Tropical Pacific Basin-Wide Adjustment and Oceanic Waves

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Abstract. This study examines the basin-wide adjustment process associated with El Niño-Southern Oscillation (ENSO) in terms of oceanic waves using the Cane-Zebiak model. The Rossby waves induce not only the meridional mass divergence in the ocean interior but also the incoming zonal mass flux at the western boundary, which generally overcompensates the interior meridional divergence. Therefore, the Rossby waves alone cannot make a negative feedback. However, the outgoing zonal mass flux associated with the reflected Kelvin wave at the western boundary reduces the incoming boundary mass flux associated with the Rossby waves, resulting in the tendency of equatorial basin-wide heat content varying in phase with the meridional divergence. Also shown is that the nonequilibrium adjustment of the tropical Pacific Ocean is mainly due to the free Kelvin waves reflected at the western boundary.

The ENSO has been recognized as a self-sustained oscillating phenomena whose origin is an ocean-atmosphere coupled instability and whose oscillatory nature is attributed to the slow oceanic adjustment. The slow oceanic adjustment has been explained by either '*delay oscillator*' theory in terms of retarded equatorial wave motion [*Battisti and Hirst*, 1989] or '*recharge oscillator*' theory in terms of the recharge/discharge of the equatorial total mass and heat content [*Jin*, 1996; *Jin and An*, 1999; *An and Kang*, 2000]. Two theories commonly emphasize the slow ocean adjustment but explain it in a different manner.

The delay oscillator theory attributes the negative feedback mechanism to the free Kelvin waves generated by a reflection of Rossby waves at the western boundary. In the recharge oscillator mechanism, on the other hand, the negative feedback mechanism is achieved by the slow adjustment of the total heat content/mass due to the meridional mass flux divergence and the zonal mass flux at the eastern and western boundaries [*An and Kang*, 2000; *Jin*, 1996]. The recharge oscillator appears not to consider the western boundary reflection. However, the reflection affects the equatorial heat content by modifying the zonal fluxes, particularly at the western boundary, such that the net boundary flux is reduced by the increase of reflection at the western boundary. This is because at the boundary, the zonal current associated with the reflected Kelvin wave has an opposite sign from that of the forced Rossby waves.

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The ocean adjustment of the tropical Pacific basin explained by the recharge oscillator can be thought of with the tropical Rossby and Kelvin wave processes. The meridional transport and the change of zonal-mean equatorial heat content in the recharge oscillator, respectively, are accomplished by the Rossby waves and the reflected free Kelvin waves. The delayed oscillator does not explicitly discuss the meridional transport, but the free Kelvin wave is in fact linked to the forced Rossby waves by the western boundary reflection. The reflected Kelvin wave changes the zonal-mean thermocline depth with a little time delay, because of the relatively fast propagation of free Kelvin wave. Therefore, the recharge oscillator theory and the delayed oscillator theory can be explained with a same-wave dynamics. In the present study, we investigate the basin-wide adjustment of ENSO in terms of the equatorial waves, with a particular attention to the free Kelvin wave, using the Cane and Zebiak model [*Zebiak and Cane*, 1987] (hereafter CZ model). We also provide a theoretical background for the delay time of equatorial wave actions necessary for the cyclic oscillation of ENSO by using a simple analog model.

In a linear shallow water system, the tendency equation of zonal-mean heat content $[h]$ can be written as [*An and Kang*, 2000]

$$[h]_t = Hu_w/L - H[v]_y - a[h], \qquad (1)$$

where [ ] indicates the zonal average over the Pacific basin; u_w is the zonal current at the western boundary; v the meridional current; L the ocean basin length; H the mean thermocline depth of 150 m; a the damping rate of (2.5 years)$^{-1}$. The zonal current at the eastern boundary was neglected under the non-slip boundary condition [*Cane and Sarachik*, 1981], and u_w consists of the outgoing free Kelvin wave from the western boundary (u_K) and the incoming Rossby waves (u_R). The relation between u_K and u_R is determined by the meridional integral of u_w vanishing at the western boundary [*Cane and Sarachik*, 1981]. Then, for the average between 5°S and 5°N, Eq. (1) can be written

$$\begin{aligned} < h >_t = H < u_K > /L + H < u_R > /L \\ -H([v]_n - [v]_s) - a < h > . \end{aligned} \qquad (2)$$

Here, $<>$ indicates the zonal average between 5°S and 5°N, and $([v]_n - [v]_s)$ is the meridional divergence term due to the zonal-mean meridional currents at the southern ($[v]_s$) and northern ($[v]_n$) boundaries. The free Kelvin wave directly contributes to the zonal-mean height through the first term of the right hand side of Eq. (2).

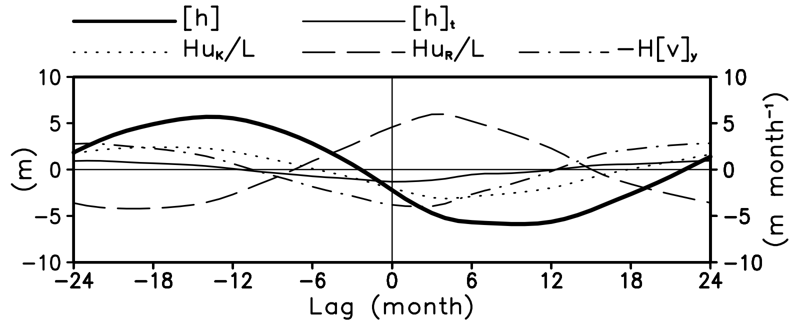

Figure 1. Lag covariance between the normalized Niño-3 index and the equatorial zonal-mean thermocline depth (thick solid line), the mass flux due to zonal currents associated with Kelvin ($H < u_K > /L$; dotted line) and Rossby ($H < u_R > /L$; dashed line) waves at the western boundary, the net mass divergence due to meridional current (-$H < [v]_y >$; dash-dotted line) in the ocean interior, and the tendency of the zonal-mean thermocline depth ($< h >_t$; thin solid line). All variables are averaged over 5°S-5°N from west to east. Units for the thermocline depth are m (left axis) and those for others are m month^{-1} per unit mass (right axis).

Figure 1 shows the lagged covariance between the normalized Niño-3 index (SST anomaly averaged over 150°W-90°W and 5°S-5°N) and each component in Eq. (2), in which the CZ model output integrated for 100 years was used. The amplitude of zonal mass flux at the western boundary associated with the Rossby waves is larger than that of the meridional divergence term and they have opposite sign. $H < u_R > /L$ has a time lag of about 4 months with the Niño-3 index. The excessive zonal mass influx associated with the Rossby waves at the western boundary during a positive SST phase can accumulate the mass. However, the zonal current associated with the reflected Kelvin wave at the western boundary ($H < u_K > /L$), which has a sign different from that associated with the Rossby waves, reduces the influx at the western boundary. The reduction of the boundary flux by the Kelvin wave insures the meridional divergence to control the height tendency shown by the thin line. Note that $< u_K >$ has a time lag of about 2 months with $< u_R >$ at the western boundary. As a result, the negative maximum of the free Kelvin wave contribution appears 6 months later than that of the SST maximum. The free Kelvin wave contribution lags the height tendency by about 6 months. As will be discussed in the next section, the 6-month delay time due to the free Kelvin wave plays an important role of negative feedback.

To examine how the free Kelvin wave controls the zonal-mean height over the Pacific basin, the height associated with the free Kelvin wave is calculated by using the following wave equation [*Kang and An*, 1998]:

$$h_K(x,y,t) = \chi \int_{-\infty}^{+\infty} u_R(x_w, t-\Delta t)dy \bullet \Psi_0(y), \qquad (3)$$

where u_R is the zonal current due to Rossby wave; x_w the western boundary; C_o the Kelvin wave speed of 2.9 m s^{-1}; Ψ_0 the zero-order hermit function; χ a constant value to get a physical dimension; $\Delta t = (x - x_w)/C_o$. As shown in Fig. 2a, the variations of $[h]$ and $[h_K]$ along the equator show different behavior with respect to the change of the Niño-3 index. The amplitude of $[h_K]$ is larger than that of $[h]$, and $[h_K]$ slightly leads $[h]$. *Kang and An* [1998] also showed a similar result with a simple coupled model. In the equatorial region, the zonal-mean height is also contributed by the forced Kelvin wave, which can be estimated by $[h]$-$[h_K]$. The forced part has an opposite sign from that of the free Kelvin wave. As a result, $[h_K]$ is larger than $[h]$, and $[h_K]$ leads $[h]$ by a few months. For the domain extended to 5°S-5°N (Fig. 2b), on the other hand, the behavior of $< h_K >$ including the amplitude and phase change with respect to the Niño-3 index is almost same as that of $< h >$. Therefore, the variations of total heat content over the tropical Pacific are controlled mainly by the free Kelvin wave. It is due to a cancellation between the forced-Kelvin and forced-Rossby waves in the domain average. As a result, the memory of the ENSO oscillation is mostly contained in the free Kelvin wave.

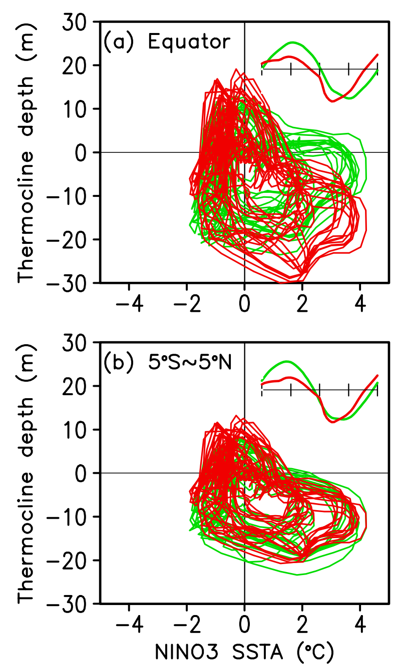

Figure 2. Trajectory plot for the zonal-mean thermocline depth anomaly ($[h]$) and Niño-3 index (green line), and that for the zonal-mean thermocline depth anomaly due to the free Kelvin ($[h_K]$) and Niño-3 index (red line). Each variable is averaged over (a) equator and (b) 5°S-5°N from west to east. Units for SST and thermocline depth are $^\circ$C and m, respectively. At upper right corners in each panel, lag correlation between $[h]$ (green line) and Niño-3 index, and that for $[h_K]$ (red line) are shown. The center tick indicates the zero lag, negative and positive lags go to the left and right from the center, respectively, and the interval between the adjacent ticks is 12 months.

To understand how the delay time scale of free Kelvin wave plays an important role for ENSO, we develop an analog model representing the tropical Pacific adjustment process. The Sverdrup balance is known to be a good approximation for expressing the equatorial ocean states [*An and Jin*, 2001; *Cane and Sarachik*, 1981]. The Sverdrup balance equation can be further approximated as Eq. (4) for the heights averaged over the equatorial eastern (h_E) and western (h_W) Pacific [*Jin*, 1996]:

$$h_E - h_W = a_L \tau(t), \tag{4}$$

where $\tau(t)$ represents the zonal wind stress averaged over the central Pacific; $a_L = L/(2\rho_o g\prime H)$, which has a value of 740 $m^3 N^{-1}$. Here, h_E is induced not only by the direct local wind forcing ($a_L \tau(t)$) but also by the remote process associated with the reflected free Kelvin waves ($-a_w \tau(t-s)$) [*Battisti and Hirst*, 1989]. The time lag (s) between the wind forcing and the free Kelvin wave is determined by the time taken for the Rossby wave to propagate from its forced region to the western boundary and that for the Kelvin wave to propagate from the western boundary to the eastern Pacific. Hence one can write h_E and h_W in terms of the wind stress:

$$h_E = a_L \tau(t) - a_w \tau(t-s) \; and \; h_W = -a_w \tau(t-s), \tag{5}$$

where a_w is related to the wind stress curl to generate the Rossby wave and the western boundary reflectivity. *Battisti and Hirst* [1989] suggested that $700 < a_L < 800$ ($m^3 N^{-1}$) and $440 < a_w < 540$ ($m^3 N^{-1}$). The zonal mean height associated with the free Kelvin wave $[h_K]$ is $-a_w \tau(t-s)$, and the zonal mean of h can be expressed as

$$[h] = 0.5 a_L \tau(t) + [h_K]. \tag{6}$$

Thus, the difference between $[h]$ and $[h_K]$ is $0.5 a_L \tau(t)$, which is proportional to the wind stress without a time lag, and $[h_K]$ is larger than $[h]$. The zonal-mean thermocline depth change associated with $0.5 a_L \tau(t)$ represents the basin average of the forced Kelvin wave established only in the eastern half of the basin. Then, the non-equilibrium process in the total heat content is mainly attributed to the free Kelvin wave.

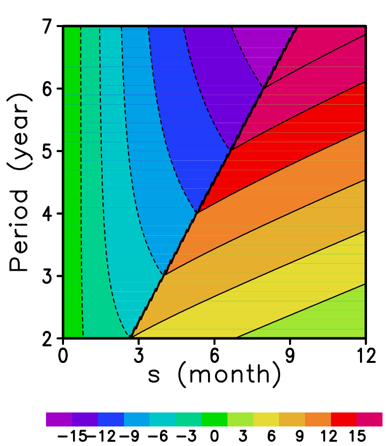

Figure 3. Time lag (φ) in month between the zonal-mean thermocline depth and wind forcing anomalies at the equator under the various periods (year in *y-axis*) and lag s (month in *x-axis*) derived from Eq. (7).

Using Eq. (6), we can get the following phase relationship among $\tau(t)$, $[h]$, and $[h_K]$, for a given oscillation frequency, ω:

$$[h] = (a_L/2)^2(\gamma^2 - 2\gamma cos(\omega s) + 1)\tau(t+\varphi),$$
$$\varphi = arctan(\gamma sin(\omega s)/(1 - \gamma cos(\omega s)))/\omega, \tag{7}$$

where φ indicates the time lag between $[h]$ and τ(t), and γ=2a$_w$/a$_L$. Thus, φ depends not only on the delayed time scale but also on the relative efficiency of wind stress to generate the forced and free Kelvin waves (γ). The positive (negative) φ indicates that $[h]$ leads (lags) $\tau(t)$. Fig. 3 shows φ as a function of various values of ω and s for a fixed value of γ =1.3. As shown in the figure, φ has a negative value for a relatively short delay time. In this parameter range, the wind (SST) forcing leads $[h]$ so that oscillation does not occur. Therefore, a sufficiently large value of delay time is necessary for the oscillation with [h] leading τ. As shown in the previous section, the CZ model has an oscillation period of about 4 years, the delay time of 6 months, and φ about 10 months. The CZ model results are remarkably consistent with what Fig. 3 shows, indicating the simple analog model explains the essence of ENSO.

For the domain expanding from the equator to the off-equatorial band between 5°S and 5°N, Eq. (5) should be replaced by the following equations:

$$\eta_E = \rho a_L \tau(t) - \rho a_w \tau(t-s),$$
$$\eta_W = -\rho a_w \tau(t-s) - (a_w/r_w)\tau(t), \tag{8}$$

where η_E and η_W are, respectively, the heights averaged over the eastern and western Pacific between 5°S and 5°N; ρ is the amplitude factor related to the zonally symmetric shape of the equatorial Kelvin wave; $-(a_w/r_w)\tau(t)$ indicates the forced Rossby wave component, which is almost zero at the equator but increases as the domain expands poleward; r_w is the reflectivity at the western boundary that is about 0.71 for the first symmetric Rossby wave [*Kang and An*, 1998]. Then, the zonal-mean height can be represented by

$$[\eta] = -\rho a_w \tau(t-s) + (\rho a_L - a_w/r_w)\tau(t)/2. \tag{9}$$

For the realistic parameter values (a_L=750, a_w=490, ρ=0.92, and r_w=0.71), $\rho a_L - a_w/r_w$ becomes almost zero. Thus, we have $[\eta] \approx -\rho a_w \tau(t-s)$. It indicates that the variation of the zonal-mean height is almost in phase with that of the reflected Kelvin wave when the domain is 5°S-5°N, resulting from the cancellation between the forced Kelvin and Rossby waves. This is consistent with the CZ model results, the variance of forced Kelvin wave (31 m^2) similar to that of the forced Rossby wave (39 m^2) over the domain of 5°S-5°N and the results as shown in Fig. 2b.

In conclusion, the Rossby waves alone cannot make a negative feedback. Since the outgoing zonal mass flux associated with the reflected Kelvin wave at the western boundary reduces the incoming mass flux associated with the Rossby waves, the tendency of equatorial basin-wide heat content is in phase with the meridional divergence. For a typical 4-year cycle of ENSO, the 6-month delayed response of free Kelvin waves to the wind stress forcing together with the simultaneous response of forced Kelvin wave are responsible for the equatorial basin-wide heat content leading the tropical SST anomalies by about one year.

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