Contribution of equatorial Pacific winds to southern tropical Indian Ocean Rossby waves

James T. Potemra

Abstract. Westward propagating features, identified as Rossby waves, have been observed and modeled in the southern tropical Indian Ocean (STIO) between 10° and 30°S. These STIO Rossby waves, which have broad zonal and meridional extents, could interact with the westward flowing South Equatorial Current (SEC) as well as coastal currents on the south shore of Java (the South Java Current) and along the western shore of Australia (the Leeuwin Current). Previous work has attributed these waves to variations in wind stress along the west coast of Australia, Ekman pumping in the STIO interior, and a combination of both. This study investigates the importance of a third factor: remotely forced coastal Kelvin waves. Observations show that changes in wind stress curl in the eastern equatorial Pacific create annual upwelling and downwelling Rossby waves. Numerical model results confirm previous studies that demonstrate these waves, upon reaching the western boundary of the Pacific, create poleward propagating coastal Kelvin waves along the western shore of the Irian Jaya/Australia land mass. Direct observations of annual sea level variations along the northwest coast of Australia show a phase lag from the northernmost station to the southernmost that is not explained by direct wind forcing, suggesting that this signal is propagating from the Indonesian seas. It is shown in this work that when these waves reach the Indian Ocean, they are in phase with the local Ekman forcing and enhance the STIO Rossby waves. In the model the signal from the equatorial Pacific accounts for almost 80% of the energy of the STIO Rossby wave near the coast of Australia and 10% of the energy offshore.

1. Introduction

The large-scale circulation in the southern tropical Indian Ocean (STIO), shown schematically in Figure 1, is dominated by the South Equatorial Current (SEC) that flows from east to west between about 10° and 30°S (although the boundaries of this current are not exactly known). The SEC terminates at the western boundary and splits at Madagascar into the northward flowing Zanzibar Current (ZC) and the southward flowing Mozambique Current (MC) and East Madagascar Current (EMC). The origin of the SEC is the South Indian Ocean Current (SIO; also referred to as the Eastern Gyral Current (EGC)) that flows to the east just north of the Antarctic Circumpolar Current (ACC). As the SIOC approaches the west coast of Australia, it turns northward to form the SEC. Along the coast, however, the flow is toward the pole, against the prevailing winds. This current, called the Leeuwin Current (LC), is thought to be driven by an alongshore pressure gradient that is established by a density gradient from the warm, fresh throughflow water [see, e.g., Church et al., 1989; Clarke, 1991; Godfrey and Weaver, 1991; Cresswell and Peterson, 1993]. The dominant variability in the STIO region on the annual timescale is due to westward propagating Rossby waves [e.g., Périaud and Delecluse, 1992; Morrow and Birol, 1998; Yang et al., 1998]. These Rossby waves are evident in both numerical models [Woodberry et al., 1989; Périaud and Delecluse, 1992; Masumoto and Meyers, 1998] and remote observations [Périaud and Delecluse, 1992; Yang et al., 1998; Masumoto and Meyers, 1998; Morrow and Birol, 1998]. The dynamics of these waves, particularly their forcing, and the effect of these waves on the large-scale circulation are still not clear.

Rossby waves in the STIO were first investigated by Woodberry et al. [1989] using a one-and-a-half-layer, reduced gravity model [forced by Hellerman and Rosen-
POTEMRA: EQUATORIAL PACIFIC CONTRIBUTION TO STIO ROSSBY WAVES

Figure 1. Schematic representation of surface flows in the southern tropical Indian Ocean (STIO). Abbreviations for the western boundary currents are defined in the text.

The present study examines a new possibility that remotely forced Rossby waves in the equatorial Pacific contribute to STIO Rossby waves. Annual Rossby waves are formed in the equatorial Pacific from changes in Ekman pumping near the eastern boundary [Meyers, 1979; Lukas and Firing, 1985; Kessler, 1990]. When these Rossby waves reach the western boundary of the Pacific, they induce poleward propagating Kelvin waves along the west side of Irian Jaya and Australia. As these Kelvin waves travel along the western coast of Australia, they excite Rossby waves.

The idea of energy "leakage" from the Pacific to the Indian Ocean was first described by Clarke [1991]. In that study the effects of the gappy western boundary of the equatorial Pacific on large-scale wave reflection were theorized. Clarke [1991] determined that 10% of the energy flux that reaches the western equatorial Pacific due to propagating long Rossby waves makes it into the Indian Ocean. Reflected equatorial Kelvin waves account for 34% of the energy flux, and the remainder gets dissipated in the boundary layers of the various islands that
make up the western boundary of the Pacific [Clarke, 1991].

Verschell et al. [1995], as part of a modeling study, showed numerically that a first-mode Rossby wave (generated in the eastern equatorial Pacific) would provide energy to the Indian Ocean. In the model a free Rossby wave impinged on the western boundary, and energy entered the Indian Ocean (presumably by a coastal Kelvin wave), as is evident by upper layer thickness anomalies along the northwest Australian coast at 10°S.

Other studies have also demonstrated the importance of equatorial Pacific forcing on sea level variations on the northwest Australian coast [e.g., Clarke and Liu, 1994; Wajsowicz, 1996; Murtugudde et al., 1998]. All of these studies, however, focused on interannual variability. In addition, these studies were involved primarily with variations along the northwest coast of Australia and did not directly address the affect these variations would have on STIO Rossby waves.

Here the Pacific to Indian Ocean connection will be investigated with respect to its effect on STIO Rossby waves. Since these waves have been observed annually (one upwelling and one downwelling wave per year), the present study will focus on monthly timescales, unlike the previous studies of the Pacific to Indian Ocean connection that focused on interannual variability (and therefore relatively low frequency Rossby waves).

The key issues that will be addressed include (1) Can sea level variability in the equatorial Pacific propagate to the Indian Ocean on annual timescales; (2) What is the phase relationship of the forcing in the equatorial Pacific and the observed STIO Rossby wave variations (Do remotely forced sea level variations act to enhance or diminish STIO Rossby waves); and (3) How much of the observed variability in STIO sea level is explained by remote Pacific forcing. A theoretical discussion aimed at the first issue is given in section 2 followed by some observational evidence and some results from large-scale general circulation models (GCMs) to address the second and third issues.

2. Theoretical Considerations

Previous studies of STIO Rossby waves have neglected remote effects from the Pacific; however, it has been shown that equatorial Rossby waves in the Pacific could provide energy to the Indian Ocean via coastal Kelvin waves. Clarke [1991] showed that when first-mode, long Rossby waves (interannual period) impinge on the western boundary of the Pacific, 10% of the energy makes it into the Indian Ocean. That work focused on the interannual waves that are created during El Niño-Southern Oscillation (ENSO) events. As part of the derivation, a limit is placed on the horizontal scale and period of the wave. Following Clarke [1991], it is assumed that motion is independent of the zonal extent of the western boundary region, $\Delta x$, which, through expansion, requires

$$\frac{k \Delta x}{2} << 1.$$  \hspace{1cm} (1)

The wave number $k$ is given by $\frac{2 \pi}{\Delta x}$, where $\omega$ is the frequency of the Rossby wave ($\frac{2 \pi}{\Delta x}$ years$^{-1}$ for the interannual waves described by Clarke [1991]) and $c$ is the first baroclinic mode phase speed (given as $3$ m s$^{-1}$ by Clarke [1991]). $\Delta x$ represents the width of the Pacific boundary region, from Kalimantan to the Solomon Islands, and is approximately 4900 km. In the case of interannual Rossby waves, $\omega$ is about $6.6 \times 10^{-8}$ s$^{-1}$, and the left side of (1) is 0.16. Therefore low-frequency Pacific Rossby waves (periods of years) have motions mainly independent of $x$ and leak some energy (10% by Clarke [1991]) into the Indian Ocean, and in fact, it has been shown that interannual variations in northwest Australian sea level are correlated to variations in the equatorial Pacific [e.g., Clarke and Liu, 1994; Wajsowicz, 1996; Murtugudde et al., 1998].

For higher frequencies, however, the left side of (1) is closer to 1. For example, at semiannual periods ($\omega = \frac{\pi}{180}$ days$^{-1}$), $\frac{k \Delta x}{2}$ is 1.13. Therefore it must first be demonstrated that variations in sea level in the equatorial Pacific due to annual Rossby waves will propagate into the Indian Ocean as coastal waves.

Once these waves reach the west coast of Australia and propagate southward, a high-frequency limit is placed on their ability to supply energy to the STIO region. Coastal Kelvin waves along an eastern boundary will generate offshore Rossby waves into the ocean interior only within certain latitudes (so-called turning latitudes; see, for example, McCreary and Kundu [1987]

![Figure 2: Equation (2) was used to compute the critical latitude, given a certain periodic forcing and $c$. The solid line is for the first baroclinic mode ($c = 2.62$ m s$^{-1}$), and the dashed line is for the second baroclinic mode ($c = 1.07$ m s$^{-1}$). The curve shows, for example, for semiannual forcing, first mode coastal Kelvin waves will only generate offshore Rossby waves equatorward of 27° latitude.](image-url)
POTEMRA: EQUATORIAL PACIFIC CONTRIBUTION TO STIO ROSSBY WAVES

where

\[ A_0 = 0.2 \text{ N m}^2; \]
\[ \omega = 2\pi/30 \text{ and } 2\pi/180 \text{ d}^{-1}; \]
\[ x_o = 20^\circ \text{W of the boundary} \]
\[ \text{(170}^\circ \text{W for this example}); \]
\[ y_o = \text{equator}; \]
\[ L_x = 2/L_y; \]
\[ L_y = \sqrt{c/2\beta}. \]

The model’s initial upper layer was set to 300 m thick and a density of 1024 kg m\(^{-3}\). The resulting first baroclinic mode phase speed \( c \) is 2.62 m s\(^{-1}\). The model was run with monthly and semiannual periodic forcing.

Figure 4 shows how the period of the forcing affects the solution. Periodic forcing was used to generate poleward propagating coastal Kelvin waves. The equatorially centered forcing actually creates westward equatorial Rossby waves as well as eastward propagating equatorial Kelvin waves. It is the latter that impinge on the eastern boundary and generate the coastal Kelvin waves.

Energy (per unit volume) was computed at each grid point \((i, j)\) from

\[ E_{i,j} = \frac{\rho}{T} \sum_{t=0}^{t=T} \left[ u_{i,j}^2 + v_{i,j}^2 + P_{i,j}^2 \right] \Delta t, \]

with the pressure \( P \) computed from the model upper layer thickness \( h \) as

\[ P_{i,j} = \frac{g'}{c} (h_{i,j} - h_o). \]

The initial upper thickness \( h_o \) was 300 m, and the period \( T \) was 30 days for the first experiment (Figure 4a) and 180 days for the second (Figure 4b). In each case, coastal Kelvin waves appear on the eastern boundary. Consistent with Figure 2, however, these coastal waves only shed energy into the basin interior within 5° of the equator for the case with 30 day forcing. For lower-frequency coastal waves (180 day forcing), Rossby wave energy is evident within about 30° of the equator. The experiment by Verschell et al. [1995] was forced by an equatorial Pacific Rossby wave with a period of 20 days, which limited the amount of energy that would be generated in the midlatitude Indian Ocean.

3. Application to the STIO Region

The theory outlined in section 2 is now applied to the STIO region. Observations have shown the existence of long Rossby waves in the equatorial Pacific, formed by Ekman pumping near the eastern boundary [Meyers, 1979; Lukas and Firing, 1985; Kessler, 1990]. Further, Péronnau and Delecluse [1992] showed, using remote ob-

Figure 3. The idealized forcing for the rectangular basin model was purely zonal (periodic in time), centered at the equator, 170°W: (a) wind stress along the equator and (b) variations with latitude at 170°W.

for a derivation). The turning latitude is determined as a function of frequency, given by

\[ \tan(\theta_c) = \frac{c}{2\omega R_{\text{Earth}}} \]

Again, \( \omega \) is the frequency of the wave, and \( c \) is the phase speed. \( R_{\text{Earth}} \) is the radius of the Earth, and \( \theta_c \) is the critical latitude. When the latitude is equatorward of \( \theta_c \), the wavenumber is real, and the waves are westward (offshore) propagating Rossby waves. Poleward of this latitude, the wavenumber is complex, and only coastal Kelvin waves are permitted. Figure 2 shows the critical latitude for values of \( \omega \) up to 1 year for two values of \( c \) (approximating the first and second baroclinic modes).

To demonstrate this limit, a simple reduced gravity model was run with idealized forcing. A rectangular grid \((1^\circ \text{ by } 1^\circ)\) representing 50° longitude and 60° latitude (from 30°S to 30°N) was used. To generate poleward propagating coastal Kelvin waves along the eastern boundary (as would be found along the west coast of Australia), the model was forced by zonal, periodic wind stress centered on the equator, 20° west of the eastern boundary (in the STIO case the coastal Kelvin waves would be formed as Pacific Rossby waves interact with the western boundary of the Pacific). The wind stress magnitude (shown in Figure 3) was tapered in both the zonal and meridional directions as follows:

\[ \tau_x = A_0 \sin(\omega t) \exp \left[ -\left( \frac{x - x_o}{L_x} \right)^2 - \left( \frac{y - y_o}{L_y} \right)^2 \right], \]

The model’s initial upper layer was set to 300 m thick and a density of 1024 kg m\(^{-3}\). The resulting first baroclinic mode phase speed \( c \) is 2.62 m s\(^{-1}\). The model was run with monthly and semiannual periodic forcing.

Figure 4 shows how the period of the forcing affects the solution. Periodic forcing was used to generate poleward propagating coastal Kelvin waves. The equatorially centered forcing actually creates westward equatorial Rossby waves as well as eastward propagating equatorial Kelvin waves. It is the latter that impinge on the eastern boundary and generate the coastal Kelvin waves.
servations, that an upwelling and a downwelling Rossby wave is formed each year in the STIO. At 90°E, 12°S, peak sea level occurs in December (downwelling wave), and the minimum sea level signal is in July (upwelling wave).

The question is, given the amplitude, phase, and period of this forcing, can the equatorial Pacific Rossby waves provide energy to the Indian Ocean and contribute to the observed STIO Rossby waves? The hypothesis is that the near-equatorial Pacific Rossby waves can interact with the various islands in the western Pacific (primarily Halmahera, Kalimantan, and Irian Jaya/New Guinea) and create poleward propagating, coastal Kelvin waves. These waves would then travel...
along the western shores of Irian Jaya and Australia into the Indian Ocean. The resulting sea level variations would then contribute to STIO Rossby waves.

As outlined above, the period of the forcing is critical to whether equatorial Pacific Rossby waves will provide energy (in the form of coastal Kelvin waves) to the Indian Ocean and at what latitude these Kelvin waves will generate Rossby waves (if at all). Further, the coastline configuration of the eastern boundary of the Indian Ocean plays a role. The Malay Archipelago stretches from west to east along 10°S (see Figure 1), thus placing a northern boundary on STIO waves formed along the eastern boundary. As discussed previously, the period of the forcing in the eastern Pacific will determine the period of the coastal Kelvin waves and the critical latitude of Rossby wave formation. Thus, for waves with periods less than 60 days, the critical latitude is 10° within the equator, and the Lesser Sunda Islands prevent these waves from entering the Indian Ocean.

Finally, remotely formed coastal Kelvin waves along western Australia will be affected by local wind stress (Ekman pumping). The relative phase of the local wind and the remotely forced waves need to be compared to the phase of the observed STIO Rossby waves. In other words, given a certain phase to the Pacific forcing and a certain phase speed of the generated waves, the remotely forced coastal waves will either enhance or diminish locally forced sea level variations.

The approach taken here is to employ the same one-and-one-half-layer reduced gravity model described previously. Realistic boundaries are used, and both the Pacific and Indian Ocean basins are included. To investigate propagation paths, the same simplified wind forcing is used (see (3)). Wind forcing of monthly, semiannual, annual, and 4 year periods was applied in a region in the eastern equatorial Pacific (see Figure 5). The model was run for 5 years to remove the initial response (15 years in the case of the 4 year period forcing), and results were computed from the final year of integration.

In the model the wind patch excites a westward Kelvin wave that reflects off the eastern boundary of

![Figure 5](image-url)

Figure 5. The wind stress used to forced the idealized model experiments was purely zonal. At its peak, the wind stress attains 0.2 N m\(^{-2}\). (a) The meridional modulation of the wind stress amplitude (in N m\(^{-2}\)) along the equator. (b) The time-varying modulation of the wind stress amplitude for the four experiments: four-year, annual, semiannual, and monthly. (c) Wind stress amplitude (the 0.05 and 0.1 N m\(^{-2}\) levels are contoured). (d) The zonal modulation of the wind stress amplitude (in N m\(^{-2}\)) along 140°W.
Figure 6. Sea level deviations at 15°S, 115°E are shown for the five model experiments. The legend refers to the period of the wind forcing in the eastern Pacific. The wind forcing varies in time as shown in Figure 5.

The relatively simple reduced gravity model forced with realistic winds [Hellerman and Rosenstein, 1983] and realistic coastlines produces STIO Rossby waves similar to observations and more complex global models (this is presented in section 4). The maximum signal in the STIO region near Australia as seen in the T/P data is at about 15°S [Yang et al., 1998]. For the model in this study the maximum is farther south. However, for the following analysis, comparisons will be made along 15°S in the STIO, unless otherwise noted.

Sea level was computed from the model upper layer thickness (long-term mean removed) along 15°S from the eastern Indian Ocean (90°E) to the west coast of Australia (120°E), and the results from three experiments will be discussed (see Figure 7).

For the baseline run (Figure 7d), when climatological wind forcing is applied, annual STIO Rossby waves are evident as high (in March through mid-August) and low (in August through April) sea level deviations along the Australian coast that propagate offshore. The peak values are at 110°E (8 cm in September and -8 cm in March).

Also shown in Figure 7 are sea level perturbations from two additional experiments. In the first experiment, the model was run with wind stress fixed to the mean December values everywhere from 10°S to 10°N. Seasonally varying winds were used everywhere outside the equatorial region. This was intended to eliminate the equatorial Pacific Rossby waves (i.e., to highlight just the locally forced component of the STIO waves). In the second experiment, seasonally varying winds were applied only in the equatorial Pacific (10°S-10°N, 135°E to the eastern boundary of the Pacific). December mean winds were applied everywhere else. In this way the model-produced STIO Rossby waves were purely a result of equatorial Pacific winds. Both results are compared with the baseline model results (seasonally varying wind everywhere) in Figure 7.

It should be noted that a more traditional method of closing the throughflow to eliminate remote Pacific effects was not used here. By closing the throughflow the coastline of the Malay Archipelago gets connected to Australia/New Guinea. Coastal Kelvin waves generated in the Indian Ocean can propagate from Indonesia to Australia and then south along the coast. In reality the gap between Indonesia and Australia is too large for coastal waves to cross. Therefore a regional wind approach was chosen to isolate the effects of the equatorial Pacific from the higher-latitude STIO. Similarly, effects from the equatorial Indian Ocean are not considered for this study and are assumed to be small [see Potemra, 1999].

The two experiments demonstrate that the model STIO signal is a superposition of remotely forced Pacific waves and locally forced waves. For the Hellerman and Rosenstein [1983] wind stress forcing and the model vertical parameterization (which results in a first baroclinic phase speed of 2.62 m s⁻¹), the equatorial Pacific forcing produces high sea level along the model
Figure 7. Model sea level deviations (in centimeters) in the Indian Ocean along 15°S are contoured (negative values shaded). The model was forced with time-varying wind in specific regions: (a) the result when only off-equator winds (outside 10°S-10°N) are allowed to vary, (b) the result when only equatorial Pacific winds (east of 135°E, 10°S-10°N) are allowed to vary, (c) the sum of the results of these two cases, and (d) the baseline model run (winds time-varying everywhere).
Figure 8. Energy was computed from the model velocity and upper layer thickness as the integral of (4) from 20° to 10°S. The solid line is the ratio (in percentage) of the off-equatorial wind case to the total, and the dashed line is the ratio of the equatorial Pacific wind case to the total.

eastern boundary (at 15°S) from early March through August. The maximum signal (7 cm) occurs right at the boundary in May. Lower than average sea level is from September through February, with a peak along the boundary in mid-November of -6 cm. The effects of local wind forcing (no equatorial winds) result in a maximum along the boundary from early May through October, almost 2 months after the Pacific signal, and the maximum signal is offshore at 110°E. This maximum (5 cm) is slightly less than that formed by the equatorial winds, and it does not occur right on the boundary. Negative sea level perturbations prevail the rest of the year along the coast, with a peak of -5 cm at 110°E.

When these results are added, the result is very similar to the baseline model run, with positive sea level perturbations at the coast from mid-March through mid-September and the maxima occurring offshore at 110°E. The sum of the two experiments does not exactly match the baseline run for several reasons, including nonlinearities in the model, equatorial Indian Ocean, and local Indonesian winds and coastal waves that may propagate entirely around Australia.

Figure 8 shows the relative contribution of each signal. Energy (per unit area) was computed as the integral of (4) from 20° to 10°S using the model results from the experiment with only Indian Ocean winds (locally forced waves) and from the experiment with equatorial Pacific winds. The relative energy from each (compared to the sum of the two experiments) is given in Figure 8. As stated, the Pacific effects have the largest effect at the coast, while the local forcing has a large effect in the interior (at 108°E).

As a measure of energy, the model velocity and upper layer thickness fields were integrated over one forcing period (see (4)) for the four experiments (monthly, semiannual, annual, and 4 year forcing). The results, given in Figure 9, show how the maximum STIO Rossby wave energy is produced by the annual forcing. In the case of monthly forcing, almost all of the energy is along the eastern boundary of the Pacific in the form of coastal Kelvin waves. Interestingly, in the semiannually forced case, little energy is seen in the eastern Pacific because of reflected waves canceling the forced waves (because of the location and period of the forcing). This, however, is beyond the scope of the present investigation.

4. Observational and Numerical Evidence

Section 3 demonstrated in a reduced gravity model that equatorially forced waves in the Pacific have a constructive effect on STIO Rossby waves. The question remains whether this connection can be observed in the real ocean.

Direct observations of the connection between the Pacific and Indian Oceans is challenging, however, since the propagation speed is relatively fast while the distance is short. First-mode coastal Kelvin waves that start near the equator (along the west coast of Irian Jaya) would take less than 2 weeks to propagate to 25°S (assuming a phase speed of 3 m s⁻¹).

The repeat orbit of the T/P satellite is not sufficient to discern this signal, but the apparent continuation of a signal from the Pacific to the Indian Ocean can be seen. Sea level deviations (long-term mean removed) from T/P are plotted along 15°S in the Indian Ocean (from 100° to 120°E) and then along 5°N in the Pacific from 120° to 160°E (see Figure 10). For comparison, sea level deviations from the model used in this study (referred to as the University of Hawaii Layered Model (UHLM)) and from two more complex GCMs (the Navy Layered Ocean Model (NLOM) [Wallcraft, 1991], and the Parallel Ocean Climate Model (POCM) [Semtner and Chervin, 1992] are also shown along the same lines).

The models have different grids (specifically, the definition of land points is different in each), and sea level can only be computed from the models to a certain point offshore from the real coastline. The T/P data and the UHLM show high sea level along the Australian coast (at 15°S) during the first half of the year and low sea level during the second half, suggesting that the more simple UHLM is getting the correct phase of the STIO waves.

Propagation from the Pacific is also suggested in the remote observations and model results. Rossby waves in the Pacific are evident along 5°N, and the signal at the western boundary of the Pacific is in phase with the signal at 15°S at the eastern boundary of the Indian Ocean.

While the GCMs and altimeter agree and seem to suggest a link between the Pacific and Indian Oceans, the only direct in situ measurement comes from tide gauges. Daily measurements of sea level are available at five stations along the northwest shore of Australia (a map is given in Figure 11) from 1984-1996 (although some data are missing for the first 2 years at certain locations). Annual and semiannual harmonics were computed for the 10 year period 1987-1996 at each location. The sums of the two harmonics at each location are given in Figure 11.

There is an apparent phase lag between each station,
Figure 9. Energy (in kg m\(^{-1}\) s\(^{-2}\); computed from (4)) from four model experiments with different period forcing.
Figure 10. Sea level (long-term mean removed) is given (in centimeters) along 5°N in the Pacific and along 15°S in the Indian Ocean from: (a) TOPEX/POSEIDON (computed as a climatology from 1993-1998); (b) the model used in this study; (c) the NLOM; and (d) the POCM. Negative values are shaded.
again suggestive of a southward propagating feature. The annual cycle at Darwin shows maximum (minimum) sea level in late February (late July). At the next location (traveling south), Wyndham, the peak is in early March. Continuing south, the annual cycle maximum is early April for Broome and Port Hedland and early June for Carnarvon. The peak in the annual cycle then appears to propagate at about 0.37 m s$^{-1}$. Interestingly, the minima in the annual cycle propagates at a slightly faster speed, about 0.61 m s$^{-1}$. Both are considerably slower than the predicted speed of a first-mode Kelvin wave but could be continental shelf waves [Robinson, 1964; Mysak, 1967; Smith, 1972].

Nevertheless, there seems to be a propagation of annual sea level variability southward along the west coast of Australia that is in phase with the sea level variability of the STIO Rossby waves. To determine if this is actually a propagating signal, or merely local wind forcing, local Ekman pumping was also examined.

Ekman pumping driven by alongshore winds will also create changes in sea level along the coast, and it is important to determine what the local wind effects are to separate them from the freely propagating wave (e.g., the apparent propagation of the signal from Darwin to Carnarvon may be explained in part by a propagation of the alongshore winds).

To investigate the local wind effects, daily wind data were obtained at each tide gauge location for 1993 (from the NCEP reanalysis). The annual cycle from the tide gauges during 1993 is similar to the annual cycle from 1984 through 1996 (compare Figures 11 and 12). Approximating the slope of the shoreline, the NCEP wind
POTEMRA: EQUATORIAL PACIFIC CONTRIBUTION TO STIO ROSSBY WAVES

Figure 12. (left) The 1993 sea level deviations (in centimeters) at five locations and (right) alongshore wind speed (in m s$^{-1}$) from the same locations (from the National Centers for Environmental Prediction (NCEP)). The wind speed is plotted such that positive is alongshore with the coast on the left (downwelling favorable). The annual and semiannual harmonics are shown with the heavy line.

speeds were converted to alongshore and offshore components. The alongshore component, along with the tide gauge data, is shown in Figure 12. The annual cycle (annual and semiannual harmonics) is also given.

There is an apparent propagation of the alongshore wind speed (excluding the record at Broome). Downwelling favorable winds occur at Darwin and Wyndham almost simultaneously, with a peak in early June. Three weeks later the peak occurs at Port Hedland, and 2 weeks after that (mid-July) the peak downwelling occurs at Carnarvon. The propagation of the annual cycle in alongshore wind speed (time of maximum downwelling) is about 1.0 m s$^{-1}$ when the data at Broome are not included (0.93 otherwise). Similar to the tide gauge data, the apparent propagation of the time of maximum upwelling is faster at about 3.7 m s$^{-1}$.

Figure 13 shows the extrema in the annual cycle of sea level and alongshore wind speed (from Figure 12). The apparent phase speed of the downwelling signal is 0.37 m s$^{-1}$, and it is 0.61 m s$^{-1}$ for the upwelling signal. It is perhaps interesting that the alongshore winds are not in the same sense as the tide gauge sea level. The tide gauges record positive sea level deviations early in the year and negative in the middle and end of the year. The alongshore wind speed indicates the beginning of the year (with the exception of Broome) to be upwelling favorable (low sea level) and the middle of the year to be downwelling favorable (high sea level).

5. Discussion

Theoretical constraints place an upper and lower frequency limit on waves that can propagate from the equatorial Pacific through the Indonesian seas and into the interior Indian Ocean as Rossby waves. Interannual Rossby waves in the Pacific (such as those generated during ENSO events) mostly get reflected at the western boundary. Higher-frequency Rossby waves form
coastal Kelvin waves in the Indonesian seas. In the model geometry these coastal Kelvin waves propagate counterclockwise around Australia (until dissipated by friction). If the frequency is too high, offshore Rossby wave solutions are not permitted. For typical stratification, Rossby waves are limited to within 27° of the equator when the forcing is semianual and to within 10° of the equator when the forcing has a 60 day period. The Indonesian archipelago stretches roughly west to east along 10°S, thus blocking the Rossby waves generated by higher frequencies. Forcing within these two ranges, i.e., of order several months to a year, will leak energy into the Indian Ocean and generate offshore Rossby waves.

Numerical model experiments show that STIO Rossby wave solutions are possible for annual forcing in the Pacific. In the model this remote forcing accounts for a significant fraction (up to 80%) of the STIO Rossby wave, particularly near the coast. Since forcing in the Pacific, in a climatological sense, produces sea level changes along the northwest coast of Australia that are in phase with the local STIO Ekman pumping, it is hypothesized that the equatorial Pacific winds act to enhance STIO Rossby waves and that this adds a third factor (along with local coastal wind fluctuations and Ekman pumping in the STIO interior) to be considered in the forcing of STIO Rossby waves.

Forcing in the Pacific, in a climatological sense, produces sea level changes along the northwest coast of Australia that are in phase with the local STIO Ekman pumping. It is therefore hypothesized that the equatorial Pacific winds act to enhance STIO Rossby waves.

Observational evidence of wave propagation from the Pacific to the Indian Ocean is not available. The typical sampling frequency of altimeters is too low to capture the coastal waves (which take about 2 weeks to propagate from the equator to 25°S off the west coast of Australia). Tide gauge data, available at five locations along the west coast of Australia, does show a propagating feature that is not fully explained by changes in alongshore winds.

Phase differences in the tide gauge records along the northwest coast of Australia support the idea that a signal is propagating from north to south along the coast.
This phase difference in the sea level record was discussed by Godfrey and Golding [1981]. Godfrey and Ridgway [1985] also discuss this propagating signal as originating in the Indonesian seas. They explain that the signal at Darwin could originate from upwelling in the Banda and Arafura Seas during the northwest monsoon, which is maximum in February and March.

The observed phase speed is lower than a first-mode coastal Kelvin wave. While the simple reduced gravity model can only support this type of wave, this region could support a continental shelf wave [e.g., Mysak, 1967; Mysak, 1968a; 1968b]. These were first observed in this region by Hamon [1966]. The phase speed for a shelf wave depends, among other things, on the slope and width of the shelf. Hamon [1966] got a phase speed close to the first-mode Kelvin wave but used data from the southwestern and eastern coasts where the shelf is much more narrow. On the northwestern coast the shelf extends for over 100 km, and the phase speed for shelf waves is inversely related to the shelf width. The model, however, does not resolve the shelf, and the only allowable solution is the Kelvin wave. The exact mechanism, however, is for future studies. In any case the observed annual sea level variability propagates south and is in phase with the STIO Rossby waves.

The change in phase speed from 0.4 m s\(^{-1}\) for the downwelling wave (in February through May) to 0.6 m s\(^{-1}\) for the upwelling wave (in mid-July through September) is also interesting. This could be explained by a seasonal change in the stratification off the northwest Australian shelf. The slow downwelling signal occurs in mid-February (at Darwin) to late May (at Carnarvon). During this time the density gradient setup by the flow from the Indonesian seas is weak (maximum throughflow occurs in northern summer). Levitus data [Levitus, 1982] show a gradient of 1.5 kg m\(^{-2}\) between the surface waters at Darwin and Carnarvon in November and about 0.6 kg m\(^{-2}\) in March. In addition, the mixed layer deepens from a minimum in January through March to a maximum in June. The relatively shallow mixed layer early in the year could also cause the wave to slow.

More observations, particularly higher temporal resolution data, are required to explain fully the coastal signal off northwest Australia. Nevertheless, the observations and model experiments suggest that energy does propagate from the Pacific to the Indian Ocean and that this represents an important consideration for the formation of STIO Rossby waves.

Acknowledgments. The author gratefully acknowledges the assistance of Dennis Moore, Mark Merrifield, and Bo Qiu for many helpful discussions and comments to the draft. Jay McCreary and the rest of the IPRC staff were instrumental throughout this research effort. The tide gauge data were provided by the Pacific Sea Level Center at the University of Hawaii. Funding was provided by JAMSTEC grant 434957 and NSF grant OCE 9819511. The International Pacific Research Center is partly sponsored by Frontier Research System for Global Change. SOEST contribution number 5269 and IPRC contribution number IPRC-60.

References


J. T. Potemra, School of Oceanography, University of Washington, Box 357940, Seattle, WA 98195. (jimp@ocean.washington.edu)

(Received August 2, 1999; revised September 19, 2000; accepted October 13, 2000.)