Can Reflected Extra-equatorial Rossby Waves Drive ENSO?*

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5 December 1989 and 28 June 1990

ABSTRACT

The possibility that the evolution of the ENSO phenomenon is determined by the reflection of extra-equatorial Rossby waves from the western boundary into the equatorial waveguide has been a subject of recent debate. Observations and some wind-driven models suggest an apparent continuity of off-equatorial signals and subsequent waveguide anomalies. On the other hand, coupled model results show that ENSO-like behavior can be simulated with no involvement of the extra-equatorial regions. Linear equatorial wave theory shows that significant reflection can only occur within about 8° of the equator, with a sharp fall-off in the reflectivity poleward of this latitude. Although the amplitude of the thermocline anomalies associated with observed ENSO-forced extra-equatorial Rossby waves can be large, it is the net zonal transport of these waves that is crucial to the reflectivity, and this net transport decreases rapidly as Rossby waves occur farther from the equator. The zonal geostrophic flows associated with observed extra-equatorial Rossby waves in the northern tropical Pacific do not provide a net transport that could make a significant contribution to equatorial Kelvin waves. If the extra-equatorial signals do exert an influence on the equatorial waveguide, it must be through a mechanism other than simple boundary reflection.

1. Introduction

The possibility that reflection of extra-equatorial Rossby waves from the western boundary is responsible for the quasi-periodic nature of the ENSO (El Niño–Southern Oscillation) phenomenon has been hypothesized in two recent papers by Graham and White (1988, 1991). In contrast, coupled model results (Zebiak and Cane 1987; Battisti 1988, 1989, 1991; Schopf and Suarez 1988) show that it is possible to simulate coupled oscillations similar in many respects to the observed ENSO with no involvement of the extra-equatorial regions. The purpose of the present note is to show, from observations and simple theory, that although the wind anomalies associated with ENSO do generate large-amplitude extra-equatorial Rossby waves that propagate to the western boundary (Pazan et al. 1986; White et al. 1987, 1989; Kessler 1989, 1990), the reflection of these waves does not produce significant equatorial Kelvin waves. Thus a simple reflection mechanism such as proposed by Graham and White (1988, 1991) cannot be responsible for triggering ENSO. Although the extra-equatorial waves are important in determining the response of the western extra-equatorial Pacific, they do not appear to significantly affect subsequent evolution on the equator.

McCreary (1983) originally proposed a simple coupled model in which the roughly 4-year timescale of ENSO was determined by the relatively slow propagation of extra-equatorial Rossby waves across the width of the Pacific. In this model, the ocean was forced by idealized patches of anomalous winds that switched on and off in response to changes in eastern equatorial Pacific SST, which was assumed to be simply related to thermocline depth. These wind patches generated alternately upwelling and downwelling extra-equatorial Rossby waves (by Ekman pumping), which in turn provided negative feedback to the equatorial thermocline after reflection from the western boundary. The model oscillated at a period set by the latitude of the extra-equatorial waves, which take approximately two to four years to cross the basin at latitudes between 10° and 15°N. A similar mechanism is at the heart of the process proposed by Graham and White (1988). They suggested that when the eastern equatorial SST is cool, anomalous equatorial easterly winds are generated, which are associated with downwelling wind stress curl in the extra-equatorial regions. This curl lowers the extra-equatorial thermocline, and the downwelling signal propagates westward as a long Rossby wave, reflecting off the western boundary and returning along the equator as a downwelling Kelvin wave which lowers the eastern equatorial thermocline. The deep eastern thermocline then warms the equatorial SST (by moving the source of cold water farther from the surface), thus turning off the easterly winds

* WHOI Contribution Number 7436.

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and triggering an ENSO event. The cycle reverses as the warm eastern SST generates westerlies on the equator. Curl due to the westerlies upwells the extra-equatorial thermocline, and the resulting Rossby wave reflects and "turns off" the El Niño, and the cycle begins again. Graham and White (1988; see their Figs. 6 and 7) suggested that Rossby waves at about 12°N and S were responsible for these phase reversals of the ENSO cycle. Although they do not assume that the waves responsible for the feedback always occur at 12° latitude, allowing for some variation in location, Graham and White (1988) show repeated examples (from models) in which the 12° waves appear to reflect and produce large equatorial signals. This note examines whether the reflection of realistic waves originating far from the equator is likely to be important in the ocean.

The distinction between the near-equatorial Rossby waves and those farther away is important because the Rossby propagation speed is a strong function of latitude. Waves within about 8° of the equator propagate rapidly from their generation region in mid-Pacific to the western boundary, and provide only a short time lag. Such waves are found to be crucial in coupled models of ENSO (e.g., Battisti 1988); however, they cannot account on their own for the 3–4 year timescale of ENSO. On the other hand, if the extra-equatorial waves prove significant, then their slow propagation could be the principal mechanism which sets the ENSO timescale, and this would imply a potential means of predicting the occurrence of ENSO a year or more in advance.

2. Theory

The dynamics of the tropical western boundary reflection process has been extensively discussed by several authors using the equatorial β-plane. Lighthill (1969) showed how time-varying winds in midbasin could generate a varying western boundary current upon the arrival of Rossby waves. Moore (1968) and Moore and Philander (1977) showed that a solution at a western boundary could be written which was explicit, involving only a finite number of reflected short Rossby waves, and a Kelvin or Yanai wave, all of which are more closely trapped to the equator than the incoming wave. The physical basis of this result is that information can only travel equatorward on a western boundary. Godfrey (1975) suggested that Rossby reflection could contribute to ENSO. Anderson and Gill (1975) modeled a suddenly applied wind stress in an inviscid ocean and found that the arrival of Rossby waves produces a thinning western boundary layer. Cane and Sarachik (1977) found that to leading order all of the net incoming mass flux due to a Rossby wave is reflected in the Kelvin mode; this result is the basis for the calculations in this note. Clarke (1983) calculated reflection coefficients for incoming Rossby waves, and found that reflectivity drops off sharply after the first meridional mode Rossby wave. In a useful study, Cane and Gent (1984) clarified the dynamics of the reflection and showed the effect of a non-meridional coast, while McCalpin (1987) extended Cane and Gent's results to realistic boundary geometries (here we consider only a straight, meridionally oriented coast). Clarke's (1983) findings are very similar to calculations reported below. Nevertheless, since the possibility of large effects due to reflection of waves originating far from the equator is still a subject of debate, it seems worthwhile to calculate the amplitude of the height field and zonal current anomalies of the Kelvin waves reflected from realistic extra-equatorial Rossby waves as a means of testing the Graham and White (1988, 1991) hypothesis.

Cane and Sarachik's (1977) result that though the short reflected Rossby waves contribute to meridional mass flux along the boundary they do not transport any mass zonally provides a simple method for calculating the amplitude of a reflected Kelvin wave due to incoming long Rossby waves. It is not necessary to resolve the details of the boundary layer dynamics to understand or calculate the reflectivity. The net zonal transport of a given long Rossby wave, integrated meridionally along some longitude east of the boundary layer, must be equal to the net transport leaving the boundary, which is accomplished entirely in the reflected Kelvin wave. For direct comparison with the results of Graham and White (1988), who used the reduced gravity single-active-layer model driven by observed winds developed by Busalacchi and O'Brien (1980, hereafter referred to as the FSU model), consider a 1½ layer ocean on an equatorial β-plane. The mean upper layer thickness (ULT) H_0 is taken to be 150 m, and the density difference $\Delta\rho/\rho$ is taken to be 5×10^{-3}, so the long gravity wave speed is $c = 2.7$ m s^{-1}, typical of the western equatorial Pacific. The ULT and zonal current anomalies of a baroclinic equatorial Kelvin wave are

$$h_k = Ae^{-(\beta y^2/2c)}, \quad u_k = (g'/c)h_k \tag{1}$$

where A is the amplitude in meters (taken positive for downwelling waves) of the Kelvin wave ULT anomaly at the equator; A may be an arbitrary (low-frequency) function of (x, t). The volume transport of the Kelvin wave, U_k, is given by

$$U_k = \int_{-\infty}^{\infty} \int_{H_0}^{0} u_k dz dy = A[2c^3\pi/\beta]^{1/2} \tag{2}$$

which gives U_k in terms of A and constants $[(2c^3\pi/\beta)^{1/2}$ is about 2.3 Sv m^{-1} (Sv $\equiv 10^6$ m^3 s^{-1})]. The amplitude of the Kelvin wave resulting from a given incoming Rossby wave is found by integrating in latitude and depth over the Rossby zonal velocity to find the net zonal transport U_r, and equating $U_r = -U_k$ in (2) to satisfy the condition of no normal flow.

Before estimating the net zonal transport of incoming Rossby waves from observations, it is useful to calculate the amplitude of the Kelvin wave due to each equatorial meridional (Hermite) mode, for typical Pacific conditions. While we do not expect to see a single meridional mode in the ocean, a particular ULT anomaly projects significantly only onto modes with turning latitudes greater than or equal to the latitude of the anomaly (higher Hermite modes). Since higher modes have weaker reflectivity, the calculation for each mode establishes an upper bound on the possible reflection from near its turning latitude.

Typical tropical Pacific extra-equatorial Rossby waves are modeled by assuming a wave period of one year (little difference in the reflectivity occurs for any period greater than six months), and a maximum ULT anomaly of 20 m for each Rossby mode. A 20 m ULT anomaly is large but not uncommon in the northern tropical Pacific (Kessler 1990). The ULT anomalies for each odd Rossby mode are shown in Fig. 1 [only the odd modes (which have symmetric ULT and zonal velocity) reflect to a Kelvin wave]. The zonal wavelengths [if the waves are assumed sinusoidal in (x, t)], also shown in Fig. 1, are very long, placing these waves close to the nondispersive limit. The 20 m maximum ULT anomaly of each meridional mode is found near its turning latitude, where the modal structure changes from oscillatory to evanescent (Fig. 1). The zonal current anomalies of these modes (Fig. 2) have nearly equal-amplitude lobes of positive and negative velocity, without a maximum near the turning latitude, since the zonal velocity is nearly geostrophic. Although the zonal currents due to the Rossby waves can be large [about 30 cm s^{-1} for the 20 m ULT anomalies (Fig. 2)], the net zonal transport is small relative to that in each lobe. Integrating the Rossby zonal velocity for each mode over the depth of the upper layer and all y, then balancing the net Rossby transport with the Kelvin transport gives the amplitude of the Kelvin wave due to reflection of each meridional mode (Fig. 3).

Figure 3 shows that only the first, and to a lesser extent the third mode long Rossby waves produce a significant Kelvin wave reflection, for reasonable amplitude ULT anomalies. Clarke (1983) obtained a similar result, as did Cane and Gent (1984). The first and third mode waves have their turning latitudes at 5.4° and 8.2° latitude, respectively. The 20 m ULT mode 7 Rossby wave, which has its turning latitude at 12° latitude, where Graham and White (1988) suggested that the trigger for ENSO could be generated, produces a Kelvin wave ULT anomaly of about 1 m (Fig. 3), which is below the noise level of currently available observations. Since $g'/c \approx 0.018\ s^{-1}$ [see Eq. (1)], the anomalous zonal current on the equator of the mode 7 wave is about 2 cm s^{-1} (Fig. 3), not enough to accomplish any further feedback through zonal advection. Therefore, it is clear from linear equatorial wave theory that if realistic extra-equatorial Rossby waves have an effect on the evolution of ENSO in the waveguide, it cannot be through simple western boundary reflection.

The physical reason for the sharp fall-off in Kelvin amplitude with latitude of the initiating Rossby wave is that the net zonal transport associated with a given ULT anomaly is due entirely to the difference in $1/f$ between the northern and southern side of the anomaly, and this difference becomes smaller farther from the equator. Consider an extra-equatorial pycnocline "hump" (uplifted pycnocline), propagating westward as a free long Rossby wave in an otherwise resting ocean. The hump is associated with eastward geostrophic flow on the equatorward side and westward flow on the poleward side. On an f-plane, the total volume transport on either side is identical; on a β-plane, however, the eastward transport is larger by being closer to the equator, and the net zonal transport is not zero. When the hump arrives at the western boundary, an equatorward western boundary current is set up which connects the two zonal transports (e.g., Cane and Gent 1984); a deficit exists, however, because of the net eastward transport. Since (low-frequency) information can only propagate equatorward on a western boundary, the net zonal mass flux at the boundary can only be balanced by the westward geostrophic flow of an upwelling equatorial Kelvin wave. Similarly, a downwelling extra-equatorial Rossby wave has a net westward transport, which is balanced by a downwelling equatorial Kelvin wave. Regardless of the details of the reflection in the western boundary layer (which remain unknown without further assumptions about the boundary layer physics), the requirement that the zonal transport must be balanced is sufficient to describe the resulting Kelvin wave completely [Eq. (2)]. For a given height and form of ULT anomaly, the net Rossby transport decreases as the anomaly occurs farther from the equator, and therefore the reflected Kelvin wave amplitude also decreases. The net transport is proportional to the difference in $1/f$ between the equatorward and poleward sides of the ULT anomaly. If we take y_0 to be the center and Δy to be a scale meridional width of the ULT anomaly, this difference (with $f = \beta y$) equals $2\Delta y/\beta(y_0^2 - \Delta y^2)$, and decreases approximately as $1/y_0^2$.

Graham and White (1991) point out that the largest zonal transports of extra-equatorial Rossby waves are found 3°–4° equatorward of the largest ULT anomalies. While this is true, the poleward side of a ULT anomaly has oppositely directed transport, which nearly cancels that on the equatorward side; it is the net or difference between these two that is important. The appropriate way to characterize the wave is by the location of the center, namely the latitude of the maximum ULT anomaly.

Summarizing the results shown in Fig. 3, we may make a practical distinction between "near-equatorial" and "extra-equatorial" Rossby waves on the basis of which waves feed significant information back to the

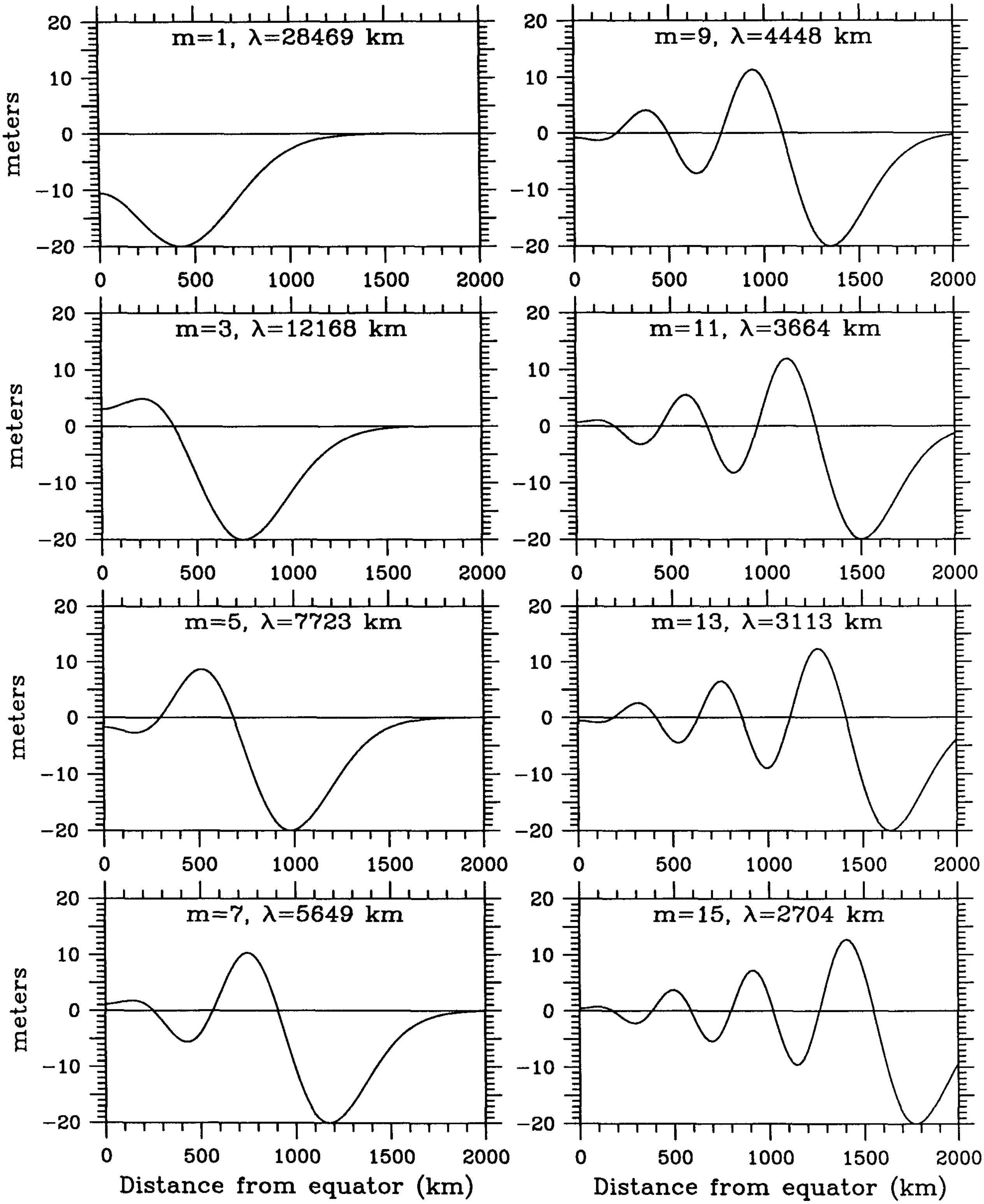

FIG. 1. Upper layer thickness (ULT) anomalies for odd (symmetric) long Rossby wave meridional modes (m) 1 through 15 as a function of distance from the equator. The period of oscillation is one year, and each mode has been scaled to a maximum anomaly of 20 m. The zonal wavelength (km) of each mode is given as λ.

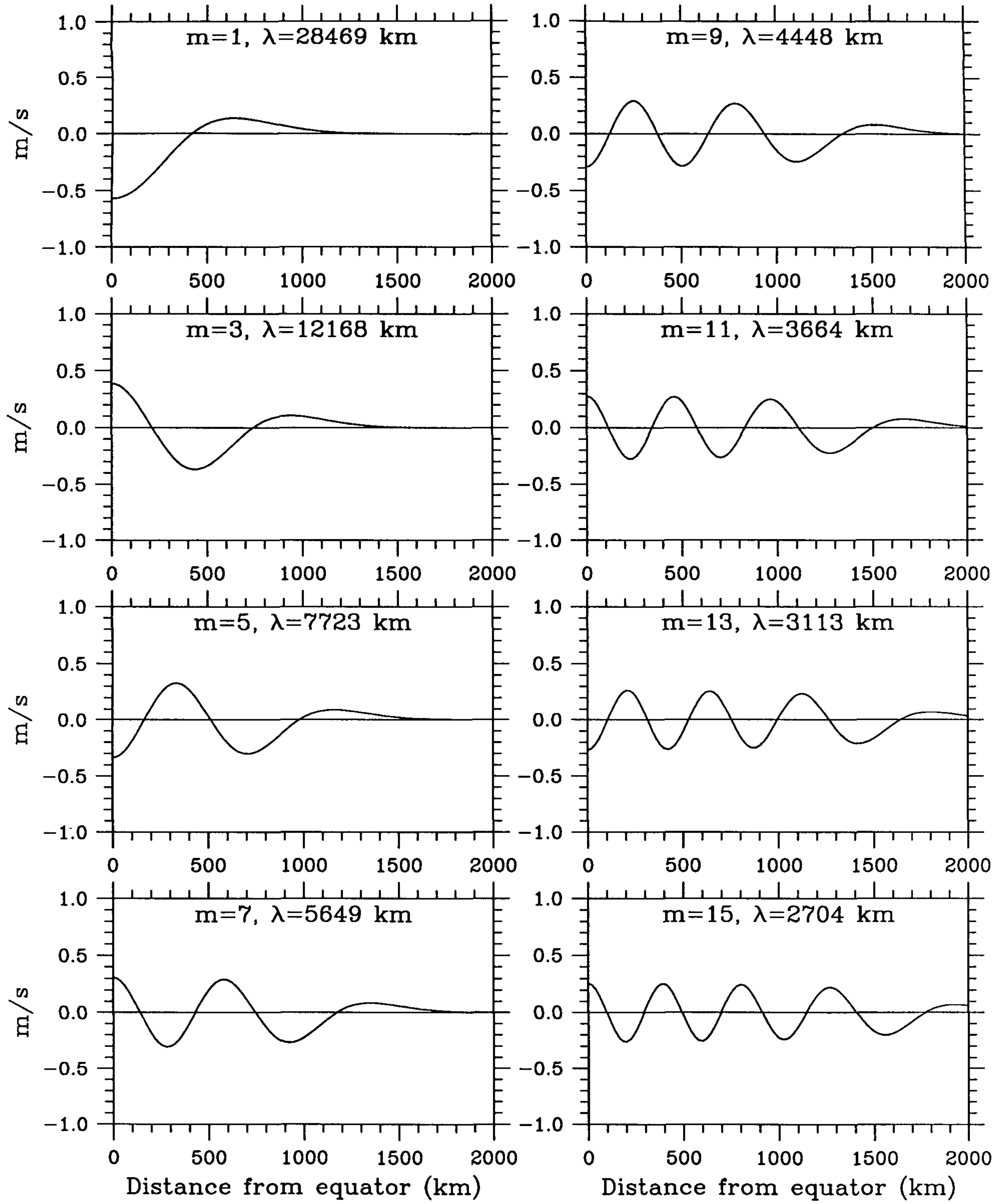

FIG. 2. As Fig. 1 but for zonal velocity (m s^{-1}).

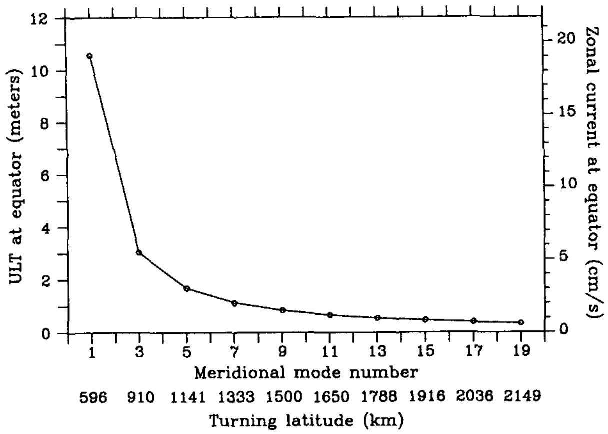

FIG. 3. The ULT (m) and zonal current (cm s^{-1}) anomalies at the equator for the Kelvin wave generated by reflection of each of the Rossby modes shown in Figs. 1 and 2, calculated by equating the net zonal transport of the Rossby modes with that of the Kelvin wave. The period of oscillation is one year and the maximum height of each Rossby mode ULT anomaly is 20 m. The scale for ULT is at left, and for zonal current at right; the two are related by the factor $g'/c \approx 0.018\ s^{-1}$ [see Eq. (1)].

equator upon reflection at the western boundary. For realistic amplitude thermocline anomalies, 8° is the approximate latitude dividing the "near-equatorial" region where all the waves can have subsequent equatorial effects from the "extra-equatorial" region where wave energy (which may have been generated by equatorial winds) is removed from and can no longer significantly affect the equator.

3. Kelvin wave amplitude for observed ULT anomalies in the tropical Pacific

Observations of ULT anomalies in the tropical Pacific have been compiled from historical data archives by Kessler (1989, 1990). The observations are nearly 200 000 XBT and MBT vertical temperature profiles collected from merchant, navy and fishing vessels during the 18-year period 1970 through 1987. The data processing eliminated fluctuations with periods less than about six months (observations are binned bimonthly); therefore these observations are suitable for examining low-frequency phenomena only (Kessler 1989). Evidence of long (nondispersive) Rossby waves was found in variations of the depth of the 20°C isotherm, which was used as a proxy for thermocline depth. Long Rossby waves were observed at both annual and interannual periods at latitudes between 4° and 18°N (Kessler 1989, 1990). These waves have also been observed by White (1977), Meyers (1979), White, et al. (1985a, 1985b, 1987, 1989), and Pazan et al. (1986).

Interannual thermocline variability in the tropical western Pacific was observed to have largest magnitude near about 12°N (Fig. 4). The standard deviation of interannual 20°C depth in the western tropical north Pacific was about 10–12 m during 1970–87 (Fig. 4). Pazan and White (1987, Fig. 5) show a slightly larger rms amplitude of observed variability in 14°C depth (about 15 m) for the period 1979–82, which included the massive anomalies associated with the 1982–83 El Niño. The FSU model driven by observed winds used by Graham and White produces off-equatorial interannual variability much larger than that observed. Kubota and O'Brien (1988, Fig. 3a) show interannual rms variability of upper-layer thickness in this model of up to 50 m at the Philippine coast at 15°N, for the period 1965–84, about 4 to 5 times larger than observed 20°C depth variability (Fig. 4). These very large signals in the model exaggerate the effects of extra-equatorial Rossby to Kelvin wave reflection. By contrast, the coupled model of Battisti (1989) has less ULT variability poleward of 10° latitude than do the observations. Much of the interannual variability seen in both the observations of isotherm depth and in the FSU model ULT was associated with the propagation into the western tropical Pacific of long Rossby waves generated in midbasin by the large wind stress curl anomalies during the ENSO events of 1972 and 1982–83 (White et al. 1987, 1989; Kessler 1989, 1990). For example, Fig. 5 shows 20°C depth anomalies at 13°N during the period spanning the El Niño of 1972. Kessler (1990) identified the shallow anomaly which propagated from the American coast in early 1971 to the western boundary by mid-1973 as a long Rossby wave generated principally by wind stress curl associated with equatorial westerly anomalies during the height of El Niño of 1972. The deep anomaly emanating from the eastern boundary in late 1972 was due to the reflection from the eastern boundary of a downwelling equatorial Kelvin wave. The deep anomaly which began in mid-1973 near 160°W and propagated to the western boundary was a long Rossby wave forced by downwelling wind stress curl associated with the return of strong equatorial easterlies in the aftermath of the 1972 El Niño. These signals have typical amplitudes of 10–30 m. Note, in comparison, that the FSU model ULT signals at 12°N have amplitudes in the western Pacific exceeding 100 m (Graham and White 1988, Fig. 7).

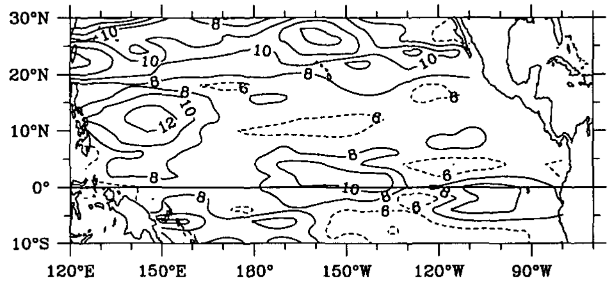

FIG. 4. Interannual standard deviation of 20°C depth (m) during 1970–87. Contours are every 2 m, with dashed contours indicating standard deviations of 6 m or less.

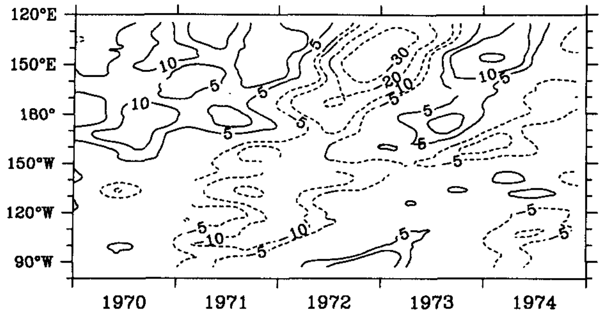

FIG. 5. Anomalies of 20°C depth (m) at 13°N during 1970 through 1974. Solid contours indicate deep and dashed contours shallow thermocline anomalies.

To calculate the Kelvin wave amplitude due to the observed long Rossby waves, the 1½-layer approximation was used to find the zonal geostrophic flows associated with observed pycnocline depth variations. As was done for the meridional modes in section 2, the intent is to integrate over the Rossby wave zonal transport, balance it with the zonal transport in the resulting Kelvin wave, and solve for the amplitude A of the Kelvin wave at the equator. There are serious difficulties with this calculation using observed thermal data, since it is impossible to extract the uncontaminated Rossby wave signal from the observations. There are two principal contaminants. First, there is thermocline depth variability due to equatorial Kelvin waves which must be excluded from the integration, and second, aliasing in the observations due to unresolved high-frequency noise overwhelms the geostrophic calculation close to the equator. Both of these problems can be partly avoided by starting the integration at 3°N, at the cost of poorly resolving the gravest mode waves which have significant amplitude equatorward of 3°N (Fig. 2). In addition, the resulting integral was found to be highly sensitive to the starting point of integration; if 2°N was chosen, the final value would be quite different because very small pressure gradients produce large transports close to the equator. Finally, contributions to the Kelvin amplitude from the Southern Hemisphere are omitted from the calculation due to very sparse data. Therefore, no claim is made that the value of the integral over y calculated in this way represents the actual contribution to equatorial Kelvin waves from observed Rossby waves. Presently available data are inadequate to answer this question definitively. Nevertheless, the calculation shows the latitudes where the integral changes value, in other words the locations where there is significant contribution to the amplitude of the Kelvin wave from observed geostrophic transport imbalances. To show the contribution by latitude, the result is given as a running integral over latitude (Fig. 6), as in Battisti (1989). The amplitude A of the Kelvin wave due to contributions between 3°N and latitude y is thus:

$$A(y) = -[\beta/2c^3\pi]^{1/2} \int_{3^\circ\mathrm{N}}^{y} \int_{H}^{0} u_{g140^\circ\mathrm{E}}\, dz dy'. \qquad (3)$$

The meridional integration was carried out in 1° latitude steps. All the anomalous zonal geostrophic transport across 140°E between 3° and 30°N was assumed to be associated with long Rossby waves; 140°E was chosen because it is near the boundary but far enough offshore to be well east of the western boundary layer (where the long Rossby signal is also contaminated by short Rossby waves in the western boundary current and where the length scales are much shorter and more difficult to sample accurately). Results from other meridians between about 135° and 160°E were similar, since the observed thermal structure variations were similar to 140°E. Figure 6 shows that most of the total Kelvin amplitude due to reflection was generated

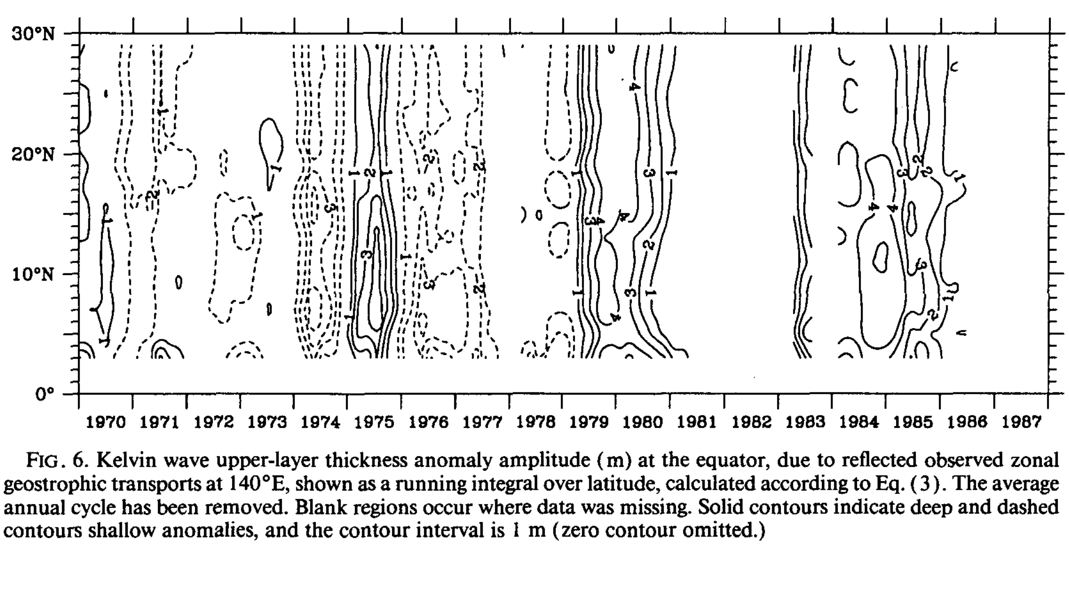

FIG. 6. Kelvin wave upper-layer thickness anomaly amplitude (m) at the equator, due to reflected observed zonal geostrophic transports at 140°E, shown as a running integral over latitude, calculated according to Eq. (3). The average annual cycle has been removed. Blank regions occur where data was missing. Solid contours indicate deep and dashed contours shallow anomalies, and the contour interval is 1 m (zero contour omitted.)

within about 8° of the equator and represents Kelvin wave amplitudes of about 4 m during several one- to two-year-long events (unfortunately the important period from mid-1981 through early 1983 is missing from the record). Poleward of 8°N the value of the integral (3) is nearly constant, indicating little contribution to Kelvin waves from the extra-equatorial region. Small increases in amplitude are seen to have been generated near 10°–12°N at the end of 1972 [the reflection of the upwelling Rossby wave forced during the peak of the 1972 ENSO (Fig. 4)] and in 1984 (the reflection of the downwelling wave associated with the return of equatorial trade winds following the 1982–83 ENSO). In accord with the simple theory discussed in part 2, both of these signals (which are the largest extra-equatorial variations in the 18-year record) contributed only about 1 m in ULT change, or about 2 cm s^{-1} zonal current, to the equatorial Kelvin wave. In sum, the observations show that although strong thermocline variability extends well outside the equatorial region, these extra-equatorial signals do not produce a significant net transport imbalance that would lead to a large waveguide signal. Unfortunately, the data do not permit direct observation of equatorial Kelvin waves, because the low sampling density dictates a minimum bimonthly binning. Since a first baroclinic mode Kelvin wave would take only two months to cross the entire basin, Kelvin waves cannot be unambiguously identified in this dataset.

4. Conclusion

The issue of whether the extra-equatorial regions of the Pacific feed information back into the waveguide through reflection of Rossby waves from the western boundary has been a subject of recent debate. Large-amplitude extra-equatorial thermocline anomalies in the western tropical Pacific associated with ENSO have been observed and identified as long Rossby waves by several investigators. However, well-known linear equatorial wave dynamics demonstrates that efficient reflection of these waves can occur only within about 8° of the equator, with a sharp poleward fall-off in the reflectivity. In addition, the observed zonal geostrophic transports do not support the hypothesis of significant contributions to equatorial signals from the northern tropical Pacific. These results are consistent with recent coupled model studies (e.g. Battisti 1989) which show that ENSO-like oscillations can be generated with no involvement of the extra-equatorial regions. Therefore, if reflection does play a major role, it must be through processes not included in the simple dynamics outlined above. Reflection of extra-equatorial waves does appear to be important in the linear, wind-driven model studied by Graham and White (1988). However, the model produces extra-equatorial Rossby waves much larger (by a factor of four or five) than those observed, which exaggerates the role of reflection. Kessler (1990) found evidence that some ULT signals appear to make a continuous circuit around the northern tropical Pacific, as if reflection were occurring. Graham and White (1988) also show evidence of continuity of off-equatorial signals with subsequent waveguide anomalies; in both cases the continuity was associated with the large anomalies due to ENSO. However, given the basinwide wind anomalies which occur during ENSO events, it is not surprising or necessarily illuminating to find correlations (and lag correlations) of variables all over the Pacific with each other. The apparent continuity seen in observations may well be due to coherence of ENSO forcing over very large regions, and the regularity of ENSO during the past several decades. The off-equatorial signals may be simply coincident with much more important near-equatorial waves. On the other hand, the interaction of SST and tropical cyclone formation is still poorly understood. It may be that coupled interactions occur in the western Pacific warm pool whereby very small oceanic signals are amplified through processes other than linear wave propagation. Therefore, the possibility must remain open that the large extra-equatorial signals do exert some influence on the subsequent evolution on the equator, although it cannot be through as simple a mechanism as proposed by Graham and White (1988).

Acknowledgments. This research was supported by the National Science Foundation under grants OCE-88-16910 and OCE-90-12508. I thank David Battisti, Nick Graham and Warren White for perceptive comments and discussion which helped clarify this note.

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