

NOTES AND CORRESPONDENCE

An Examination of Some ENSO Mechanisms Using Interannual Sea Level at the Eastern and Western Equatorial Boundaries and the Zonally Averaged Equatorial Wind

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ABSTRACT

Theory suggests that the interannual equatorial Kelvin wave sea level (η_W) in the western Pacific should be directly proportional to the interannual western Australian sea level, and the interannual equatorial Kelvin wave sea level (η_E) in the eastern Pacific should be directly proportional to the equatorial interannual sea level at the South American coast. Forced equatorial Kelvin wave dynamics implies, approximately, that

$$\eta_E(t + \Delta t) - \eta_W(t) = B\bar{\tau}(t + \theta\Delta t), \quad (1)$$

where Δt , $\bar{\tau}$, B , and θ , are, respectively, a lag corresponding to equatorial Kelvin wave propagation, zonally averaged equatorial interannual eastward wind stress, a forcing coefficient and a number between 0 and 1. A regression analysis of the sea level difference above using western Australian and South American sea levels against $\bar{\tau}$ shows that the correlation is high and is highest (0.83) when $\Delta t = 3$ months and $\theta\Delta t = 1$ month. The values for Δt , θ , and B are reasonable for they are appropriate for a signal dominated by the first and second vertical modes and forced mainly slightly west of the central equatorial Pacific.

Other relationships between η_W , η_E , and $\bar{\tau}$ are examined in addition to (1) to test some common ENSO ideas. An important component of the delayed oscillator theory of ENSO is that reflected Rossby waves at the western Pacific Ocean boundary should trigger wind changes in the equatorial Pacific, that is, η_W should lead $\bar{\tau}$. Surprisingly, the authors' correlation analysis shows that η_W is not significantly positively correlated with $\bar{\tau}$ at any lag or lead. It seems that while reflection of Rossby waves at the western boundary may influence ocean and atmosphere dynamics in the Pacific interior, extra physics beyond delayed oscillator theory is necessary to describe ENSO. However, we found that other ideas about ENSO were consistent with the data. In agreement with the idea that strengthening trade winds result in a build up of sea level in the western equatorial Pacific and also in agreement with Rossby wave and equatorial wave reflection dynamics, η_W is maximally correlated with $-\bar{\tau}$ when it lags $-\bar{\tau}$ by the theoretically reasonable value of about 3 months. In accordance with the notion that El Niño results when there is a relaxation of the trade winds, η_E is highly correlated with $\bar{\tau}$. But analysis also shows that the commonly held idea that η_E should lag $\bar{\tau}$ because of the equatorial Kelvin wave propagation from the main region of forcing in the west-central equatorial Pacific is questionable because of the influence of the out of phase reflected equatorial Kelvin wave signal η_W .

1. Introduction

Over the last several years much has been learned about El Niño and the Southern Oscillation. Probably the most widely discussed theory for this oscillation is the "delayed oscillator" coupled ocean-atmosphere theory first suggested by Schopf and Suarez (1988), Battisti (1988), Suarez and Schopf (1988), and Battisti and Hirst (1989). An important com-

ponent of this mechanism is that the equatorial Kelvin wave signal resulting from the reflection of Rossby waves at the western Pacific Ocean boundary triggers ENSO events. If this mechanism were operating, then, in terms of sea level, the interannual equatorial Kelvin wave sea level signal at the Western Pacific boundary should lead interannual atmospheric changes in the winds in the equatorial Pacific. Does it? We will test this idea and other ENSO ideas using the interannual zonally averaged equatorial eastward wind stress [$\bar{\tau}(t)$], the interannual equatorial Kelvin wave sea level signal [$\eta_W(t)$] at the equatorial western boundary, and interannual equa-

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torial Kelvin wave signal [$\eta_E(t)$] at the eastern equatorial boundary.

In the next section we will establish theoretically a relationship between η_W , η_E and $\bar{\tau}(t)$ and explain how these quantities can be determined. Following this, in section 3 we will present correlation results to check this theoretical relationship. In section 3 we will also correlate $\bar{\tau}$ and η_W at various lags to test whether the delayed oscillator mechanism is operating, and we will use η_W , η_E , and $\bar{\tau}$, to test some classical ENSO ideas originally due to Wyrtki (1975) and Godfrey (1975). Concluding remarks then follow in section 4.

2. Theory

a. Relationship between η_W , η_E and $\bar{\tau}$

In standard fashion assume that the stratified ocean is of constant depth and describe the ocean response to large scale low-frequency wind forcing in terms of vertical modes. For each vertical mode the linear ocean response to large-scale, low-frequency wind forcing is in terms of a forced long equatorial Kelvin wave and forced long equatorial Rossby waves (e.g., see Gill and Clarke 1974; Cane and Sarachik 1976; or a concise derivation in Clarke and Liu 1993). The equatorial Kelvin wave contribution to sea level for vertical mode j is of the form

$$\eta_j(x, t) \exp(-\beta y^2/2c_j),$$

where x , y , t , β , and c_j refer, respectively, to distance eastward, distance northward from the equator, time, the northward gradient of the Coriolis parameter, and the Kelvin wave phase speed for vertical mode j . The vertical mode j sea level signal at the equator, $\eta_j(x, t)$, satisfies a forced first-order wave equation, which when integrated from $x = 0$, the intersection of the western Pacific boundary with the equator ($y = 0$), to $x = L$, the intersection of the eastern Pacific boundary with the equator, gives

$$\begin{aligned} \eta_j(L, t + \Delta t_j) \\ = \eta_j(0, t) + B_j L^{-1} \int_0^L \tau(x, t + x/c_j) dx. \end{aligned} \quad (2.1)$$

In (2.1)

$$\Delta t_j = L/c_j \quad (2.2)$$

is the time it takes the mode j equatorial Kelvin wave to cross the ocean from the western to the eastern boundary, τ is the eastward component of the equatorial wind stress and B_j , the forcing coefficient for vertical mode j , is given by

$$B_j = (\sqrt{2}\rho_0 g)^{-1} L \int_{-H_{\text{mix}}}^0 F_j dz / \left[H_{\text{mix}} \int_{-H}^0 F_j^2 dz \right]. \quad (2.3)$$

In Eq. (2.3), $F_j(z)$ is the vertical mode j eigenfunction, ρ_0 the mean density of the water, H the total water depth, and H_{mix} the water depth where the wind-driven stress vanishes. In deriving (2.2) and (2.3), we have normalized $F_j(z)$ by taking it to be one at the ocean surface and have assumed that τ varies only slowly over the north-south equatorial Kelvin wave trapping scale (in practice 5°S to 5°N or less for the dominant vertical modes).

When there is no forcing, $\tau = 0$ and (2.1) states that a Kelvin wave signal seen at the eastern boundary at time $t + \Delta t_j$ is equal to the Kelvin wave signal seen at a time t . This is what is expected from free Kelvin wave propagation. When wind forcing is included, the equatorial Kelvin wave signal at the eastern boundary is a sum of that due to free equatorial Kelvin wave propagation from the western boundary and a part forced by the wind as the equatorial Kelvin wave crosses the basin.

On interannual time scales, sea level in the equatorial Pacific is well approximated by a sum of only vertical modes one and two (Cane 1984; Busalacchi and Cane 1985). Using this approximation and Taylor series expansion, we show in appendix A that modes one and two in (2.1) may be summed to give

$$\eta_E(t + \Delta t) = \eta_W(t) + B\bar{\tau}(t + \theta\Delta t), \quad (2.4)$$

where

$$\Delta t = \frac{1}{2} (\Delta t_1 + \Delta t_2), \quad (2.5)$$

$$B = B_1 + B_2, \quad (2.6)$$

and θ is a number between 0 and 1 (probably $\approx 1/3$). For the equatorial Pacific, the times Δt_1 and Δt_2 for the vertical modes 1 and 2 equatorial Kelvin waves to cross the basin are 2.44 and 3.98 months (mo), respectively. Hence we expect $\Delta t \approx 3.2$ mo. To check (2.4) and the expected theoretical values for θ , Δt , and B we must obtain estimates of η_E , η_W and $\bar{\tau}$ from data.

b. Estimation of $\bar{\tau}$, η_W , and η_E

The linear theory of large-scale, low-frequency, wind-forced equatorial ocean response indicates that the only signal that can reach the eastern ocean boundary from the interior is in the form of a forced or freely propagating equatorial Kelvin wave. Therefore, the eastern ocean boundary is a special place where it is possible to easily detect the equatorial Kelvin wave signal. Using sea level records at four locations along the South and Central American coast, Clarke (1992) estimated the equatorial Kelvin wave signal at the intersection of the equator and eastern Pacific Ocean boundary from October 1908. We will use this estimate for $\eta_E(t)$.

Clarke (1991) and du Penhoat and Cane (1991) described how low-frequency energy reflects from the

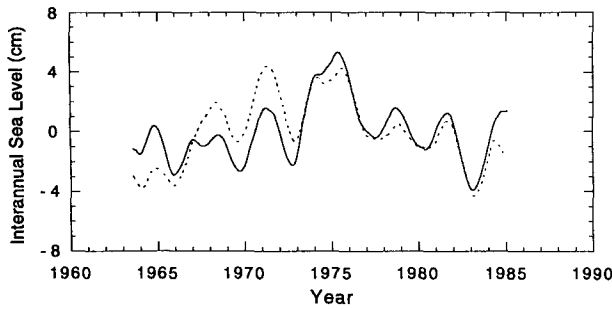


FIG. 1. The $\eta_w(t)$ obtained from (2.7) (solid line) and $\eta_w(t)$ found from a two vertical mode version of the linear forced equatorial ocean model of Cane and Patton (1984) (dashed line). The western boundary condition has been modified using Clarke (1991) to take into account gaps and irregularities in the western boundary. The correlation coefficient between the two time series is 0.76 (95% confidence level is 0.52). The regression coefficient of the model η_w on η_w obtained from (2.7) is 0.86 ± 0.45 (95% confidence interval). The confidence level and interval were found using Chelton (1983).

broken and irregular western Pacific Ocean boundary. The interannual sea level signal on Australia's northwestern coastline ($\eta_A(t)$) can be related to the interannual equatorial Kelvin wave signal at the intersection of the western Pacific boundary and the equator ($\eta_w(t)$). We show in appendix B that

$$\eta_w(t) = 0.52\eta_A(t). \quad (2.7)$$

The time series $\eta_w(t)$ can thus be obtained from the above equation using Clarke's (1991) interannual northwestern Australian sea-level time series for $\eta_A(t)$.

Support for relationship (2.7) has been provided theoretically by Clarke and Liu (1994) and Mantua and Battisti (1994). Clarke and Liu used essentially the physics of (2.7) and a forced long equatorial wave model with Pacific wind stress to simulate quite well the observed northwestern Australian sea level. Mantua and Battisti (1994) found very good agreement between $\eta_w(t)$ obtained from a $1\frac{1}{2}$ -layer linear model forced with Pacific wind stress with $\eta_w(t)$ obtained from (2.7). Figure 1 shows similar good $\eta_w(t)$ agreement for a two vertical mode linear ocean model based on Cane and Patton (1984) and forced by Pacific wind stress. In addition, Clarke (1991) provided observational support for (2.7) by showing that $\eta_A(t)$ correlates well with the interannual sea level record at Truk. Truk is the longest west Pacific sea level record near the equator and approximates the incoming Rossby wave signal and the reflected equatorial Kelvin wave signal $\eta_w(t)$ [see Figs. 16 and 17 of Clarke (1991)].

We calculated $\bar{\tau}(t)$ from The Florida State University monthly mean Pacific "pseudostress" fields generously provided by J. J. O'Brien. Having obtained wind stress from pseudostress with values 1.5×10^{-3} for the drag coefficient and 1.2 kg m^{-3} for the air density, we averaged the eastward windstress component across the basin from 128°E to the South American

coast in the band 1°S to 1°N . After removing mean and trend from the resultant time series, we formed an interannual time series in a similar way to sea level by filtering twice with a 13-month running mean filter (Chelton and Davis 1982).

3. Lagged correlation results

All confidence levels discussed below were found using the analysis of Chelton (1983).

a. Lagged sea level difference across the Pacific and $\bar{\tau}$

We begin by analyzing the theoretically derived relationship (2.4) for time series beginning in July 1963 and ending in January 1985. Correlation calculations at several different monthly lags Δt and $\theta\Delta t$ showed that a maximum correlation of 0.83 occurred when $\Delta t = 3$ months and $\theta\Delta t = 1$ month (see Fig. 2). This correlation is significant at the 95% confidence level

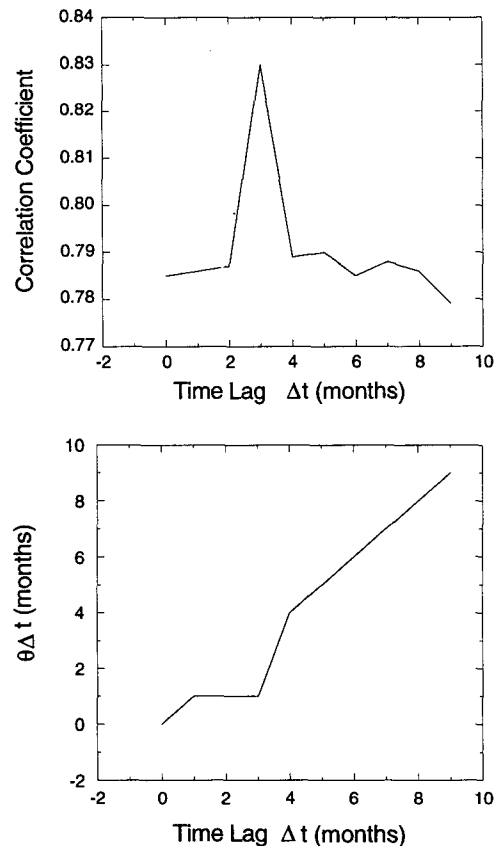


FIG. 2. (a) Maximum correlation between $\eta_E(t + \Delta t) - \eta_w(t)$ and $B\bar{\tau}(t + \theta\Delta t)$ as a function of lag Δt . (b) The values of $\theta\Delta t$ for which the correlation is maximum. Note that the theoretical values are $\Delta t \approx 3$ mo and $\theta\Delta t \approx 1$ mo and these occur at maximum correlation (0.83).

(critical correlation coefficient 0.39). As noted earlier, the theoretical values are $\Delta t \approx 3.2$ months and $\theta\Delta t \approx 1$ month, so the agreement is very good. Because the comparison is between very low-frequency time series, the correlation only falls by about 0.04 from the maximum correlation for nearby values of Δt (see Fig. 2a). However, it seems more than chance that there is a sharp maximum in correlation at the correct theoretical values of Δt and $\theta\Delta t$ and that other lesser correlations do not have Δt and $\theta\Delta t$ values that fit as well (see Fig. 2).

Note that $\theta \approx 1/3$ is due to the forcing being stronger in the western part of the equatorial region (see Fig. 3). If the wind would have been uniform east-west, then $\theta \approx 1/2$ (see appendix A).

From our analysis in appendix A we expect that the regression coefficient B found empirically in (2.4) should approximately match the theoretical value $B_1 + B_2$ [see (2.6) and (2.3)]. Calculations using measured stratification near the date line and $H_{\text{mix}} = 50$ m gave $B_1 + B_2 = 4.72 + 3.00 = 7.72 \text{ m}^2 \text{ s}^2 \text{ kg}^{-1}$. The regression value is $6.82 \text{ m}^2 \text{ s}^2 \text{ kg}^{-1}$ with 95% confidence interval (5.71, 7.93) so the theoretical value lies within the 95% confidence interval. We note that this theoretical value remains inside the 95% confidence interval when H_{mix} varies realistically; the theoretical value hardly changes for $H_{\text{mix}} < 100$ m. Figure 4 shows a comparison of lagged sea level difference with zonally averaged eastward wind stress for the best fit values $B = 6.82 \text{ m}^2 \text{ s}^2 \text{ kg}^{-1}$, $\Delta t = 3$ months and $\theta\Delta t = 1$ mo.

b. η_w and $\bar{\tau}$

Because interannual eastward equatorial wind stress is strongest from about 160°E to 160°W (see Fig. 3), define $\bar{\tau}_e$ to be the zonally averaged interannual eastward wind stress between these longitudes. Note that the correlation of η_w and $\bar{\tau}_e$ should be similar to that of η_w with $\bar{\tau}$ since $\bar{\tau}$ and $\bar{\tau}_e$ are highly correlated (correlation coefficient 0.89) at zero lag (see Fig. 5).

What should a plot of the lagged correlation between η_w with $\bar{\tau}_e$ look like? Consistent with Wyrtki's (1975) idea that strengthening trade winds pile up water in the western Pacific, the correlation of either $\bar{\tau}$ or $\bar{\tau}_e$ with η_w should be negative, with negative wind stress generating and therefore *leading* positive η_w . Eastward equatorial wind stress should similarly generate negative western Pacific sea level and also contribute to the negative correlation.

But this is not the only correlation we should expect. As mentioned in the Introduction, if the delayed oscillator mechanism for ENSO were operating, then the interannual equatorial Kelvin wave signal at the western Pacific boundary should lead interannual changes in the winds in the equatorial Pacific. Since positive η_w corresponds to downwelling and an anomalous eastward current, advecting warm water eastward, we see that positive η_w should result in warmer central Pacific sea surface temperatures and anomalous westerly winds; that is, $+\eta_w$ should lead $+\bar{\tau}_e$. Note that this is a more complicated process than the previous negative correlation one, since it involves both oceanic advection and heating of the atmosphere by anomalous sea surface temperature. Consequently, the positive correlation of η_w leading $\bar{\tau}_e$ could be lower than the negative one of $\bar{\tau}_e$ leading η_w .

Correlation results between $\bar{\tau}_e$ and η_w for D.S. Battisti's version (Battisti 1988) of the Cane and Zebiak coupled ocean-atmosphere model (Cane and Zebiak 1985; Zebiak and Cane 1987) are consistent with the above arguments (see Fig. 6). Here $\bar{\tau}_e$ and η_w correlate strongly with $-\bar{\tau}_e$ leading η_w by about 3 months and also nearly as strongly with η_w leading $+\bar{\tau}_e$ by about 16 months. Because the model is periodic, there are also appropriate correlations at larger lags.

What is the lagged correlation of η_w and $\bar{\tau}_e$ based on observations? Overlapping time series for η_w and $\bar{\tau}_e$ from observations began in July 1963 and ended in January 1985 (see Fig. 7). The model delayed oscillator correlation results for η_w and $\bar{\tau}_e$ are only in partial

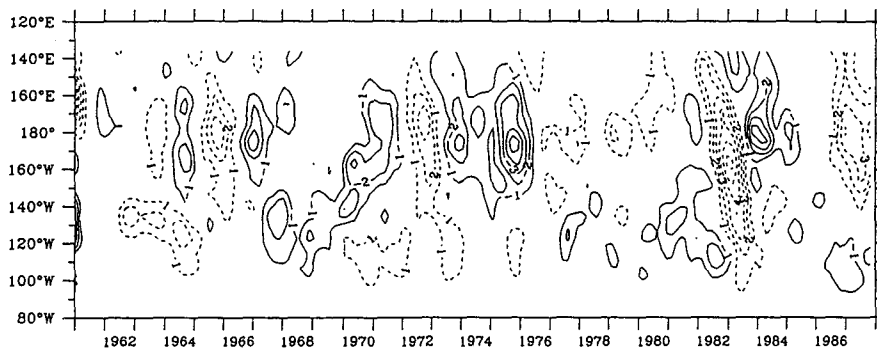


FIG. 3. Anomalies from the average annual cycle of zonal equatorial wind stress (10^{-2} N m^{-2}), with positive (eastward) anomalies dashed and negative (westward) anomalies solid lines (after Kessler 1990).

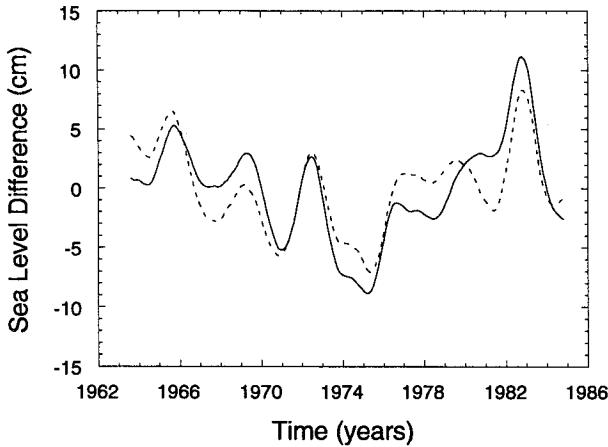


FIG. 4. Equatorial Kelvin wave interannual sea level difference across the Pacific ($\eta_E(t + \Delta t) - \eta_W(t)$) in cm (solid line) and the zonally averaged wind stress times regression coefficient (dashed line) for the optimal values $\Delta t = 3$ months and $\theta\Delta t = 1$ month.

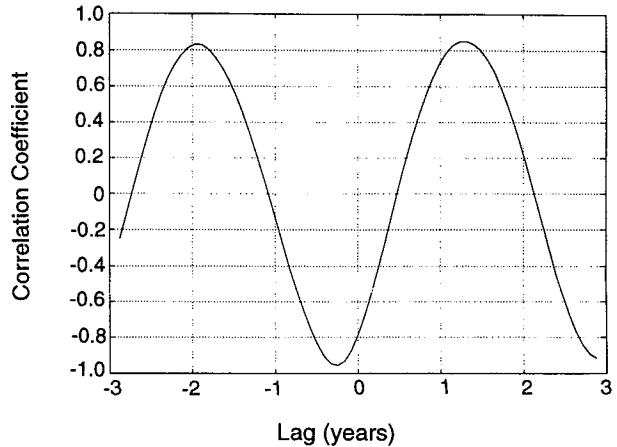


FIG. 6. Lagged correlation of $\bar{\tau}_c$ against η_W with $\bar{\tau}_c$ and η_W from the coupled atmosphere-ocean model of Battisti (1988). Negative lag means η_W lags $\bar{\tau}_c$. [After Battisti (1993, personal communication).]

agreement with similar lagged correlations for the observed time series (see Figs. 6 and 7). They are in agreement in that they are both strongly negatively correlated with $-\bar{\tau}_c$ leading η_W by about 3 months; they are in disagreement in that the observations do not have a significant positive correlation. We will first discuss the negative correlation result to understand dynamically the “pile up” of water in the western Pacific and why it should take 3 months.

When the trade winds strengthen in the central Pacific, the increased westward wind stress generates Rossby waves with westward equatorial currents. Such large-scale waves have westward group velocity and a consequent negative time-averaged energy flux. Since the latter quantity is associated with pressure times eastward velocity, Rossby waves with westward currents will be associated with a positive pressure per-

turbation and higher sea level. These waves propagate to the western Pacific and raise the sea level there. The reflected equatorial Kelvin wave has an eastward equatorial current in order to cancel, as much as possible, the westward equatorial current associated with the Rossby wave. By geostrophy, an equatorial Kelvin wave with an eastward current has raised sea level; that is, η_W is positive. Hence strengthening trade winds lead to raised sea level in the western Pacific, η_W lagging $-\bar{\tau}_c$ by a time corresponding to Rossby wave propagation.

While Fig. 7 shows that η_W and $\bar{\tau}_c$ are most correlated (-0.82 , 95% significance -0.40) when η_W lags $-\bar{\tau}_c$ by about 3 months, this lag is not clearly defined because nearly as strong negative correlation also occurs at nearby phase lags. Based on Rossby wave propagation,

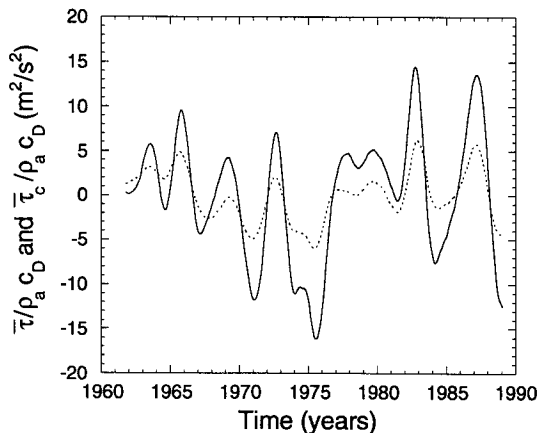


FIG. 5. Wind stress $\bar{\tau}$ (dashed line) and $\bar{\tau}_c$ (solid line) normalized by $\rho_a c_D = 1.8 \times 10^{-3} \text{ kg m}^{-3}$.

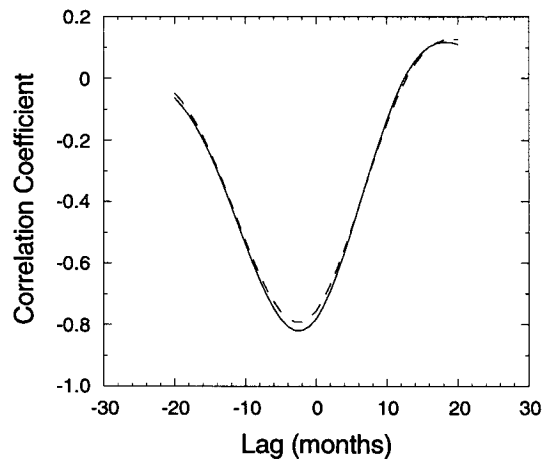


FIG. 7. Correlation of $\bar{\tau}_c$ against η_W (solid line) and $\bar{\tau}$ against η_W (dashed line) as a function of lag. Negative lag means that sea level lags wind stress.

what lag should we expect theoretically? The major part of the equatorial sea level signal reaching the western boundary from the interior is in the form of the first meridional mode equatorial Rossby wave. This wave, having a westward speed $1/3$ of the eastward speed of the equatorial Kelvin wave, takes about 3 months to travel from 180° to the western boundary at 120°E for the first vertical mode and 5 months for the second. An average of these lags is comparable to the approximate three or so month lag seen in Fig. 7.

Note that the above arguments are based on the assumption that the overall contribution to η_W from other Rossby wave signals generated either by the equatorial wind outside of 160°E to 160°W or by reflection at the eastern ocean boundary is minor. Calculations verify that this is indeed the case.

We now turn to the area of disagreement between the expected delayed oscillator η_W and $\bar{\tau}_c$ correlation and that observed. As noted earlier, the observed correlation at an expected lag of about 1 to $1\frac{1}{2}$ years is negligible while that for the Battisti (1988) model is large and positive. Fig. 13 of Schopf and Suarez (1988) suggests that the correlation results for their coupled ocean-linear atmosphere model should also display a strong positive correlation with η_W leading $\bar{\tau}_c$ by about 10 months. Since the observed correlation does not show this large positive peak, the delayed oscillator mechanism would seem to be invalid in nature.

A preprint of our paper prompted Mantua and Battisti (1994) to suggest that although the delayed oscillator theory might be invalid in the strict sense, all but one of its basic elements may still apply. Specifically, they showed that a low positive correlation of η_W and $\bar{\tau}_c$ would result if the equatorial Kelvin wave resulting from reflection at the western Pacific boundary were to terminate central and eastern Pacific ENSO anomalies but often not initiate new ones. We note that Schopf and Suarez (1988) also obtained a much reduced positive correlation (0.2) between η_W and $\bar{\tau}_c$ when they used a nonlinear instead of linear model atmosphere, but along with that the negative correlation (-0.5) was considerably less in magnitude than the -0.82 for the observations (see Fig. 6c of Schopf and Suarez, where their η_R is used as a proxy for η_W , a reasonable approximation based on their Figs. 10a and 10b). The Mantua and Battisti (1994) scenario has the advantage of having a negligible positive correlation peak while retaining a strong negative one.

c. η_E and $\bar{\tau}_c$, η_E and $\bar{\tau}$

One of the most widely quoted ideas concerning El Niño is Wyrtki's (1975) hypothesis that El Niño results when the trade winds, which pile up warm water in the western Pacific, relax. When the trade winds weaken the zonal equatorial sea level slope is reduced and a higher sea level and depressed thermocline are seen in the eastern equatorial Pacific. By a similar argument a lower eastern equatorial Pacific sea level and raised

thermocline should result when the trades strengthen. In both cases the sea level adjustment is accomplished by long equatorial waves. Cane and Sarachik (1981) and Chao and Philander (1993) have pointed out that we may not see individual waves clearly because in general the wind-forced low-frequency ocean response will consist of a sum of many interfering wave signals. However, as noted earlier, the eastern equatorial ocean boundary is one place where the interannual sea level signal depends directly on the equatorial Kelvin wave alone. But even in this case the signal consists of two parts. One part is due to an equatorial Kelvin wave that has propagated freely across the Pacific and has resulted from Rossby wave reflection at the western boundary; the other part is due to an equatorial Kelvin wave forced by the zonal wind stress.

As shown in Fig. 3, the interannual equatorial forcing has a maximum in the central Pacific between about 160°E and 160°W , quite remote from the eastern equatorial Pacific boundary (about 80°W). If there were no western boundary, then η_E is determined directly by this remote forcing and a lag between forcing and boundary sea level can be expected. Since the first and second vertical mode equatorial Kelvin waves travel at about 2.7 m s^{-1} and 1.6 m s^{-1} , respectively, and these dominate interannual sea level and are approximately in the ratio 2:3 (Clarke and VanGorder 1994), we would expect η_E to lag by $\bar{\tau}_c$ by about 2 months.

However, the above mechanism is incomplete for it does not take into account the reflected equatorial Kelvin wave signal η_W at the western boundary. This must affect η_E since it has a similar amplitude ($\text{RMS } \eta_E / \text{RMS } \eta_W \approx 1.4$). As noted in section 3b, η_W lags $\bar{\tau}$ by about 3 months and when this result is substituted into the sea level slope formula (2.4) we find that η_E no longer lags $+\bar{\tau}$ but leads it slightly by about 1 month (see appendix C). A similar result can be expected for $\bar{\tau}_c$ since $\bar{\tau}_c$ and $\bar{\tau}$ are in phase. In summary, although

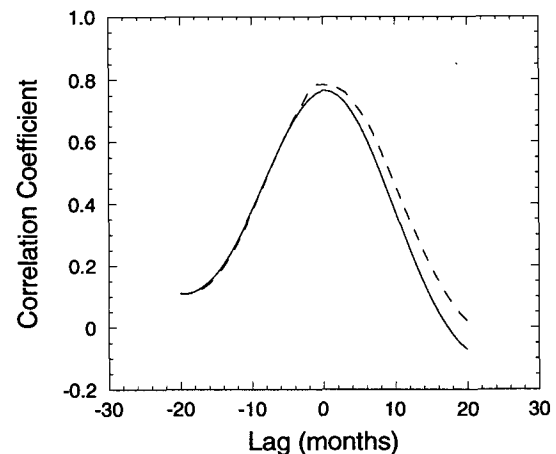


FIG. 8. As for Fig. 7 except with η_E instead of η_W .

eastern boundary sea level is due to equatorial Kelvin wave propagation from the western Pacific, we do not expect to see an appropriate Kelvin wave phase lag because of the interference of two equatorial Kelvin wave signals that are approximately out of phase; one Kelvin wave signal, directly generated by $\bar{\tau}_c$, propagates eastward of the region of forcing while the other, of nearly opposite phase, results from westward Rossby wave propagation from the region of forcing and then subsequent reflection at the western boundary.

What do the lagged correlations of $\bar{\tau}_c$ and $\bar{\tau}$ against η_E actually look like? Figure 8 shows results for the period July 1964 to January 1985. As expected from simple ideas about zonal wind stress causing sea level tilt, η_E is positively correlated with $\bar{\tau}$ and $\bar{\tau}_c$. The maximum correlation between η_E and $\bar{\tau}_c$ is 0.77 (95% sig-

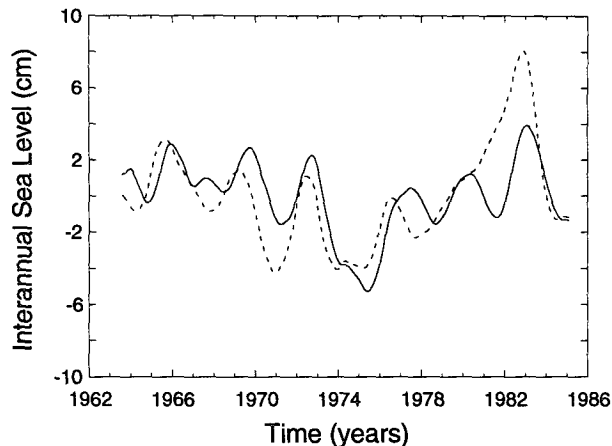


FIG. 10. Sea level boundary signals η_E (dashed line) and $-\eta_W$ (solid line).

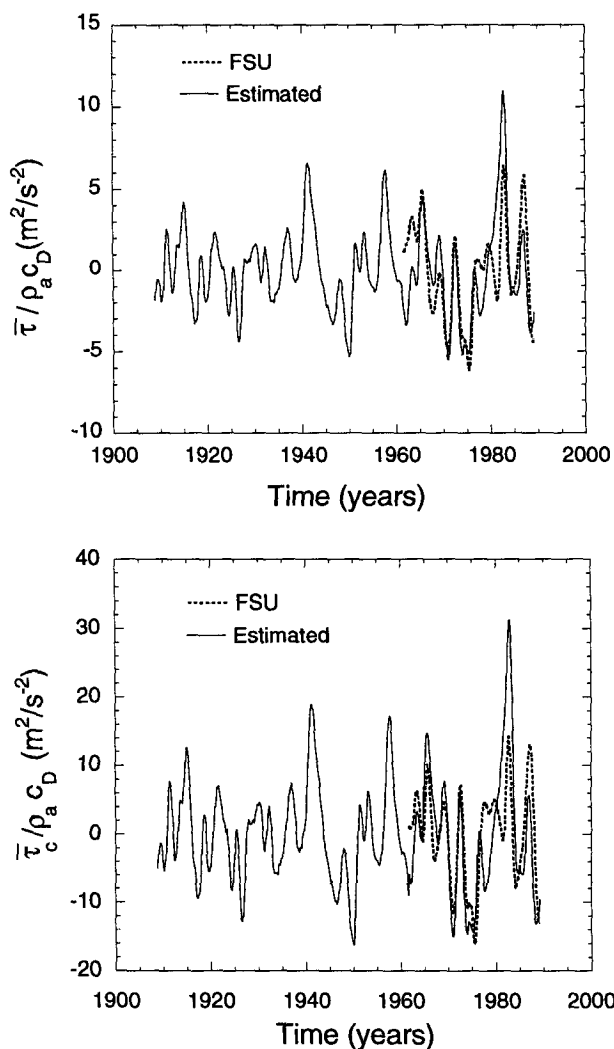


FIG. 9. (a) FSU pseudo stress $\bar{\tau} / \rho_a c_D$ (dashed line) and pseudo stress $\bar{\tau} / \rho_a c_D$ estimated back to October 1908 (solid line) using η_E ; $\rho_a c_D$ takes the constant value $1.8 \times 10^{-3} \text{ kg m}^{-3}$. (b) As in (a) except for $\bar{\tau}_c$ instead of $\bar{\tau}$.

nificance level = 0.39) at zero lag while that for η_E and $\bar{\tau}$ is 0.78 (95% significance level = 0.39) with η_E lagging $\bar{\tau}$ by 1 month. These lags are not clearly defined as nearly as strong correlations occur at nearby time lags; correct to two decimal places, the correlations at 1 month lag, no lag and 1 month lead are all 0.78. The data therefore cannot verify or disprove the lag expected theoretically. The regression coefficients with 95% confidence intervals are $1.69 (\pm 0.58) \text{ m}^2 \text{ s}^2 \text{ kg}^{-1}$ for η_E and $\bar{\tau}_c$ and $4.06 (\pm 1.35) \text{ m}^2 \text{ s}^2 \text{ kg}^{-1}$ for η_E and $\bar{\tau}$.

The above high correlations imply that η_E can be used as an indirect zonal wind stress gauge. This is especially useful prior to about 1960 when the equatorial wind stress measurements are fewer and our knowledge is consequently more unsure. Since η_E is available beginning October 1908, we have an estimate of $\bar{\tau}$ and $\bar{\tau}_c$ back to that time. Figure 9 shows this estimate together with measured $\bar{\tau}$ and $\bar{\tau}_c$ since October 1961.

Finally, note that since η_W is nearly in phase with $-\bar{\tau}_c$ and η_E is nearly in phase with $+\bar{\tau}_c$, we expect that sea levels on opposite sides of the Pacific should be approximately out of phase. Figure 10 shows that this is indeed the case.

4. Concluding remarks

We have used zonally averaged interannual equatorial wind stress and interannual sea level at the eastern and western Pacific Ocean boundaries to examine some commonly held ideas about ENSO. Surprisingly, in apparent contradiction with delayed oscillator theory, η_W is not positively correlated with $\bar{\tau}_c$ at any lead time. Wakata and Sarachik (1991) used a simple numerical model to provide support for the delayed oscillator mechanism, but they did not calculate correlations as in our case. Mantua and Battisti (1994) have suggested that although the delayed oscillator theory might be

invalid in the strict sense, all but one of its basic elements may still apply. Specifically, a negligible lagged positive correlation would result if the equatorial Kelvin wave resulting from reflection at the western Pacific boundary acts to terminate ENSO anomalies but not always to initiate new ones.

We found that other ENSO ideas were mostly consistent with the correlation analysis. The difference in interannual sea level across the Pacific, taking into account the appropriate time lags, is directly related to the zonally averaged eastward equatorial wind stress. Interannual western equatorial Pacific boundary sea level rises (falls) when the trade winds strengthen (weaken) and lags interior wind forcing by about 3 months, a lag appropriate for equatorial Rossby wave propagation. On the other hand, while it is true that interannual eastern equatorial Pacific sea level is driven by the interior zonal equatorial wind stress, we found that the commonly held idea that η_E should lag $\bar{\tau}$ because of equatorial Kelvin wave propagation from the main region of forcing in the west-central equatorial Pacific is questionable. This is because η_E really consists of two interfering equatorial Kelvin wave signals; one is directly driven by $\bar{\tau}_c$ and propagates eastward of the region of forcing, while the other, of nearly opposite phase, results as the forcing generates Rossby waves that propagate westward of the region of forcing and then reflect as an equatorial Kelvin wave. The resultant response has η_E leading $\bar{\tau}$ by about 1 month. The lagged correlation analysis is unable to distinguish clearly between this 1 month lead and η_E lagging $\bar{\tau}$ by 2 months, the expected result when the reflected equatorial Kelvin wave signal is ignored.

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APPENDIX A

Establishing a Relationship between Interannual Equatorial Sea Level Difference across the Pacific and the Zonally Averaged Equatorial Eastward Interannual Wind Stress

Our aim is to establish (2.4). We proceed by first noting that, by Taylor series expansion, for $j = 1$ and 2,

$$\eta_j(x, t + \Delta t_j) = \eta_j(x, t + \Delta t) + (\Delta t_j - \Delta t) \frac{\partial \eta_j}{\partial t}(x, t + \Delta t) + \dots, \quad (\text{A1})$$

where Δt_j and Δt are defined in (2.2) and (2.5) of the main text and η_j is the equatorial Kelvin wave contribution to sea level for vertical mode j at the equator. Equation (A1) can be simplified to

$$\eta_j(x, t + \Delta t_j) = \eta_j(x, t + \Delta t) \quad (\text{A2})$$

with error $\omega(\Delta t_j - \Delta t)$, ω being a representative ENSO frequency. Since $\Delta t_1 = 2.44$ mo, $\Delta t_2 = 3.98$ mo, and $\Delta t = 1/2(\Delta t_1 + \Delta t_2) = 3.21$ mo, for $\omega = 2\pi/3$ yr, the error is small (13%). Substitution of (A2) into (2.1) and adding the $j = 1$ and $j = 2$ equations gives, by the definition of η_E and η_W ,

$$\eta_E(t + \Delta t) = \eta_W(t) + \sum_{j=1}^2 B_j L^{-1} \int_0^L \tau(x, t + x/c_j) dx. \quad (\text{A3})$$

By Taylor series expansion of $\tau(x, t + x/c_j)$ about $\tau(x, t + \theta\Delta t)$ we have, by the definition of $\bar{\tau}$, that (A3) can be written

$$\eta_E(t + \Delta t) = \eta_W(t) + (B_1 + B_2)\bar{\tau}(t + \theta\Delta t) + E, \quad (\text{A4})$$

where

$$E = \sum_{j=1}^2 B_j L^{-1} \int_0^L (x/c_j - \theta\Delta t) \tau_t(x, t + \theta\Delta t) dx + B_j L^{-1} \int_0^L (0.5)(x/c_j - \theta\Delta t)^2 \times \tau_{tt}(x, t + \theta\Delta t) dx + \dots \quad (\text{A5})$$

The desired result (2.4) then follows from (A4) and (2.6) if E is negligible compared with $(B_1 + B_2)\bar{\tau}(t + \theta\Delta t)$. We will prove this below.

First consider the second- and higher-order terms on the right-hand side of (A5). By approximating $\tau_{tt}(x, t + \theta\Delta t)$ by $\bar{\tau}_{tt}(t + \theta\Delta t)$, we deduce that the error in neglecting these terms is of order

$$\sum_{j=1}^2 (1/6) B_j \bar{\tau}_{tt}(t + \theta\Delta t) [(\Delta t_j)^2 - 3(\Delta t_j)(\theta\Delta t) + 3(\theta\Delta t)^2] / (B_1 + B_2) \bar{\tau}(t + \theta\Delta t). \quad (\text{A6})$$

Using the calculated values $B_1 = 4.72 \text{ m}^2 \text{ s}^2 \text{ kg}^{-1}$, $B_2 = 3.00 \text{ m}^2 \text{ s}^2 \text{ kg}^{-1}$, $\Delta t_1 = 2.44$ mo, and $\Delta t_2 = 3.98$ mo, the statistically estimated value $\theta\Delta t = 1$ mo, and approximating $\bar{\tau}_{tt}/\bar{\tau}$ by $-\omega^2$, with $\omega = 2\pi/3$ yr, this error is $0.019 \ll 1$. The first term on the right-hand side of (A5) is also negligible for appropriate choice of θ . To estimate θ , approximate $\tau_t(x, t + \theta\Delta t)$ by $\bar{\tau}_t(t + \theta\Delta t)$.

Then the first term on the right hand side of (A5) is equal to zero when

$$\theta \Delta t = \frac{\sum_{j=1}^2 B_j \Delta t_j}{2 \sum_{j=1}^2 B_j} = 1.52 \text{ mo} \quad (\text{A7})$$

and $\theta \approx 1/2$. If we were to take into account that the interannual wind stress is stronger in the west-central part of the equatorial Pacific (see Fig. 3), then the estimate for $\theta \Delta t$ would be nearer to 1 month and $\theta \approx 1/3$.

APPENDIX B

Establishing a Relationship between Interannual Equatorial Sea Level on Australia's Western Coast and η_W

Clarke (1991) and du Penhoat and Cane (1991) showed that the equatorial Kelvin wave reflected from the gappy, irregular western Pacific Ocean boundary mostly results from the reflection of a first meridional mode equatorial Rossby wave. The dimensional pressure field for such a wave at the western boundary can be written in the form

$$p_R = \sum_{j=1}^2 \alpha_j [2^{-1/2} \psi_2(y_j) + \psi_0(y_j)] F_j(z) c_j^2 \rho_0, \quad (\text{B1})$$

where $\alpha_j(t)$ is the amplitude for vertical mode j , ψ_0 and ψ_2 are Hermite functions, and

$$y_j = y \left(\frac{\beta}{c_j} \right)^{1/2} \quad (\text{B2})$$

with y distance northward from the equator and β the northward gradient of the Coriolis parameter.

For such an incoming Rossby wave pressure field, theory (see Clarke 1991) shows that the reflected equatorial Kelvin wave pressure field is

$$p_K = \sum_{j=1}^2 \gamma_j \alpha_j \psi_0(y_j) F_j(z) c_j^2 \rho_0, \quad (\text{B3})$$

and the pressure field on Australia's northwestern coast is

$$p_A = \sum_{j=1}^2 d_j \alpha_j F_j(z) c_j^2 \rho_0. \quad (\text{B4})$$

In (B3) the reflection coefficients $\gamma_1 = 0.415$ and $\gamma_2 = 0.435$, while the coefficients $d_1 = 0.595$ and $d_2 = 0.581$. Since $p(z=0) = \rho_0 g \eta$, with g the acceleration due to gravity and η the sea level, it follows from (B3) and (B4) and the definitions of η_W and η_A that

$$\begin{aligned} \eta_W / \eta_A &= p_K(y=0, z=0) / p_A(z=0) \\ &= \sum_{j=1}^2 \gamma_j \alpha_j \pi^{-1/4} c_j^2 / \sum_{j=1}^2 d_j \alpha_j c_j^2. \end{aligned} \quad (\text{B5})$$

In obtaining (B5) we used $\psi_0(0) = \pi^{-1/4}$ and $F_j(0) = 1$. Equation (B5) may also be written

$$\eta_W / \eta_A = \pi^{-1/4} \gamma_1 d_1^{-1} [1 + c_2^2 \alpha_2 \gamma_2 / c_1^2 \alpha_1 \gamma_1] / [1 + c_2^2 \alpha_2 d_2 / c_1^2 \alpha_1 d_1]. \quad (\text{B6})$$

But since $\gamma_2 / \gamma_1 \approx d_2 / d_1$,

$$\eta_W / \eta_A \approx \pi^{-1/4} \gamma_1 d_1^{-1} = 0.52 \quad (\text{B7})$$

which is the required result.

APPENDIX C

Estimating the Lag between η_E and Lag $\bar{\tau}$

Consistent with dynamics and the high lagged correlation in Fig. 7, we may write, approximately, for some constant D

$$\eta_W(t) = -D \bar{\tau}(t - \Delta t), \quad (\text{C1})$$

where $\Delta t = 3$ months. Substitution of this result into (2.4) gives

$$\eta_E(t + \Delta t) = -D \bar{\tau}(t - \Delta t) + B \bar{\tau}(t + \theta \Delta t). \quad (\text{C2})$$

We modify the right-hand side of C2 by expanding in a Taylor series that yields

$$\begin{aligned} (B - D)[\bar{\tau}(t) + \alpha \Delta t \bar{\tau}_t(t) + \dots] \\ = (B - D) \bar{\tau}(t + \alpha \Delta t), \end{aligned} \quad (\text{C3})$$

where

$$\alpha = (B \theta + D) / (B - D). \quad (\text{C4})$$

It follows from (C2) and (C3) that η_E should lag $\bar{\tau}$ by $(1 - \alpha) \Delta t$.

Using $\theta = 1/3$ and the regression coefficients $D = 3.14 \text{ m}^2 \text{ s}^2 \text{ kg}^{-1}$ and $B = 6.82 \text{ m}^2 \text{ s}^2 \text{ kg}^{-1}$ gives $\alpha = 1.47$ and $(1 - \alpha) \Delta t = -1.41$ months. Thus η_E should lead $\bar{\tau}$ by about 1.4 months.

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