Observations of Long Rossby Waves in the Northern Tropical Pacific

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1. Introduction

Long baroclinic Rossby waves are potentially important in the adjustment of the tropical (extra-equatorial) Pacific pycnocline to both annual and interannual wind stress curl fluctuations. The purpose of this paper is to document the occurrence of these waves from historical data and to show that much of the observed low-frequency pycnocline variability in the northern tropical Pacific may be represented by a simple model expressing the physical processes of Ekman pumping, the radiation of long (non-dispersive) Rossby waves due to such pumping in mid-basin, and the radiation of long Rossby waves from observed eastern boundary thermocline depth fluctuations. Evidence of long Rossby waves is found in bathythermograph (BT) observations of the depth of the 20°C isotherm, which is used as a proxy for pycnocline variability. The progenitor of the present study is Meyers' (1979a) investigation of the annual cycle of 14°C depth, which used a similar model of a low-frequency linear quasi-geostrophic two-layer ocean. Much of Meyers' (1979a) work was updated (and largely confirmed) in Kessler (1989a, b), with the addition of ten more years of both BT and wind observations.

In a series of papers, White and collaborators (White, 1977, 1983; White, et al., 1982, 1985; Inoue, et al., 1987; Pazan, et al., 1986; Pazan and White, 1987; Graham and White, 1988) have sought to interpret BT observations in the extra-equatorial north Pacific in terms of similar simple long Rossby wave dynamics, often using the linear reduced-gravity model developed by Busalacchi and O'Brien (1980). Some of their results are consistent with conclusions reported here, however the expendable BT (XBT) data available to these investigators are distributed primarily along a few shipping lines, with large gaps in some important areas, requiring relatively elaborate interpolation schemes between the ship tracks or the use of model output to fill data gaps. Because of this, their interpretation of observed interannual variability as evidence of zonal propagation has been controversial. An essential element of the present study is in the addition of a large set of mechanical BT (MBT) data taken by Japanese fishing vessels, which is concentrated in a previously data-poor region in the western tropical Pacific and makes possible the construction of long time series of thermocline depths in the western Pacific with a minimum of zonal interpolation.

Section 2 discusses the BT data sets, their processing and gridding. The simple model of low-frequency pycnocline variability is developed in Section 3, and the approximations made are discussed. Section 4 focuses on the role of long extra-equatorial Rossby waves in the evolution of El Niño, and the variability is discussed in terms of the simple dynamics outlined above, comparing the observations with hindcasts derived assuming either local Ekman pumping or including the radiation of low-frequency Rossby waves. Waves due to two principal forcing sources are identified: upwelling long Rossby waves generated in mid-basin by the large scale wind stress curl pattern associated with strong equatorial westerly wind anomalies; and downwelling long Rossby waves which radiate from the eastern boundary following the reflection of the deep equatorial thermocline anomaly which occurs during El Niños. The suggestion that long Rossby waves in the extra-equatorial tropics can be a trigger for subsequent El Niños, which has been debated in the recent literature, is evaluated from several time series which extend back to the 1950's, spanning seven major El Niño events.

A brief note on terminology is necessary: the word "tropical" is used in the sense of "extra-equatorial", to refer to the regions between about 5° and about 25° latitude. As discussed in Section 2, the observations are inadequate to study the rapid zonal propagation which occurs in the equatorial waveguide, so dynamical interpretation is confined to the extra-equatorial, "tropical" regions.

2. Data collection and processing.

a. Introduction

Several organizations hold large archives of historical BT observations, each of which densely samples particular regions; however none of these data sets individually has sufficient coverage to document zonally...
Figure 1a. Distribution of XBT observations during 1970 through 1987.

Figure 1b. Distribution of MBT observations during 1970 through 1980.
propagating events over the width of the tropical north Pacific. In particular, the XBT data sets which have formed the basis for many studies of tropical Pacific variability (e.g. Meyers and Donguy, 1980, 1984; Kessler, et al, 1985; Rebert, et al, 1985; White, et al. 1985; Pazan and White, 1987; Kessler and Taft, 1987; Graham and White, 1983; McPhaden, et al., 1988a,b) are derived principally from merchant and navy vessels, which densely sample a few specific shipping lanes, but poorly observe a large region of the western north Pacific southwest of Hawaii (Figure 1a). It has proven difficult to draw convincing evidence of zonal propagation from these data. On the other hand, many MBT profiles were made (mostly by Japanese fishing vessels) in the western tropical Pacific during the 1970's, which nicely complement the XBT data distribution (Figure 1b). Even the combined data are adequate to construct long time series only in the region between 5°N and 30°N, where the observations are densest and zonal propagation of long waves is relatively slow. Although it would be desirable to develop a historical time series of thermocline fluctuations in the equatorial waveguide, the data are insufficient to do so, particularly since time scales near the equator are so short. In most of the southern hemisphere east of 150°W the observations are so sparse that even the mean cannot be found with confidence. In this section the various data sets are described and the methodology and criteria for combining and gridding them are set forth. Although the BT sampling is highly irregular and the data are aliased by high-frequency phenomena, so gridding, interpolation and smoothing are necessary, every effort has been made to keep the processing of data to a minimum, in order to let the data "speak for itself". In particular, care was taken that interpolation in longitude was minimized to avoid the possibility that such interpolation would lead to a misleading interpretation as zonal propagation. Secondly, since the tropical Pacific is characterized by sharp zonally oriented ridge and trough features, interpolation in latitude might tend to smooth these structures; thus no interpolation in latitude was used.

b. Description of bathythermograph data sets

The largest single temperature profile data collection is in the archives of the National Oceanographic Data Center (NODC), consisting of 108632 XBT observations in the Pacific Ocean part of the region 30°S-30°N, 110°E-70°W, during 1967 through 1985. This data set provides much of the basis for the Levitus (1982) climatology. The NODC data comprise most of the XBT observations shown in Figure 1a, with the exception of the sharply defined route from New Caledonia (21°S, 166°E) to Japan and the routes emanating from Tahiti (17°S, 150°W), which are due to the French-U.S. ship of opportunity (SOP) program discussed below. About half the observations shown on the mid-basin routes from the southwest Pacific to Hawaii and North America were also made by SOP ships. The NODC data are densest on ship tracks traveled by United States Navy and merchant vessels, particularly between the U.S. mainland and Hawaii, Guam (13°N, 145°E), the Philippines, Japan and Australia (Figure 1a). However, south of the Hawaii-Guam-Philippines ship track and west of the Dateline the NODC data set has only sparse and sporadic sampling (Figure 1a). It is clearly an insufficient basis for study of time-dependent phenomena in the western tropical and equatorial Pacific.

The French-American SOP program began collecting surface observations (temperature, salinity and biological variables) in 1972 (Donguy and Dessier, 1983) and XBT profiles in 1979 (Meyers and Donguy, 1980) from volunteer merchant ships operating out of Noumea, New Caledonia. The ships traverse primarily three routes as mentioned in the previous paragraph (Figure 1a). A total of 26,373 XBT profiles were collected and made available under the SOP program during 1979 through 1987, only about one-quarter as many as in the NODC set, however the SOP data is much more concentrated in the tropics, with relatively dense sampling along the three tracks.

The combined XBT data shown in Figure 1a amounts to 135,005 total observations during 1967 through 1987. Almost all of the XBT profiles extend to 450 m depth, which is well below the thermocline in most of the tropical Pacific. Although this combined data set has been used to study zonal propagation of thermocline depth anomalies (e.g. White, et al, 1985), it is clear from Figure 1a that a very large region southwest of Hawaii was sampled principally on a single meridional section (crossing the equator near 160°E), which did not begin until 1979, so the XBT data alone is a poor antenna for zonally propagating variability in that important region, requiring relatively elaborate statistical interpolation schemes between the tracks (Pazan and White, 1987). Because of this interpolation, interpretation of observed variability as evidence of zonal propagation has been controversial. The present work avoids this problem by merging the XBT data with MBT data collected by Japanese fishing vessels, which fortuitously fills the large gap southwest of Hawaii (Figure 1b) and makes possible the construction of time series of thermocline depth across most of the northern tropical Pacific with a minimum of zonal interpolation.
The MBT data set studied here was compiled by the Japanese Far Seas Fisheries Research Laboratory (FSFRL, Shimizu, Japan, and was generously made available through a cooperative agreement with PMEL (K. Mizuno, personal communication). There is also a large collection of MBT observations taken before 1970 in the NODC archives, which were included in the Levitus (1982) climatology. These data are for the most part not used in the present study, since the data coverage in the tropical Pacific is not dense enough to grid on the relatively fine scales possible with the combined data after 1970. However, in a few regions (for example along the ship route from the U.S. West Coast to the Panama Canal) the NODC MBT data has been used to construct several long time series which are discussed in Section 4.

A total of 64,062 FSFRL MBT profiles were received, covering the period 1970 through 1980. The MBT observations extend only to 250 m depth, which is below the thermocline near the equator, but within the lower thermocline in the subtropical gyres. The 20°C isotherm is practically the deepest isotherm sampled over the entire tropical Pacific by MBTs. The MBT profiles studied here are taken principally in the course of fishing operations, in which many ships work for several months in a small region, and so have a different type of spatial distribution than the XBT data, which were generally sampled along underway ship routes (Figure 1b). Additionally, there is some seasonal variability in the MBT data distribution (Kessler, 1989a).

The interannual time distribution of the sampling was roughly constant during 1970 through 1980, when data from both the NODC and the FSFRL are available, with the highest sampling density during 1970 through 1975. The SOP data begins in 1979 while the FSFRL set ends in 1980, so the distribution in the large region between Hawaii and the Indonesian Archipelago changed at that time from the fishery MBT pattern (Figure 1b) to mostly meridional XBT tracks (Figure 1a). During the El Niño years of 1982-83 the observations available in this region were essentially just from the SOP program and have large gaps between the tracks. The total data set is thus most useful for studying zonally propagating events during the 1970's.

Kessler (1989a) compared the mean depth of several isotherms from the MBT and XBT data sets in regions where there was substantial overlap. He found an apparently systematic deep bias of MBT isotherm depths compared to the XBT, averaging 2 to 5 m for 20°C. However, in this paper it is principally anomalies of isotherm depth which are of interest, so this small instrument bias can be avoided by finding anomalies from the mean of the corresponding data set, rather than from the mean of the total. With this caveat, it is permissible to merge the MBT and XBT data sets, giving a total of 199,067 profiles of both types. After quality control, the cleaned data were gridded with a gridsize of 2° latitude, 5° longitude and 2 months. See Kessler (1989a, b) for a discussion of quality control and data gridding.

c. Wind data

In Section 2, a simple wind-forced model is developed to help explain observed pycnocline depth fluctuations. The winds which are used to force this model are produced by the Mesoscale Air-Sea Interaction Group at Florida State University under the direction of J. J. O'Brien and D. Legler, from ship winds subjectively analyzed onto a 2° latitude by 2° longitude grid (Legler, personal communication), and generously made available by them. This ship wind processing effort was begun at the University of Hawaii by G. Meyers and K. Wyrtki (Wyrtki and Meyers, 1975a, 1975b, 1976) and continued at Florida State University (Goldenberg and O'Brien, 1981). This wind data set has been used to force a variety of models of the tropical Pacific (e.g. Busalacchi and O'Brien, 1980; Busalacchi, et al., 1983; McPhaden, et al., 1988a,b; Harrison, et al., 1989). The pseudo-stress values provided were converted to stress using a value of $1.5 \times 10^{-3}$ for the drag coefficient. Since the meridional scale of the wind stress is typically much smaller than the zonal scale, the wind stress curl was found by differencing $\tau^c$ over 2° latitude and $\tau^s$ over 4° longitude.

3. A simple model of low-frequency pycnocline depth variations.

a. Introduction

The field of 20°C isotherm depth was chosen as the clearest, simplest representation of the observed variability. Implicit in this choice is the idea that the tropical Pacific may be approximated as a two-layer system divided by a sharp pycnocline, for which the 20°C isotherm is a proxy. Meyers (1979a,b) used the depth of the 14°C isotherm in this way to study the annual cycle. Kessler and Taft (1987) showed that volume transports of the North Equatorial Current and Countercurrent derived from the two-layer assumption, using the vertical motion of the 20°C isotherm to represent interfacial movement, give a reasonably good estimate of zonal geostrophic volume transport variability. In the same spirit, many modelling studies have used a reduced
gravity, single-active-layer approximation (e.g. Busalacchi and O’Brien, 1980, and successors; see also McCreary, 1985). The principal advantage of this choice is that it simplifies the three-dimensional variability into a two-dimensional field which can be mapped and studied conveniently. For practical reasons, since the BT profiles extend to a variety of depths, a calculation needing a common reference level (such as dynamic height) poses problems and requires further assumptions necessary to extrapolate all profiles to the same depth (but see Kessler, et al., 1985) for a discussion of a scheme for dealing with this). As discussed in Section 2.b, the Japanese MBT data extends only to 250 m, so 20°C is effectively the deepest isotherm fully sampled over the entire tropical Pacific. The observations cannot be reasonably gridded with a time resolution of less than two months (and must be smoothed in time (see Section 2.c)), so the BT data set is suitable only for the study of low-frequency variability. Thus, the simple model was developed assuming that all variability can be described by low-frequency, large-scale (long wave limit) fluctuations of a sharp pycnocline.

b. The simple model

A simple quasi-geostrophic model of low-frequency pycnocline variability was employed by Meyers (1979a) to study annual fluctuations of the 14°C isotherm in the tropical (not equatorial) Pacific. The physical processes allowed by the model are Ekman pumping, the radiation of long (non-dispersive) Rossby waves due to such pumping in mid-basin, and the radiation of long Rossby waves from observed eastern boundary depth fluctuations. The ocean is assumed to consist of a rigid lid over a single active layer, separated by a sharp pycnocline from a deep motionless lower layer; both layers are homogeneous (the reduced gravity, \( g' = g \Delta \rho / \rho \) describes the potential difference between them). The wind stress \( \tau \) (see Section 1.c) is assumed to act as a body force in the upper layer. Since monthly average wind data is used, the forcing frequency is much slower than inertial (\( \partial / \partial t \ll f \)). The long wave approximation is made (see Kessler, 1989b, for discussion of this simplification). The momentum equations are linearized by assuming that the mean upper layer depth is spatially uniform. Then the low-frequency forced quasi-geostrophic vorticity equation as was used by Meyers (1979a) is:

\[
\frac{\partial h}{\partial t} - \frac{\partial c^2 f^2}{\partial x} = \nabla \cdot (\tau + f \rho f), \tag{3.1} \]

where \( c^2 = g' H \) is the long baroclinic gravity wave speed squared. Balance (3.1) is the simplest relation obtainable describing time-dependent \( \beta \)-plane dynamics. The long wave, low frequency limit has been assumed by neglecting the local accelerations; this approximation makes wave solutions to (3.1) non-dispersive, so Rossby waves in the simple model propagate exactly westward. Thus, in this limit solutions at each latitude are independent. Furthermore, there is no need to assume that the forcing or solution have a sinusoidal form. There are no short Rossby waves in the model, so it is only valid where the zonal length scale of solutions is much greater than the local gravest internal Rossby radius. In addition, neglecting \( u_0 \) excludes the equatorial Kelvin wave and is not valid within about 5° of the equator, so (3.1) is a mid-latitude approximation. Garzoli and Katz (1983) estimated the magnitude of the contribution of the acceleration terms which have been neglected to the balance (3.1), using BT data in the tropical Atlantic. They concluded that these terms did not contribute significantly to the thermocline displacement anywhere in the region which they studied (3° to 10°N).

Solutions \( h(x,t) \) to (3.1) are referred to as the "upper layer thickness anomalies" (ULT) (measured positive down, or thicker), continuing terminology used by Busalacchi and O’Brien (1980). These solutions were found according to the procedure given by Gill and Clarke (1974, section 7), integrating along characteristics (which are just latitude circles in the long wave limit) from the eastern boundary, where \( h(x=0,t) \) was taken to be the observed annual cycle of 20°C depth. Meyers (1979a) gives an equivalent procedure. If the wind stress \( \tau \) is given, then (3.1) can be straightforwardly solved if a value for the long Rossby wave speed \( c_r = \beta c^2 f^2 \) is known. As did Meyers (1979a), \( c_r \) was treated as an unknown and studied by integrating (3.1) at each latitude, trying all values of \( c_r \) from 1 cm/s to 1 m/s at 1 cm/s intervals, then correlating each solution \( h(x,t) \) with the observed 20°C depth variations, either at the annual period (Kessler, 1989a,b) or for interannual variability (Section 4), to find a best-fit value for \( c_r \) at each latitude. Since only the long gravity wave speed \( c \) is truly unknown, effectively the fitting process found an estimate of \( c \) which maximized the correlation between model and data over the width of the basin at each latitude.

Several limitations of the simple model (3.1) are evident: the assumption that all waves can be described by the long wave limit, the linearizing assumption in the derivation of the model that the mean pycnocline is flat and the strong zonal geostrophic currents of the tropical Pacific are absent, and the neglect of...
Figure 2. Best-fit estimates of the propagation speed of long Rossby waves. The two smooth curves represent the theoretical speeds \( c_r = \beta c^2/f^2 \); the thin solid line uses \( c \) found from a first baroclinic mode decomposition given in Emery, et al. (1984); the dotted line uses \( c = \sqrt{gH} \), where \( H \) is the depth of 20°C at 160°W. The two heavy lines represent best-fit estimates of \( c_r \). The dashed line (discussed in Section 4.e) shows estimates for the annual cycle (no estimates were made at 10° or 12°N for annual cycle). The solid line shows best-fit estimates for interannual variability (discussed in Section 4.d). The "error bars" refer to these interannual estimates; the method used to calculate these is discussed in the text.

Figure 3. Interannual anomalies of 20°C depth (m) along the American coast. Contours are given at ±2, 5, 10, 15 and 20 m, with positive (deep) anomalies solid, negative (shallow) anomalies dashed.
the more complicated observed vertical structure in favor of simple single-active-layer dynamics. Kessler (1989a, b) discusses these limitations, concluding that for waves with periods greater than about one year in an ocean with a sharp pycnocline the long wave limit is appropriate, and that the zonal geostrophic currents do not modify wave propagation, due to the non-Doppler effect (see also Chang and Philander, 1989).

The observations are also compared with hindcasts derived assuming local Ekman pumping alone, that is, solutions \( h_E(t) \) to:

\[
\frac{\partial h_E}{\partial t} = -\text{Curl}(\tau/\rho f),
\]

with the boundary condition that the mean \( h_E \) must be zero.

4. Interannual variability of 20°C depth

a. Introduction

Interannual variability was studied according to the same simple model as was annual variability, using the same vorticity equation (3.1). The reader is referred to Section 3 for discussion of the model and its limitations. The primary issue addressed in this section is the role of long extra-equatorial Rossby waves in the evolution of El Niño. In general, two types of long Rossby waves are thought likely to be produced in the tropics during El Niños: First, upwelling long Rossby waves may be generated in mid-basin by the large-scale wind stress curl pattern associated with equatorial westerly wind anomalies. Secondly, downwelling waves may occur following an El Niño, initiated by reflection off the eastern boundary of deep equatorial pycnocline anomalies. Although both waves occur in the tropical, extra-equatorial region, both originate from events in the equatorial waveguide. Such wave-induced variability has been noted in many model experiments. McCreary (1983) and McCreary and Anderson (1984) developed a simple model of coupled ocean-atmosphere interaction in which long Rossby waves were generated in the central tropical Pacific by idealized wind stress curl anomalies associated with changes of SST in the eastern equatorial Pacific. In this model, the travel time of these extra-equatorial waves set the oscillation period of El Niño events, and they played a crucial role in both triggering and turning off El Niños through reflection off the western boundary. Busalacchi, et al. (1983) used a linear reduced-gravity model forced by observed winds and showed that in this model shoaling of the western Pacific pycnocline observed in the latter stages of El Niño was due to the arrival of long Rossby waves. These waves were generated by the off-equatorial wind stress curl pattern associated with strong westerly wind anomalies near the equator at the height of the event. The physics of the Busalacchi, et al. (1983) model are similar to those expressed in the vorticity equation (3.1) for extra-equatorial regions, and one of the principal results of this section is that wave-induced fluctuations in the western Pacific like those seen in the model of Busalacchi, et al. (1983) can be observed in 20°C depth. More recently, Graham and White (1988) suggested that a mechanism similar in some important respects to that proposed by McCreary (1983) could be verified as triggering several observed El Niños. Results presented in this chapter suggest problems with the Graham and White (1988) scenario.

b. Interannual variability on the eastern boundary

As mentioned in Section 2, the NODC archives contain MBT data taken before 1970 which in a few regions of the Pacific is sufficient to construct long time series. One of these regions is along the shipping route between the U.S. West Coast and the Panama Canal, which hugs the coast of Mexico and Central America. A long time series of anomalous 20°C depth on the eastern boundary from 1954 through 1987 can be made by combining the early NODC MBT data with the later XBT data set. This time series shows El Niño events clearly as deep thermocline anomalies, which occurred in 1957, 1963, 1965, 1969, 1972, 1976, weakly in 1979, and in 1983 (Figure 3). The large El Niños (in 1957, 1965, 1972, 1976 and 1983) had amplitudes of 10-15 m at the equator on the eastern boundary, and the anomalies generally extend north to about 22°-25°N (which, possibly coincidentally, is where the relatively smooth coast is broken by the Gulf of California). Thermocline deepening in these events typically lasted for 12-18 months (Figure 3). Kessler (1989a) discusses in detail how an idealized ocean would respond to such pulse-like forcing along the eastern boundary by radiating Rossby waves offshore; however, as will be shown later in this chapter, the signal of such offshore radiation was observed only far from the equator and did not appear to be a major contributor to the subsequent evolution of El Niños.
Figure 4. Anomalies from the average annual cycle of zonal wind stress ($10^{-2} \text{ N m}^{-2}$) on the equator. Contours are given every $10^{-2} \text{ N m}^{-2}$, with positive (eastward) anomalies dashed and negative (westward) anomalies solid lines.

Figure 6. Contours of the curl of the wind stress ($10^{-8} \text{ N m}^{-2}$) along 13°N (longitude scale on left axis) during 1971 through 1973. Dashed contours are positive curl (Ekman upwelling); solid contours are negative curl (Ekman downwelling). The heavy overlaid line is the time history of 20°C depth (m) on the equator at the eastern boundary, which is used as an index of the peak of El Niño of 1972. The scale for 20°C depths is on the right axis, with negative (shallow) 20°C depths shown upwards and positive (deep) anomalies shown downward.

Figure 5. Third empirical orthogonal function of zonal wind stress (1961-87), representing 12% of the total interannual variance in the northern tropical Pacific. The top panel is the eigenvector, in units of $10^{-2} \text{ N m}^{-2}$, while the bottom panel is the (dimensionless) time amplitude function. A positive product between the eigenvector and time amplitude function represents an eastward stress anomaly.
c. El Niño-related variability of the wind field

Although this paper does not focus on the wind field, it is appropriate to briefly discuss the type of wind fluctuations associated with El Niño which are important in generating extra-equatorial Rossby waves. The most prominent of these wind events is the occurrence of westerly anomalies in the equatorial waveguide at the height of El Niños, which leads to upwelling wind stress curl in the tropical regions. The time history of zonal wind stress on the equator (Figure 4) shows that during several strong El Niños (in 1965, 1972, 1982 and 1986, but not in 1976) major westerly wind anomalies (amplitudes typically $3 \times 10^{-2}$ N m$^{-2}$ or roughly 4-5 m/s) occurred near the Dateline. During the following year, easterly wind anomalies occurred in the same region with similar amplitude. The wind stress curl pattern associated with these equatorial anomalies is exemplified by the third Empirical Orthogonal Function (EOF) of the zonal wind stress $\tau_{xx}$ in the northern hemisphere, representing 12% of the total interannual variance in this region (Figure 5). That this third EOF is characteristic of El Niño variability is shown by the high correlation ($r=0.75$) of the time amplitude function (bottom frame) with the time history of 20°C depth anomalies at the equator on the eastern boundary (the equatorial time series from Figure 3, discussed in 4.b above), which is a good index of El Niño events. The other EOF time amplitude functions had much lower correlations with this El Niño index ($r<0.4$) and are not shown. The third EOF (Figure 5) shows that the El Niño time history was associated with the westerly anomalies on the equator near the Dateline seen in Figure 4, and that these anomalies were confined south of about 5°-6°N, with easterly anomalies centered near 15°N. This pattern of wind stress anomalies implies that strong upwelling curl characteristically occurs in the tropical region between 5° and 15°N near the Dateline at the height of El Niños, and is often followed by a return of anomalously strong equatorial easterlies and hence by downwelling curl in the same region one to two years later.

This sequence of events is seen in a detail of the wind stress curl field at 13°N and eastern equatorial thermocline depth during the El Niño of 1972 (Figure 6). (13°N is approximately typical of latitudes between 5° and 15°N during this period, and is plotted to correspond to BT data at 13°N discussed below). 20°C depth at the South American coast peaked downward in May-June 1972, marking the height of the event. The wind stress curl field at 13°N indicates strong upwelling between about 170°W and 150°E slightly earlier than this, in March through July 1972. The following year, both fields had reversed sign, and the anomalies were smaller. A corresponding plot of the curl at 13°N during the El Niño of 1982-83 shows a similar phasing of curl variations relative to the eastern boundary peak, with the upwelling curl maximum occurring near the Dateline while the eastern boundary thermocline was falling, and a downwelling curl event about a year later. The effect of these curl fluctuations was to force an upwelling Rossby wave in the tropics near the Dateline during the height of El Niño, then a downwelling wave a year or so later.

d. Observed interannual 20°C depths and hindcasts from the simple model

The long Rossby speed $c = \beta c^3 / f^2$ was treated as an unknown in the simple vorticity equation (3.1), and studied by solving (3.1) at each latitude using the interannual FSU winds, trying all values of $c$ from 1 cm/s to 1 m/s, then correlating each solution $h(x,t)$ with the observed interannual variations of 20°C depth. Significant peaks of correlation were found at all latitudes from 8°N through 22°N, with a summary of the results in Figure 2. An ad hoc estimate of the uncertainty of fitting $c$, was made by finding the range of $c$, at each latitude such that the correlation of the solution $h(x,t)$ with observed 20°C depth was within 10% of the best correlation; these uncertainties are shown as the "error bars" about the best-fit meridional profile in Figure 2. The uncertainties ranged from about ±3 cm/s at 22°N to about ±20-30 cm/s at 8°N, and were typically about ±10 cm/s in the region 10° to 18°N (Figure 2). The interannual best-fit long Rossby speeds were generally smaller than the annual speeds in the NEC region, but within the estimated uncertainty, except at 22°N. Despite the fact that evidence of an annual Rossby wave was found at 4° through 6°N (Kessler, 1989a,b), interannual propagation was not clearly detected by the present method at these latitudes.

Correlation of solutions to the long Rossby wave model (using the best-fit speeds $c$, shown in Figure 2) with observed interannual 20°C depths was high in the west, with values above 0.7 west of the Dateline between 8° and 16°N, and also high east of the Dateline at 16° to 22°N (Figure 7, top). If we take the interannual autocorrelation timescale to be roughly four years (as estimated by Kessler (1989a)), then with bimonthly data there are about 24 degrees of freedom, and a correlation of 0.4 is significant at the 95% level. Since the simple model took the observed 20°C depths at the American coast as a boundary condition, the correlations were exactly 1 at the coast; the fact that correlations fell off rapidly offshore south of about 14°N suggests that
Figure 7. Top: correlation of observed interannual 20°C depth anomalies with the ULT hindcast made using the vorticity equation (3.1). Bottom: correlation of observed interannual 20°C depth anomalies with the hindcast made assuming Ekman pumping acting alone. In both, positive correlations are shown as solid contours, negative correlations as dashed contours. Dot shading is used to indicate correlations greater than 0.7, and hatching is used to indicate correlations greater than 0.5.
offshore radiation of boundary anomalies (which dominated the model variability in the region east of about 150°W) was probably not a major contributor to the variability in mid-basin equatorward of 14°N. Near 140°-150°W equatorward of about 12°N is a region of near-zero correlations. There is no obvious explanation for the poor model results in this area, which could be due to errors in the wind, aliasing or errors in the BT data (this is a relatively data-sparse area), or more complicated physical mechanisms at work than are expressed in the simple model (3.1) (an obvious possibility is the 20-30 day instability waves). In most other areas of the northern tropical Pacific, however, the simple model hindcasts were considerably more closely correlated with observed conditions than hindcasts made assumed Ekman pumping alone (equation (3.2); Figure 7, bottom). Over much of the central basin, the Ekman pumping hindcast was essentially uncorrelated with the data; only in small regions in the far west and near Hawaii does the Ekman pumping hindcast appear significantly correlated with the observations, and even in these regions the correlation was weaker than the simple model including long Rossby waves. In the region near 140°-150°W where the simple model hindcast was poorly correlated with the observations, the Ekman pumping hindcast was also uncorrelated. The poor interannual correlation with Ekman pumping contrasts with the results found in (Kessler, 1989a,b) for the annual cycle, in which at 10°-12°N Ekman pumping was found to be the dominant process at the annual period. Interannual solutions of the vorticity equation (3.1) (referred to as the upper layer thickness anomaly, or ULT), are compared with observations and with pycnocline anomalies hindcast by Ekman pumping alone (the solution to (3.2)) in time-longitude plots at 13°N, Figures 8 and 9, for the periods 1970 through 1975 and 1980 through 1986. The comparison is made at 13°N because first, there is a greater density of data at that latitude than at any other, particularly in the west; second, because 13°N is sufficiently far from the equatorial waveguide that the variability there is clearly governed by the quasi-geostrophic dynamics expressed by (3.1); and third, because this latitude has been suggested as a likely region for long Rossby wave propagation which might be important in the evolution of El Niño (Graham and White, 1988). The observed variability of both the wind stress curl and the depth of 20°C at 13°N are relatively typical of latitudes from about 10° to 15°N (Kessler, 1989a). The periods 1970-75 and 1980-86 were chosen because they span the strong El Niños of 1972 and 1982-83. As noted in Section 1, the 1970-75 period was relatively well sampled due to the availability of data from both the Japanese fisheries and the NODC, with an average of one observation every 13 days in each 5° longitude box at 13°N during that six-year period. The 1980-86 period, which was observed only by XBTs, was much more sparsely sampled, with an average of one observation every 38 days in each box; gaps appear as "holes" in Figure 9. Note in particular the absence of observations between the Dateline and 160°E (Figure 9).

For interannual frequencies, Ekman pumping alone hindcasts unrealistically large amplitudes of variability within about 15° of the equator. At 13°N, the zonal average standard deviation of the Ekman pumping solution to (3.2) was about 25 m, whereas the standard deviation of the ULT solution to (3.1) was about 10 m and that of the observations of 20°C depth about 8 m (all three were higher in the west). The unrealistic magnitude of Ekman pumping is even more pronounced closer to the equator. This contrasts with the results reported in (Kessler, 1989a,b) for the average annual cycle where the amplitude of the Ekman pumping solution was found to be the same as that for the total ULT solution including Rossby waves. The difference is due to the fact that at the annual period there are at most six continuous months of one sign of pumping, while interannual wind stress curl variations frequently persist for several years, so the (integrated in time) interannual Ekman pumping solution can grow very large (bottom panels of Figures 8 and 9). If the dynamics includes Rossby waves, however, these anomalies radiate away before building up large amplitudes, thus the ULT (middle panels) had approximately the same amplitude as the observations (top panels).

Many overall aspects of the forcing were similar over much of the northern tropical Pacific during several El Niños. Upwelling wind stress curl is typically generated to at least 16°N in mid-basin by equatorial westerly wind stress anomalies during the peak of the event (see Section 4c). Along the eastern boundary, 5 to 15 m shallow thermocline anomalies were observed in the year preceding several events (Figure 3); these may also be a characteristic feature. This boundary signal is hindcast to radiate across the basin by the simple vorticity equation (3.1). However, equatorward of about 15°N, this boundary-induced shoaling was usually secondary in its effects on the mid-basin thermocline to that associated with the wind stress curl. In the final phase of the event, the deep equatorial pycnocline anomaly which is the primary signal of El Niño spread up the American coast (Figure 3), with an amplitude of 10-15 m; the ULT solution shows this signal radiating offshore. However, similarly to the shallow anomalies preceding the event, the boundary signal was in many cases a relatively small part of the thermocline deepening observed in the west, most of which was due to downwelling wind stress curl as the equatorial easterlies returned in force. In general, the simple model brings
Figure 8. Top: interannual anomalies of 20°C depth (m) at 13°N during 1970 through 1975. Solid contours indicate deep anomalies, dashed contours shallow anomalies. Middle: hindcast of interannual upper layer thickness (ULT) anomalies (m) made using the vorticity equation (3.1) with best-fit speed $c_r = 23$ cm/s shown in lower right. The heavy line running up from the lower left corner shows this speed. Bottom: Hindcast of interannual pycnocline depth anomalies (m) made assuming Ekman pumping alone (equation 3.2).
Figure 9. As Figure 8 but during 1980 through 1986.
out the fact that although the observations may appear as if a boundary reflection is propagating freely across the entire basin, this can be due to a coincidental occurrence of mid-basin curl-generated anomalies (Kessler, 1989a). It may be that the coincidence of timing has a deeper significance associated with the quasi-periodic nature of El Niño (which cannot be determined here), however the fact remains that at most locations in the western Pacific most of the amplitude of long Rossby waves generated by El Niños was produced in mid-basin by the wind stress curl. An exception is the region north of 15°N, where particular boundary forcing events can be traced for several years across much of the basin and accounted for significant fraction of the variability.

At 13°N during the 1970-75 period (Figure 8), the observations (top) show that in the two years leading up to the El Niño of 1972 the thermocline was deep west of 150°W but shallow in the east. The ULT (middle) has a roughly similar pattern; the deep anomalies in the west were forced in the model by wind stress curl in the final stage of the 1969 El Niño, while the shallow pycnocline in the east has emanated from the boundary. Note that in the east during this period the Ekman pumping solution (bottom) hindcasts anomalies of the opposite sign from those observed, indicating that boundary radiation and not wind stress curl probably accounted for the observed eastern shoaling. During the height of El Niño in mid-1972, the observed thermocline was anomalously shallow everywhere west of 90°W, while deep anomalies appeared in the east as the equatorial deepening arrived at the boundary and spread up the American coast. The simple model (middle) appears to show the shoaling event of 1972 as a continuous upwelling radiating from the eastern boundary and propagating across the basin. However, this boundary signal was quite weak and grew very slowly east of 150°W before amplifying in mid-basin due to the curl (to a much larger size than the original boundary fluctuation), so the western Pacific anomalies cannot be attributed primarily to the boundary reflection. Following the peak of El Niño, deep anomalies on the eastern boundary were observed in late 1972 and then were seen in early 1974 in the west, with some suggestion of propagation. The ULT (middle) makes clear that this is not the result of free propagation of the boundary anomalies, since the speed of free propagation means that waves radiating from the American coast in late 1972 would not have arrived in the west until late 1974; in fact the deep anomalies observed in the west were generated near the Dateline in mid-1973. There is little correlation between the Ekman pumping solution (bottom) and the observations (top); as noted, Ekman pumping hindcasts amplitudes more than twice as large as those observed.

The 1982-83 El Niño (Figure 9) had a somewhat similar sequence of events at 13°N, although observation is more difficult because of the sparser sampling and data gaps. The shallow western Pacific thermocline observed in 1980 (top), followed by a downward peak at the end of 1981, was due to wind stress curl fluctuations associated with the weak El Niño of 1979; these signals are accurately portrayed by the simple vorticity equation (middle). As in the El Niño of 1972, the 1982-83 event began with weak shallow anomalies on the eastern boundary which gained much larger amplitude in mid-basin; this sequence is seen both in the observations (top) and the simple model hindcast (middle). Unlike the El Niño of 1972, during 1982 strong upwelling curl covered the entire 13°N band west of about 150°W, and the Ekman pumping hindcast (Figure 9, bottom) also shows a shoaling event west of 150°W, so the observed shoaling cannot be attributed entirely to the radiation of long Rossby waves, but was locally forced over a wide region.

The eastern boundary deepening following the El Niño of 1982-83 was stronger and longer-lasting than the corresponding event in late 1972, with about a year of anomalies greater than 10 m and a peak of more than 15 m in mid-1983 (Figure 3). Both the observations (Figure 9, top) and the ULT (middle) show this signal apparently radiating across the entire basin at 13°N. However, the maximum deepening in the west at 13°N was observed at the end of 1984, only 18 months after the maximum on the eastern boundary, which is too fast for free propagation (this would imply a speed of about 35 cm/s). In fact, strong downwelling wind stress curl began in early 1983 near 150°E, which started the deepening in the west before the wave arrived from the eastern boundary. The combination of the two forcings produced a deep anomaly in the west which was about twice as large as the original eastern boundary signal (Figure 9, middle). Therefore, as in 1972, it would be an oversimplification to interpret this anomaly as due primarily to free radiation of an eastern boundary anomaly; both the boundary and mid-basin forcing played a strong role.

The above discussion, showing that long extra-equatorial Rossby waves of significant amplitude arrive at the western boundary (whether due to boundary reflection or mid-basin curl), raises the question of whether reflection of these waves from the western boundary as equatorial Kelvin waves could be a contributor to the subsequent evolution of El Niño. As noted in the introduction to this Chapter, such reflection has been suggested as a mechanism for triggering or turning off El Niño. In a recent paper, Graham and White (1988) proposed that downwelling long Rossby waves generated by wind stress curl anomalies near 12°N in mid-basin...
ANOMALIES OF 20°C DEPTH

AVERAGE ANNUAL CYCLE REMOVED

Figure 10. Time series of anomalous 20°C depth (m) at 13°N on the western boundary (solid line) and at the equator on the eastern boundary (dashed line) during 1956 through 1986. Deep anomalies are shown downwards and shallow anomalies upward.

CORRELATION OF 125°E, 13°N with 80°W, EQUATOR

Figure 11. Lag correlation between anomalous 20°C depths at 13°N on the western boundary and at the equator on the eastern boundary (the two time series in Figure 10).
are reflected from the western boundary as downwelling equatorial Kelvin waves and act as triggers for El Niño. Aside from the questions of how efficient such a reflection could be given the irregular nature of the Philippine and Indonesian coastline, and the relatively high latitude, which are beyond the scope of this work (see Battisti, 1989), the data available here can be used to examine the lag relationship between off-equatorial Rossby pulses arriving at the western boundary and the subsequent occurrence of El Niños. To make this comparison long time series are desirable. In section 4.b a long time series of 20°C depth anomalies along the eastern boundary was shown to be a good index of El Niños (Figure 3); this time series can be summarized by the history at the equator along the eastern boundary. Along the western boundary it is not possible to construct a similarly long time series at most latitudes, however one location where data exist back to the 1950's is at 13°N, 125°E, which is where shipping headed for Manila from the east enters the Philippines Archipelago (San Bernardino Strait) (see Figure 1a). This location is almost exactly where Graham and White (1988) have proposed that the long Rossby waves in their scenario would arrive at the western boundary, and since Rossby events large enough to trigger El Niño should have a relatively broad meridional extent, this time series is thought to be representative of extra-equatorial large-scale western boundary variability. The two time series (now referred to as the eastern boundary, equator and the western boundary, 13°N) are overlaid in Figure 10, during 1956 through 1986, which includes eight El Niños (an El Niño is indicated by a deepening of the eastern boundary thermocline). Inspection of these time series shows that each El Niño in fact preceded by a deep event in the west, as suggested by Graham and White (1988). However, there is not a consistent lag relation between the western and eastern deepening as would be expected if the connection between the two was a (nearly linear) oceanic wave process; the observed lag ranges from about three months to about fifteen months.

Second, there were many more western boundary events than subsequent El Niños; during the thirty-one years shown in Figure 10 there were thirteen major western boundary downwellings but only eight El Niños. Given the inconsistency of the lags observed, it would be possible to find western boundary downwelling to precede almost any sequence of events. Third, there is no relation between the magnitude of the western deepening and that of the subsequent eastern event; for example a relatively large event in the west in 1978 was followed by a very weak El Niño in 1979, while a much weaker western downwelling in 1971 was followed by the strong El Niño of 1972. Lag correlation between the two time series (Figure 11) shows only a very weak anticipation of eastern events by those in the west, with correlations just above 0.2 for the west leading by one to one and one-half years, and about 0.3 for the west leading by two and one-half years (which seems much too long for a triggering mechanism). The only significant peak of correlation (r = 0.5) between the two time series occurs with the east leading the west by six months, with opposite sign (Figure 11). This represents the mechanism discussed in the main part of this section, in which upwelling long Rossby waves are generated in mid-basin near the height of El Niños (slightly earlier than when the eastern boundary thermocline peaks downward), so their arrival at the western boundary lags the eastern downward peak by about half a year. Such fluctuations can be seen in the two time series (Figure 10), for example in the 1972 El Niño, where the eastern boundary downward peak occurs in late 1972, while the western boundary shoaling peak at 13°N arrives in early 1973. In sum, while the observations presented here are inadequate to examine the reflection process in detail, the long time series suggest that the Graham and White (1988) scenario is unconvincing in its assertion that extra-equatorial long Rossby waves are the trigger for El Niños. A justifiable conclusion from the long time series is the more modest one that it is unlikely that an El Niño would occur at a time when the western tropical thermocline is anomalously shallow.

5. Summary and conclusions

A new set of BT data makes it possible to examine zonal propagation of thermocline fluctuations without extensive interpolation. The archives of the NODC contain several hundred thousand BT observations in the Pacific which have formed the basis for several previous studies; however these data are primarily distributed along particular shipping routes and leave large gaps, particularly in an important region of the western tropical Pacific. This has resulted in the necessity of relatively elaborate interpolation schemes in order to interpret thermocline variability in terms of zonal long-wave propagation, and has made such interpretation controversial. The new data set consists of MBT observations taken by Japanese fishing vessels during the 1970's; the sampling distribution complements the XBT data set in the western Pacific and allows the gridding of observations on a relatively fine 2° latitude by 5° longitude grid with a minimum of zonal interpolation. On this gridscale, the combined BT observations average one profile every 35 days in each gridbox during 1970.
through 1987, with a higher density during the early 1970’s. The sampling is concentrated in the northern hemisphere and in the southwest Pacific; the southeast Pacific is very poorly sampled. After extensive quality control, more than 185K profiles were available for analysis in the Pacific between 30°N and 30°S.

The field of 20°C isotherm depth was chosen as the clearest, simplest representation of the observed variability. The principal advantage of this choice is that it simplifies the three-dimensional thermal field into a two-dimensional variable which can be mapped and viewed conveniently. The assumption is made that the tropical Pacific may be approximated as a two-layer system divided by a sharp pycnocline, for which the depth of the 20°C isotherm is a proxy. This simplification has commonly been used in observational studies of the tropical Pacific, and in the same spirit many modelling experiments have used a reduced gravity, single-active-layer approximation.

A simple low-frequency long Rossby wave model (3.1) was used to study interannual pycnocline variability, with particular concentration on the El Niño of 1972 when the sampling was most dense. Ekman pumping alone hindcast unrealistically large amplitudes of interannual variability within about 15° latitude of the equator. This was due to the relative persistence of interannual wind stress curl anomalies, which commonly remain of the same sign for several years, so the (integrated in time) Ekman pumping solution can get very large. If the dynamics include Rossby waves, however, the anomalies radiate away before building up large amplitudes, so the simple model solution had approximately the same amplitude as the observations. The interannual hindcast assuming Ekman pumping alone was poorly correlated with the observations almost everywhere, while the Rossby wave solution had high correlations in the western Pacific; neither was significantly correlated in the east central Pacific (Figure 7), a region of poor data coverage (both wind and BT data). The simple model (3.1) was best-fit to the observations of interannual 20°C depth variability to find the phase speed which gave the highest correlations. The planetary wave character of the variability is shown by the meridional profile of these best-fit speeds (Figure 2), which falls roughly on the line $c = \beta v_\pi^2/\rho$, with the long gravity wave speed $c$ chosen to be the zonal average of the first baroclinic mode long gravity speed found by Emery, et al. (1984).

A characteristic zonal wind stress anomaly pattern associated with El Niño is equatorial westerlies during the height of the event in the central Pacific, followed by a return of stronger than average equatorial easterlies the following year (Figure 4). Although the strongest wind stress anomalies are confined to the equatorial zone, the large scale wind stress curl pattern resulting from these anomalies is upwelling curl extending to about 15°N during the equatorial westerlies. In this tropical, extra-equatorial region, upwelling waves were generated during the height of the event which carried the upwelling signal to the western boundary, arriving about six months later (Figures 8 and 9). Thus although the anomalous wind stress curl forcing in the western extra-equatorial Pacific was relatively weak during several El Niños, large-scale shallow thermocline depth anomalies were observed there; these were forced in the central Pacific and radiated west. During the year following several El Niños strong easterlies returned to the equatorial region, generating downwelling wind stress curl in the tropical central Pacific. The deep anomalies forced in this way in the central basin were also observed to propagate to the western boundary.

In addition to the mid-basin forcing, El Niños also generate boundary reflections of the deep equatorial thermocline anomalies which are the primary oceanic signal of the event. Along the American coast, deep anomalies of 20°C depth were observed to occur northward to at least 22°N, with typical amplitudes of 10-15 m (Figure 3). The speed of propagation along the coast was observed to be about 32 cm/s, which is slower than estimates of coastal Kelvin wave propagation found from long time series of sea level. The simple model radiates these anomalies offshore and predicts that they should form an dominant part of the variability in the eastern and central Pacific. However, equatorward of 15°N, the observations do not appear to be consistent with this hypothesis, which is largely why the simple model was found to be poorly correlated with the observations in this region (Figure 7). North of 15°N, however, reflections can be observed propagating across the entire basin over several years (Kessler 1989a,b).

It is clear from the observations reported here that long Rossby waves propagate across the tropical, extra-equatorial north Pacific, and that these waves can arrive at the western boundary with significant amplitude. It has been proposed that these waves reflect from the western boundary as equatorial Kelvin waves and contribute to the triggering of subsequent El Niños (Graham and White, 1988). This mechanism has also been suggested in theoretical and modelling studies which hypothesize that the approximate three to four year periodicity of these events is due to the relatively slow travel time of the long tropical Rossby waves. In a few locations in the Pacific, it is possible to construct long time series of 20°C depths going back to the 1950’s, and
these time series, which span seven major El Niños, indicate problems with this idea. Figure 10 shows such long time series at $13^\circ N$ on the Philippines coast (which gives a good index of waves arriving from the northern tropical Pacific, and is almost exactly where Graham and White have suggested that the triggering waves arrive at the boundary) and at the equator at the South American coast (which is a good index of El Niños). If a (nearly linear) oceanic wave mechanism connects these two locations, then there should be a significantly positively correlated lag relation between them, with the west leading the east. In fact the only significant correlation shows the east leading the west by about six months, with opposite sign. This represents the process discussed above whereby upwelling Rossby waves are generated in mid-basin by wind stress curl associated with equatorial westerly wind anomalies (occurring only slightly before the deep equatorial thermocline anomalies arriving in the east), which then take about six months to reach the western boundary. The proposed correlation with the west leading the east is a weak, broad peak with correlations about 0.2, which is probably not significantly different from zero. There is a suggestion in the time series (Figure 10) that many El Niños are preceded by a deep thermocline in the west, however there are almost twice as many deep events in the west as El Niños, and no apparent relation between the size of the western event and the subsequent eastern deepening, nor is the time lag at all regular, as would be expected if a nearly linear oceanic wave process were the primary mechanism. A more justifiable conclusion from these data is that it is apparently unlikely to have an El Niño occur following a period when the western Pacific thermocline is anomalously shallow. The data do not support the idea that there is a triggering of eastward propagation along the equator by long Rossby waves arriving at the tropical western boundary.

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PROCEEDINGS

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