vertical advection of temperature is the primary mechanism by which El Niño–Southern Oscillation (ENSO) time-scale sea surface temperature (SST) anomalies are generated in the eastern equatorial Pacific. Variations in vertical advection are mediated primarily by remote wind-forced thermocline displacements, which control the temperature of water upwelled to the surface. However, during some ENSO events, large wind stress variations occur in the eastern Pacific that in principle should affect local upwelling rates, the depth of the thermocline, and SST. In this study, the impact of these wind stress variations on the eastern equatorial Pacific is addressed using multiple linear regression analysis and a linear equatorial wave model. The regression analysis indicates that a zonal wind stress anomaly of 0.01 N m$^{-2}$ leads to approximately a 1°C SST anomaly over the Niño-3 region (5°N–5°S, 90°–150°W) due to changes in local upwelling rates. Wind stress variations of this magnitude occurred in the eastern Pacific during the 1982/83 and 1997/98 El Niños, accounting for about 1/3 of the maximum SST anomaly during these events. The linear equatorial wave model also indicates that depending on the period in question, zonal wind stress variations in the eastern Pacific can work either with or against remote wind stress forcing from the central and western Pacific to determine the thermocline depth in the eastern Pacific. Thus, zonal wind stress variations in the eastern Pacific contribute to the generation of interannual SST anomalies through both changes in local upwelling rates and changes in thermocline depth. Positive feedbacks between the ocean and atmosphere in the eastern Pacific are shown to influence the evolution of the surface wind field, especially during strong El Niño events, emphasizing the coupled nature of variability in the region. Implications of these results for understanding the character of event-to-event differences in El Niño and La Niña are discussed.

1. Introduction

The El Niño–Southern Oscillation (ENSO) phenomenon is the most prominent year-to-year climate variation on the planet. Although it originates in the tropical Pacific, ENSO extends its influence to the globe through the atmospheric teleconnections (Trenberth et al. 1998; Alexander et al. 2002). Sea surface temperature (SST) anomalies in the tropical Pacific play a crucial role in air–sea feedback during the ENSO cycle (e.g., Neelin et al. 1998). The physical mechanisms responsible for giving rise to these anomalies have therefore been an area of intense interest over the past two decades.

Vertical advection of temperature is the primary mechanism by which ENSO time-scale SST anomalies are generated in the eastern equatorial Pacific. Thermocline displacements in the presence of mean upwelling alter the temperature of water upwelled to the surface (Zebiak and Cane 1987; Battisti and Hirst 1989). This “thermocline feedback” (e.g., Jin and An 1999) results primarily from remotely wind-forced thermocline displacements associated with equatorial wave dynamics and is a key element of most ENSO theories. The effects of local wind variations on interannual SST variability in the eastern equatorial Pacific have received relatively little attention because of the prominence of remote wind forcing in generating ENSO SST anomalies. However, winds in the eastern Pacific vary

Corresponding author address: Michael J. McPhaden, NOAA/Pacific Marine Environmental Laboratory, 7600 Sand Point Way NE, Seattle, WA 98115.
E-mail: michael.j.mcphaden@noaa.gov

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on ENSO time scales as well (Rasmusson and Carpenter 1982; Harrison and Larkin 1998) and several studies have noted that wind variations in the equatorial cold tongue region can have an impact on the evolution of individual El Niño events. Among these are the 1982/83 event (Harrison 1989), the 1991/93 event (Kessler and McPhaden 1995a), the 1997/98 event (McPhaden 1999; McPhaden and Yu 1999), and the 2002/03 event (McPhaden 2004; Vintzileos et al. 2005).

A comparison of the 1997/98 El Niño and the 2002/03 El Niño illustrates the point (Fig. 1). During the 1997/98 El Niño, by some measures the strongest on record (McPhaden 1999), westerly anomalies penetrated into the eastern Pacific (east of 140°W) by late 1997. In contrast, during the moderate intensity 2002/03 El Niño, anomalously strong easterlies were present east of 160°W throughout the event. SST anomalies were significantly weaker and the thermocline depth (as indicated by the depth of the 20°C isotherm) was shallower in 2002/03 than in 1997/98. These differences can in large part be ascribed to differences in the intensity of remote wind forcing in the central and western Pacific during the two events. However, the longitudinal structure of the SST and thermocline depth anomalies were very different for the two events, with relatively weak SST and thermocline depth anomalies in the eastern Pacific where the winds were anomalously strong during the 2002/03 event.

The purpose of this study is to systematically evaluate the role of wind stress forcing in the eastern equatorial Pacific on ENSO time scales. Our premise is that while remote forcing is a dominant feature common to all ENSO warm and cold events, winds in the eastern Pacific vary from event to event and can contribute to the unique characteristics that distinguish one event from another. Thus, while we do not intend to challenge the primacy of remote forcing, we want to quantify the relative importance of local versus remote wind forcing during those periods when wind variations in the eastern Pacific are pronounced.

The paper is organized as follows. In section 2, we discuss the datasets and data processing procedures used in this study. Section 3 discusses the importance of remote wind forcing and thermocline feedback in the eastern Pacific. In section 4 we develop and analyze a physically based multiple linear regression (MLR) statistical model that relates thermocline depth and zonal wind stress to SST. Section 5 describes a linear equatorial wave model designed to examine the relative importance of wind stress variations in the eastern Pacific on thermocline depth. Additional interpretation of these results is presented in section 6. We then conclude with a summary in section 7.

2. Data and processing

We use a gridded subsurface temperature dataset derived from an ocean subsurface analysis system (Smith 1995a,b) prepared by the Australian Bureau of Meteorology Research Centre (BMRC). The subsurface temperature dataset has 14 levels in the upper 500 m with 1° latitude × 2° longitude horizontal resolution and monthly temporal resolution starting from 1980. The BMRC dataset combines XBT and mooring data by optimal interpolation. The Tropical Atmosphere and Ocean (TAO) Array (Hayes et al. 1991; McPhaden et al. 1998) provides most of the subsurface temperature information for the BMRC dataset for the equatorial band between 5°S and 5°N during this time period (Smith and Meyers 1996). Meinen and McPhaden (2000, 2001) previously used this dataset to analyze warm water volume changes and volume transports in the equatorial Pacific. The BMRC subsurface temperature was used to derive 20°C isotherm depth (Z20) by linear interpolation in the vertical direction. Anomalies of Z20 and subsurface temperature were calculated relative to a monthly climatology with a base period ranging from 1980 to 2000.

The SST used in this study is derived from blended in situ and satellite analyses of Reynolds and Smith (1994). SST anomalies were computed relative to a monthly climatology for a 1971–2000 base period constructed following Reynolds and Smith (1995) and Smith and Reynolds (1998).

We used twice daily wind field data from the European Centre for Medium-Range Weather Forecasts (ECMWF) 40-yr reanalysis (information available online at http://data.ecmwf.int/data/era40_daily) available from 1 September 1957 to 31 August 2002. After August 2002, the ECMWF operational product was used. We computed daily zonal wind stress from wind velocities with a constant drag coefficient of $1.43 \times 10^{-3}$ and air density of 1.225 kg m$^{-3}$ (Weisberg and Wang 1997). Wind stress anomalies were computed relative to a monthly climatology for a base period of 1971–2000, consistent with that for SST.

We also make use of the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) data product, which is a combination of observations from rain gauges and several satellite sensors. Monthly data are provided with a spatial resolution of 2.5° latitude × 2.5° longitude for the period from 1979 to the present. More information can be found online about this product (http://www.cpc.ncep.noaa.gov/products/global_precip/html/wpae.cmap.html). We compute
Fig. 1. (a) Time–lon plot for interannual anomalies of zonal wind stress ($\tau^x$), 20°C isotherm depth ($Z_{20}$), and SST during the 1997/98 El Niño event. (b) Same as in (a) but for the 2002/03 El Niño event.
precipitation anomalies relative to a monthly climatology with a base period of 1979–2000.

The base periods for the BMRC subsurface temperature climatology and CMAP precipitation climatology are shorter than for winds and SST, but results are not qualitatively different if a base period of 1980–2000 is used for all variables. Prior to further analysis, monthly anomalies around the climatologies are smoothed with a double 5-month running-mean filter to emphasize interannual variability.

3. The importance of remote forcing and thermocline feedback in the eastern Pacific

The importance of remote forcing on interannual time scales in the eastern Pacific is underscored by the spatial patterns of zonal wind stress (hereafter \( \tau^x \)), SST, and \( Z_{20} \) variability (Fig. 2). Zonal wind stress fluctuations are on average largest between 160°E and 160°W in the central and western Pacific (Fig. 2a), whereas \( Z_{20} \) variations along the equator are largest between about 95° and 140°W (Fig. 2b). Interannual variations in SST are also largest east of the date line along the equator and between 0° and 10°S off the South American coast (Fig. 2c).

Time series of SST and \( Z_{20} \) anomalies in the Niño-3 region (5°N–5°S, 160°E–160°W) and zonal wind stress anomalies over the central Pacific (5°N–5°S, 160°E–160°W). (b) Lagged correction between \( \tau^x \) and Niño-3 SST anomalies along the equator. Calculations are based on first averaging \( \tau^x \) data into 10° lon \times 10° lat boxes in the equatorial region between 5°N–5°S and 15°E–90°W. Negative (positive) time lags mean \( \tau^x \) leads (lags) Niño-3 SST. (c) Same as in (a) but for \( \tau^x \) and Niño-3 \( Z_{20} \). Correlations greater than 0.3 are significantly different from zero with 90% confidence based on methods described in Emery and Thomson (2001).

in the central and western Pacific (Fig. 2a), whereas \( Z_{20} \) variations along the equator are least between about 95° and 140°W (Fig. 2b). Interannual variations in SST are also largest east of the date line along the equator and between 0° and 10°S off the South American coast (Fig. 2c).

Time series of SST and \( Z_{20} \) anomalies in the Niño-3 region (5°N–5°S, 160°E–160°W) are most significantly correlated with zonal wind stress in the western and central Pacific (Figs. 3a–c). There is a tendency for eastward propagation in the maximum correlation between \( \tau^x \) and both Niño-3 SST and \( Z_{20} \) (Figs. 3b,c) with \( \tau^x \) leading in the west and lagging in the east. This lead–lag relationship highlights the coupled nature of ocean–atmosphere variability associated with ENSO. SST and \( Z_{20} \) anomalies in the cold tongue are generated primarily by remote wind-forced ocean dynamical processes.
However, once thermocline-depth-mediated SST anomalies develop, they can feedback to the atmosphere to reinforce surface wind stress tendencies through the so-called Bjerknes feedback mechanism (Neelin et al. 1998).

Eastward-propagating equatorial Kelvin waves are the mechanism by which zonal winds remotely affect the eastern basin (e.g., Wyrtki 1975; Busalacchi et al. 1983; Kessler and McPhaden 1995b). These waves lead to changes in thermocline depth, which in turn modify SST by changing the temperature of upwelled water. First baroclinic mode Kelvin waves take only about 2 weeks to travel from the date line to 150°W and about 6 weeks to travel from the date line to 90°W. For the time series shown in Fig. 3a, τ20 anomalies in the region 5°N–5°S, 160°E–160°W lag Niño-3 Z20 anomalies by about 1 month and lead Niño-3 SST anomalies by about 1 month. These lags relative to τ20, which are virtually indistinguishable from zero for interannual time-scale fluctuations, are consistent with the combined effects of remote wind-forced Kelvin wave dynamics on Z20 and SST, and the Bjerknes feedback on surface zonal wind stress.

The thermocline feedback referred to in the introduction is evident in Fig. 3a, which shows high correlation between SST and Z20 in the eastern Pacific cold tongue region. Zero-lag correlations between SST and Z20 (Fig. 4a) are >0.6 east of 150°W and >0.9 east of 110°W where wind variations tend to be weakest. Maximum correlation between SST and Z20 occurs with SST lagging by 1–2 months as can be inferred from visual inspection of Fig. 3a, suggesting a slight delay in the effects of thermocline variations on surface temperatures. However, the difference between zero-lag correlations and those at 1–2-month lag is not statistically significant and for the purposes of this analysis can be ignored. Significant positive correlation between Z20 and SST is also observed in the far western Pacific, but there SST anomalies are very weak (Fig. 2c) because the mean thermocline depth is so much greater in the west than in the east (Fig. 4b). It is the shallowness of the mean thermocline in the east as a result of mean easterly trade wind forcing that makes the eastern equatorial Pacific especially susceptible to thermocline feedbacks on SST.

The above discussion underscores the importance of remote zonal wind forcing on Z20 and SST in the eastern Pacific and the strong connection between thermocline depth and SST variations there. However, Figs. 2–4 represent statistical averages over many events, obscuring event-to-event differences. Some of these differences relate to wind variations in the eastern Pacific and it is these wind variations and their effects we wish to examine in more detail.

4. A dynamically based statistical model for the eastern Pacific

Wang and McPhaden (2000, 2001; hereafter WM) found that on interannual time scales, the dominant terms in the surface layer temperature balance in the eastern Pacific involved surface heating, vertical entrainment, and mixing. Entrainment is a process largely modulated by variations in vertical velocity (upwelling) in the cold tongue region while surface heat fluxes tend to damp SST anomalies that are dynamically generated.
A simplified mixed layer temperature equation can be used to describe this dominant low-frequency SST balance in the eastern equatorial Pacific (Battisti 1988):

$$T'_t \approx - \delta \bar{w} T'_t - \alpha_t T' - \beta_2 T'_{20}/H - \alpha_2 T'_t,$$

(1)

where the overbar denotes the long-term mean and prime denotes an interannual anomaly, $w$ is the upwelling velocity, $\delta$ is an upwelling efficiency coefficient, $H$ is the mean mixed layer depth, $T'$ is the anomalous SST, $T'_{20}$ is the anomalous subsurface temperature, and $\alpha_t$ is a thermal damping coefficient. The first term on the right-hand side of (1) represents mean vertical advection of the anomalous temperature gradient, while the second term represents the thermal damping effect of the surface fluxes.

The anomalous subsurface temperature $T'_{20}$ is related to anomalous thermocline depth displacements $Z'_{20}$ and is often represented as through a linear constant of proportionality (Battisti 1988). Thus (1) becomes

$$T'_t \approx - \delta \bar{w}(T - T'_{20})/H - \alpha_t T = \alpha^* T,$$

(2)

where $K_z = \delta \bar{w}/H$ and $\alpha^* = \delta \bar{w}/H + \alpha_t$. We have dropped the primes in (2) and hereafter for simplicity. Unless otherwise stated, the unprimed variables below represent interannual anomalies.

In the eastern equatorial Pacific, the mixed layer is much shallower than that in the central and western equatorial Pacific. As a result, the thermal inertia of the mixed layer, though not exactly zero as suggested by the 1–2-month phase lag between SST and $Z'_{20}$ (Fig. 3a), is relatively small compared to terms on the right-hand side of (2). Thus, $T'_t$ can be neglected (Chang 1993; WM). Consequently, (2) with the left side equal to zero represents the thermocline feedback on SST with $T$ and $Z'_{20}$ approximately in phase.

Equation (2) ignores the effects of local wind-induced upwelling along the equator. This process can be readily accounted for since local upwelling velocity is roughly equal to the divergence of the meridional Ekman transport. Mathematically, we may express this relationship as (Kessler and McPhaden 1995a)

$$w = \Delta V_E/\Delta y = (-2\tau/\rho_0\beta y_u)/(2y_u) = -\tau/(\rho_0\beta y_u^2) = -\gamma \tau$$

(3)

where $V_E$ is the meridional Ekman transport per unit width, $\rho_0$ is the density of seawater, $y_u$ is the latitudinal range over which equatorial upwelling is important, $\beta = \delta \bar{w}/\delta y$ is the meridional gradient of the Coriolis parameter, and $\gamma = 1/(\rho_0\beta y_u^2)$ is a proportionality coefficient. Anomalous upwelling of the mean temperature gradient, expressed as $wT'_t$, can then be included in (2) such that

$$0 \approx K_z Z'_{20} - \alpha^* T - \delta \bar{w} T'_t = K_z Z'_{20} - \alpha^* T + K_z \tau,$$

(4)

where $K_z = \delta \gamma T'_t = \delta T'_t(\rho_0 \beta y_u^2)$.

The simplified mixed layer temperature balance (4) inspires a MLR model with thermocline depth and wind stress as independent variables and SST as a dependent variable. The model takes the specific form

$$\hat{T} = C_1 \tau + C_2 Z'_{20} + T_0,$$

(5)

where $\hat{T}$ is the estimated SST, $C_1$ and $C_2$ are regression slopes, and $T_0$ is the intercept. Strictly speaking, we do not expect the physical assumptions underlying (5) to apply west of about 150°W in the central Pacific where the mixed layer is significantly deeper than in the eastern Pacific (Ando and McPhaden 1997), where the thermocline feedback weakens, and where zonal advection becomes more prominent (Gill and Rasmusson 1983; McPhaden and Picaut 1990; Picaut et al. 1996; WM). However, Burgers and Oldenborgh (2003) and Zelle et al. (2004) have derived a linear statistical model for SST anomalies similar to (5) with zonal wind stress as a proxy for zonal advection in the central Pacific. Hence, we will fit this model with data for the period 1982–present over a range of longitudes from 95° to 170°W, interpreting the results in terms of the physics processes appropriate to each region.

Comparisons of observed ($T$) and estimated ($\hat{T}$) SST anomalies along the equator (Fig. 5) indicate that this simple regression model performs reasonably well in representing SST at these locations. The correlation coefficient ($R$) between the observed and estimated SST is uniformly high at around 0.9 throughout the study region (Table 1). In terms of $\tau$ and $Z'_{20}$ individually, both the correlation with and the magnitude of the effect on SST increase toward the west for $\tau$ and toward the east for $Z'_{20}$ (Table 1; Fig. 5). Locally, $Z'_{20}$ and $\tau$ are only weakly correlated with one another (Table 1, last column), so that they represent quasi-independent predictors of SST.

In terms of physical controls, the coefficient $C_2$ (Table 2) indicates that the sensitivity of SST to $Z'_{20}$ variations is, as expected, largest in the east. Con-
versely, the coefficient $C_1$, which indicates the sensitivity of SST to $\tau^x$ variations, is relatively constant at about 60°–70°C (N m$^{-2}$)$^{-1}$ [equivalently to 0.6°–0.7°C (0.01 N m$^{-2}$)$^{-1}$] such that SST is equally sensitive to zonal wind stress variations in the eastern and central Pacific. While wind stress anomalies in the east are most often relatively weak compared to those farther west (Fig. 2), there are times when $\tau^x$ anomalies are large in the eastern Pacific as well (e.g., during the 1982/83 and 1997/98 El Niños). Thus, at these times we might expect a measurable impact of the eastern Pacific winds on SST.

The effect of zonal wind stress on SST in the eastern Pacific is most clearly illustrated by computing the correlation of $\tau^x$ with $T^* = T - C_2Z_{20}$, where $T^*$ is the SST anomaly after removing the effects of thermocline depth variations. For the Niño-3 region, the correlation

![Fig. 5. Anomalies of (a) $C_1\tau^x$, (b) $C_2Z_{20}$, (c) $\hat{T}$, and (d) $T$ from Eq. (5) along the equator for the period from 1980 to 2003.](image)

<table>
<thead>
<tr>
<th>Location (°W)</th>
<th>$R$</th>
<th>$R_z$</th>
<th>$R_Z$</th>
<th>$R_{Z_2}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>170</td>
<td>0.88</td>
<td>0.87</td>
<td>0.05</td>
<td>-0.13</td>
</tr>
<tr>
<td>155</td>
<td>0.88</td>
<td>0.81</td>
<td>0.38</td>
<td>0.03</td>
</tr>
<tr>
<td>140</td>
<td>0.90</td>
<td>0.49</td>
<td>0.71</td>
<td>-0.12</td>
</tr>
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<td>125</td>
<td>0.90</td>
<td>0.23</td>
<td>0.78</td>
<td>-0.28</td>
</tr>
<tr>
<td>110</td>
<td>0.95</td>
<td>0.09</td>
<td>0.90</td>
<td>-0.39</td>
</tr>
<tr>
<td>95</td>
<td>0.92</td>
<td>0.06</td>
<td>0.91</td>
<td>-0.13</td>
</tr>
</tbody>
</table>

Table 1. Correlation coefficients between observed SST and SST predicted by the multiple linear regression model ($R$, second column), between observed SST and SST predicted by zonal wind stress $\tau^x$ alone ($R_z$, third column), between observed SST and SST predicted by thermocline depth $Z_{20}$ alone ($R_Z$, fourth column), and between $Z_{20}$ and $\tau^x$ ($R_{Z_2}$, fifth column). Estimates of 90% confidence limits are shown in parentheses. Values above the 90% significance are in boldface.
is \(0.75\) (greater than the 90\% confidence limit), with positive (negative) zonal wind stress anomaly generally related to the warm (cold) SST anomaly (Fig. 6a). Using standard linear regression analysis, the slope of the regression line \(\pm 90\%\) confidence limits for Niño-3 SST anomalies and \(\tau^*\) averaged over the same region is \(0.87^\circ \pm 0.25^\circ\) \((0.01 \text{ N m}^{-2})^{-1}\) (Fig. 6b), similar to the pointwise regression slopes at \(110^\circ\) and \(140^\circ\)W (Table 2). The slope of the regression line increases to about \(1.16^\circ\) \((0.01 \text{ N m}^{-2})^{-1}\) if we use orthogonal (also referred to as neutral) regression analysis rather than standard linear regression. [See Emery and Thomson (2001) for a discussion of the distinctions between these techniques.] In either case though, this relationship suggests that during the 1982/83 and 1997/98 El Niño events, about \(1^\circ\) \(\text{C} (or 1/3 the total Niño-3 anomaly) was attributable to Niño-3 zonal wind stress variations. In contrast, for the period since 2000, persistently strong easterlies in the eastern Pacific lead to somewhat cooler Niño-3 SST anomalies than would otherwise be expected.

### Table 2. Multiple linear regression slope \(C_1\) and \(C_2\), and the intercept \(T_0\) along the equator. The 90\% confidence limits (based on Emery and Thomson 2001) are also shown.

<table>
<thead>
<tr>
<th>Location (°W)</th>
<th>(C_1) (°C N(^{-1}) m(^2))</th>
<th>(C_2) (10(^{-2}) °C m(^{-1}))</th>
<th>(T_0) (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>170</td>
<td>58.58 ± 6.91</td>
<td>1.22 ± 0.81</td>
<td>28.59 ± 2.33</td>
</tr>
<tr>
<td>155</td>
<td>68.27 ± 8.96</td>
<td>2.70 ± 0.84</td>
<td>26.90 ± 2.85</td>
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<td>140</td>
<td>71.52 ± 11.87</td>
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<td>72.16 ± 15.89</td>
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<td>22.51 ± 3.79</td>
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<tr>
<td>95</td>
<td>71.06 ± 35.94</td>
<td>6.25 ± 0.57</td>
<td>22.68 ± 4.74</td>
</tr>
</tbody>
</table>

5. Linear equatorial wave model

The above analysis indicates that \(Z_{20}\) variations have a dominant effect on SST anomalies in the eastern Pacific, but there are times when the effects of eastern Pacific zonal wind stress \(\tau^*\) variations are nonnegligible. Our interpretation of the direct effect of these eastern Pacific wind stress variations has been in terms of local upwelling [Eqs. (3), (4)]. However, we would also expect these winds to affect thermocline depth and thus SST through \(Z_{20}\) variations. The question of how large these eastern Pacific forced versus remotely forced thermocline depth variations is addressed in this section using a simple linear equatorial wave model.

The model we use is based on the method of characteristics and is developed to simulate the low-frequency wave dynamics in the tropical Pacific. Similar models have been used in previous studies of seasonal (Yu and McPhaden 1999) and interannual variations during the 1997/98 El Niño event (McPhaden and Yu 1999). Only the four gravest baroclinic modes are considered here. For each baroclinic mode, the six gravest meridional equatorial Rossby waves and the equatorial Kelvin wave are included. The only forcing for the linear model is the daily ECMWF zonal wind stress anomaly (section 2). Isopycnal displacements at 125 m...
are extracted from the solutions to represent the $20^\circ$C isotherm depth variations (Kessler and McPhaden 1995b). Results are not sensitive to the choice of isopycnal displacements from other nearby depths (such as 100 or 150 m) to characterize $Z_{20}$ variations.

To examine the relative importance of zonal wind stress variations in the eastern Pacific on $Z_{20}$, we make two model runs. One is with full zonal wind stress anomaly forcing (referred to as the control run). The other is with zonal wind stress anomaly set to zero east of 130°W, then linearly increasing to its full value at 150°W (referred to as the masked run). This masking scheme is based on the lagged correlation between $\tau^w$ and Niño-3 $Z_{20}$ (Fig. 3c). There is a zero correlation between these two quantities centered around 130°W at zero phase lag, while positive (weak negative) correlation can be found to the west (east) of this longitude. Variations in $Z_{20}$ from the control run with full wind stress forcing compare well with those derived from the BMRC data (Figs. 7a,b). Model amplitudes are slightly weaker than those in the observations by 2–4 m (Fig. 8a), but the zonal structure along the equator and the timing of the $Z_{20}$ variations associated with El Niño and La Niña events since 1980 are well captured. Correlation between the modeled and observed $Z_{20}$ is between 0.7 and 0.9 at all longitudes (Fig. 8b).

Variations in $Z_{20}$ for the model run with zonal wind stress masked out in the eastern Pacific are in many respects similar to those of the control run (Figs. 7c, 8a) though amplitudes tend to be weaker. The difference between the control and masked run (Fig. 7d) moreover indicates an out-of-phase relationship between $Z_{20}$ differences in the region east of the transition zone (between 150° and 130°W) and those in the region west of it. The out-of-phase relationship can be explained in terms of wind-forced equatorial wave dynamics. For

![Fig. 7. Interannual variations of $Z_{20}$ (in m) along the equator from (a) BMRC data, (b) the control run, (c) the masked run, and (d) the difference between the control run and the masked run. The contour intervals are 10 m in (a)–(c) and 5 m in (d).](image-url)
example, positive (westerly) zonal wind stress anomalies excite both eastward-propagating downwelling equatorial Kelvin waves and westward-propagating upwelling equatorial Rossby waves, while negative (easterly) wind stress anomalies excite waves of opposite sign. The magnitude of the $Z_{20}$ difference is largest in the eastern Pacific where the standard deviation is 4–6 m between 95° and 120°W compared to control run $Z_{20}$ standard deviations of 16 m (Fig. 8a). Temporally, the largest differences (10–20 m) occur during the strong El Niño events of 1982/83 and 1997/98 (Fig. 7d). During the height of these two El Niños, westerly wind stress anomalies in the eastern Pacific worked in concert with remote westerlies to depress the thermocline in the eastern Pacific. In contrast, during the weaker 2002/03 event, somewhat stronger than normal easterlies in the eastern Pacific (Fig. 1b) opposed the tendency of remote wind forcing to depress the thermocline east of 130°W. Thus, in late 2002 and early 2003, the thermocline was 5–10 m shallower in the eastern Pacific than it would have been based on remote zonal wind forcing alone.

In summary, zonal wind stress variations in the eastern Pacific can affect SST not only directly through changes in local upwelling rates as described in section 4, but also indirectly by changing $Z_{20}$ depths and the efficacy of the thermocline feedback. This latter effect is accomplished by wind-forced equatorial Kelvin waves whose largest effects are felt east of 130°W. The impact of these processes on SST depends on the strength, sign, and fetch of the zonal wind stress anomalies in the eastern equatorial Pacific.

6. Discussion

a. Interpretation of the thermocline feedback

Our analysis reveals that while remotely forced thermocline feedback is the dominant mechanism for generating SST variability on ENSO time scales. In the eastern Pacific, wind forcing in the eastern Pacific can contribute to significant event-to-event differences. Failure to account for these eastern Pacific wind variations can lead to confusion in the interpretation of observed variability. For example, Harrison and Vecchi (2001, hereafter HV01) questioned the effectiveness of the thermocline feedback in generating eastern Pacific SST anomalies based on a correlation analysis between SST and $Z_{20}$ from TAO moored time series data for the period from 1986 to 1998. They found significant correlation between SST and $Z_{20}$ at 0°, 110°W. Thus, VH01 concluded that the thermocline feedback was not operative during El Niño at the latter site. However, they did not consider the effects of local winds on SST in the eastern Pacific.
Here, we use our MLR model to repeat their analysis in order to separate out the effects of zonal wind stress and thermocline depth variations on SST. Whereas HV01 used mooring data, we use the Reynolds SST, BMRC $Z_{20}$, and ECMWF-derived $\tau^*$. However, using exactly the same dataset as HV01, we arrive at essentially the same conclusions.

Following HV01, we first reproduce the scatterplots of $T$ versus $Z_{20}$ (left column in Fig. 9) for $0^\circ$, $140^\circ$W and $0^\circ$, $110^\circ$W. Then we adjust $T$ for local wind effects and make new scatterplots of $T^* = T - C_1\tau_s$ versus $Z_{20}$ (right column in Fig. 9). These latter scatterplots show that correlation increases from 0.71 to 0.87 at $140^\circ$W and from 0.89 to 0.95 at $110^\circ$W when $T$ is replaced by $T^*$. Close inspection indicates that the greatest improvement in correlation comes about when accounting for significant westerly anomalies that occurred during the strong El Niño events of 1982/83 and 1997/98. Adjusting for local winds reduces the scatter between $T^*$ and $Z_{20}$, especially at $140^\circ$W. Thus, we conclude that the thermocline feedback is operative during El Niño events at $140^\circ$W but that at times it can be obscured by local wind stress variations.

b. Positive feedbacks between the ocean and the atmosphere in the eastern Pacific

Both the dynamically based multiple regression model (section 4) and the linear equatorial wave model

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**Fig. 9.** Scatterplots of (left) $T$ vs $Z_{20}$ and (right) $(T - C_1\tau_s)$ vs $Z_{20}$ for $0^\circ$, $140^\circ$W and $0^\circ$, $110^\circ$W (see section 6a). Also shown are the evolutions of the 1982/83, 1986/87, and 2002/03 El Niño events in $(T, Z_{20})$ phase space.
(section 5) consider variability in the eastern equatorial Pacific in terms of atmospherically forced ocean responses. However, in nature, the ocean and atmosphere are coupled so that the ocean not only responds to but also can force the atmosphere. In section 3 we discussed the Bjerknes feedback that operates in the zonal direction along the equator during ENSO events. In this section we focus on coupled ocean–atmosphere interactions in the north–south direction that affect variability in the eastern Pacific.

In their composite analyses, Rasmusson and Carpenter (1982) and Harrison and Larkin (1998) found that anomalous near-equatorial surface wind convergence can develop over the central and eastern equatorial Pacific at the height of El Niño events and linger for several months afterward. In the eastern Pacific, this anomalous convergence is related to equatorward movement of the intertropical convergence zone (ITCZ) in response to anomalously warm equatorial SSTs. Equatorward excursions of the ITCZ are most pronounced typically during late boreal winter and early spring because this is a time of year when El Niño SST anomalies are well developed and equatorial SSTs are seasonally at their warmest (Wallace et al. 1998). The ITCZ is associated with weak and variable winds so that equatorward shifts in the ITCZ, which are especially pronounced during strong events, can lead to a significant weakening of the zonal wind stress in the eastern Pacific. Consequently, equatorial upwelling is reduced, which further enhances the original positive SST anomalies. Vecchi and Harrison (2006) and Vecchi (2006 illustrate these feedbacks in oceanic and atmospheric modeling studies of the 1997/98 El Niño event.

These positive feedbacks imply a lag between initial development of SST anomalies in the eastern Pacific and ITCZ-mediated eastern Pacific zonal wind anomalies. This lag is evident in time series of SST, precipitation, and zonal wind anomalies in the Niño-3 region (Fig. 10a). Niño-3 precipitation anomalies, used here to indicate meridional movements of the ITCZ, generally lag Niño-3 SST anomalies by about 1 month. Westerly wind stress anomalies indicative of a local weakening of the trades tend to develop at a lag of about 1–2 months following the precipitation anomalies, most notably for the strong El Niños of 1982/83 and 1997/98.

Meridional movements of the ITCZ in the Niño-3 region (Fig. 10b) and the zonal wind stress and precipitation anomalies that accompany them (Fig. 10c) show these processes at work during the 1997/98 El Niño. The ITCZ began to migrate equatorward in mid-1997 in response to the development of anomalously warm equatorial SSTs. The maximum southward excursion occurred in January–February 1998 followed by the maximum westerly wind anomalies in February–March 1998. Thus, during development and maturation of the 1997/98 El Niño, positive feedbacks between SST, precipitation, and zonal wind stress in the eastern Pacific helped to amplify and sustain anomalously warm SSTs in the region. The abrupt termination of the El Niño in May–June 1998 (McPhaden 1999) was associated with a reestablishment of the ITCZ at more normal latitudes near 10°N and a substantial reduction in zonal wind stress anomalies near the equator.

c. Effects of meridional wind stress

Our MLR model is inspired by a simple physical model for the eastern Pacific, so some processes have been neglected. For example, we have not considered the effects of interannual meridional wind stress (τy) variations on cold tongue SSTs. Although meridional wind stress can affect the seasonal cycle of SST (Philander and Pacanowski 1981; Mitchell and Wallace 1992), there are reasons to believe that τy is not as important as τx in generating equatorial SST variations in the eastern Pacific on interannual time scales. First, τx variations are significantly weaker than those for τy and are centered near 5°N rather than close to the equator (Fig. 11a). Second, the highest correlation between τx and Niño-3 SST and Z20, for meridional wind stress variations in the central and western Pacific (Figs. 11b,c). The implied remote response to this forcing in the eastern Pacific is not as physically plausible as that for τy. Meridional wind stress variations are less effective in generating low-frequency dynamical ocean responses as can be determined from a simple scaling of the equations of motion (e.g., Cane and Patton 1984) and in particular they do not project efficiently onto eastward-propagating Kelvin waves. Meridional wind stress variations, either local or remote, would moreover tend to excite antisymmetric Z20 variations around the equator in the Niño-3 region. These antisymmetric variations would appear only weakly in a 5°N–5°S average. High negative correlations with τy in the central and western Pacific lagging Niño-3 SST and Z20 by 2–4 months (Figs. 11b,c) argue against τy as a cause of Niño-3 variability. High positive correlations between western and central Pacific τy and Niño-3 SST and Z20 exist at leads of 8–12 months, but it is not obvious what physical processes could account for such correlation. The most likely explanation for these high positive and negative correlations is that they are an artifact of the correlation between τx and τy along the equator (Fig. 11d).


\[ \frac{T}{H} \]

There are other processes that we have neglected to explicitly consider in our simple model [Eqs. (4) and (5)]. For example, we have not considered nonlinearity, and in particular the term \( \dot{w} T \) (primes reinstated for clarity). Battisti and Hirst (1989) found that nonlinearity in their coupled atmosphere–ocean model acted to bound the amplitude of ENSO oscillations, but that the essential dynamics of interannual variability was linear. Jin et al. (2003) and An and Jin (2004) later argued that nonlinear vertical advection can amplify El Niño warming, but damp La Niña cooling, thus introducing asymmetry into the ENSO cycle. Nonlinear effects should be most evident in our data during the strong El Niño events of 1982/83 and 1997/98. However, determining these effects with confidence is difficult with the relatively short records available and the methodologies we have chosen. Hence, we defer examination of nonlinear effects in the temperature balance of the eastern equatorial Pacific to a follow-up study as discussed in section 7.

Anomalous zonal wind stress can also drive anomalous zonal current variations, which results in anomalous advection of SST (WM; Picaut et al. 1996). Anomalous wind stress may also lead to anomalous
evaporative heat loss (Burgers and Oldenborgh 2003) and turbulent mixing (WM). Some of these effects may be represented implicitly in our model through $H_{9270}$.

Thus, while explicit inclusion of these wind-related processes could affect the detailed interpretation of our results, our basic conclusion that wind forcing in the eastern Pacific has important and measurable effects on SST would most likely not change.

7. Summary and conclusions

In this paper, we have investigated the effects of wind anomalies in the eastern Pacific on the evolution of interannual $Z_{20}$ and SST variations for the period 1980–2003. In most ENSO theories, wind stress variations in the eastern Pacific are regarded as negligible on ENSO time scales compared to remote wind forcing in the western and central Pacific. We have found though that zonal wind stress variations in eastern Pacific can in some circumstances lead to significant SST anomalies. Moreover, at those times when substantial zonal wind stress anomalies develop in the eastern Pacific there can be both a direct effect on SST through local changes in upwelling and an indirect effect on SST through changing thermocline depth and the efficacy of
the thermocline feedback. Since 1980, the most prominent periods of local wind influence in the eastern Pacific were during the strong 1982/83 and 1997/98 El Niño events when westerly anomalies developed across the basin. During these events, local and remote forcing worked in concert to strengthen warm anomalies at the height of the El Niños. In contrast, stronger than normal easterlies in the eastern Pacific in late 2002–early 2003 tended to both elevate the thermocline and produce more local upwelling in the eastern Pacific. These tendencies may have contributed to the concentration of the largest warm SST anomalies in the central Pacific during the 2002/03 El Niño (Fig. 1b).

The results of this study are based on a simple linear statistical model and a wind-forced linear equatorial wave model. We have restricted our attention to the period 1980–2004 because this is a period particularly rich in oceanic data due to programs like the Tropical Ocean Atmosphere (TOGA) program (McPhaden et al. 1998) and the Climate Variability and Predictability (CLIVAR) program (information available online at http://www.clivar.org/). However, in a follow-up study we will apply the methodology we have established for examining local versus remote forcing effects in the eastern Pacific to ocean reanalysis products (e.g., Carton et al. 2000) extending back in time before 1980. Though the data constraining these reanalyses are sparser than for the period following 1980, it will be of interest to know whether any decadal changes in the character of eastern Pacific wind influences on ENSO can be detected as might be expected from studies such as Harrison and Larkin (1998). The reanalyses can also be used for an explicit dynamical diagnosis of the mixed layer temperature balance, including nonlinear effects, in the cold tongue region. In parallel we will perform a comprehensive evaluation of local versus remote forcing effects in the eastern Pacific analogous to those described in section 5, but for simulations with a wind-forced ocean general circulation model.

Finally, as described in section 6b, the ocean and the atmosphere are coupled in the eastern Pacific so that SST can feedback to the atmosphere to help generate dynamically significant ENSO time-scale wind stress variations. Better understanding of these coupled processes is necessary to determine how variability in the eastern Pacific affects the ENSO cycle. Using a simple coupled ocean–atmosphere model, An and Wang (2000) found that zonal wind stress variations in the eastern Pacific can modulate the amplitude of ENSO events through changes in local upwelling and through wind-forced equatorial Kelvin waves acting on thermocline depth. These results are in qualitative agreement with our analyses and motivate further investigation using coupled models to help elucidate in greater detail the mechanisms by which ocean–atmosphere feedbacks in the eastern Pacific may contribute to ENSO time-scale variations.

Acknowledgments. We thank Charlie Eriksen, Lu-anne Thompson, Paul Quay, and Mike Wallace of the University of Washington, for their advice and encouragement. Three anonymous reviewers provided helpful comments. This research was supported by NOAA’s Office of Oceanic and Atmospheric Research and Office of Global Programs.

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