On the ENSO Mechanisms①

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ABSTRACT

The El Niño–Southern Oscillation (ENSO) is an interannual phenomenon involved in the tropical Pacific Ocean–atmosphere interactions. The oscillatory nature of ENSO requires both positive and negative ocean–atmosphere feedbacks. The positive feedback is dated back to Bjerknes' hypothesis in the 1960s, and different negative feedbacks have been proposed since the 1980s associated with the delayed oscillator, the western Pacific oscillator, the recharge–discharge oscillator, and the advective–reflective oscillator. The delayed oscillator assumes that wave reflection at the western boundary provides a negative feedback for the coupled system to oscillate. The western Pacific oscillator emphasizes equatorial wind in the western Pacific that provides a negative feedback for the coupled system. The recharge–discharge oscillator argues that discharge and recharge of equatorial heat content cause the coupled system to oscillate. The advective–reflective oscillator emphasizes the importance of zonal advection associated with wave reflection at both the western and eastern boundaries. All of these physics are summarized in a unified ENSO oscillator. The delayed oscillator, the western Pacific oscillator, the recharge–discharge oscillator, and the advective–reflective oscillator can be extracted as special cases of the unified oscillator. As suggested by this unified oscillator, all of the previous ENSO oscillator mechanisms may be operating in nature.

Key words: El Niño–Southern Oscillation, ENSO theory, Climate variability, Ocean–atmosphere interactions

1. Introduction

The earliest studies for causing tropical Pacific climate variability associated with the El Niño–Southern Oscillation (ENSO) can be dated back to Bjerknes (1966, 1969). Bjerknes provided evidence that the long–term persistence of climate anomalies associated with Walker's Southern Oscillation (Walker and Bliss, 1932) is closely associated with slowly evolving sea surface temperature (SST) anomalies in the equatorial eastern and central Pacific. Bjerknes hypothesized that a positive ocean–atmosphere feedback involved in the Walker circulation is responsible for the SST warming observed in the equatorial eastern and central Pacific. Consider an initial positive SST anomaly in the equatorial eastern Pacific. This anomaly reduces the zonal SST gradient and hence the strength of the Walker circulation, resulting in weaker trade winds around the equator. The weaker trade winds in turn drive the ocean circulation changes that reinforce SST anomaly. This positive ocean–atmosphere feedback or coupled ocean–atmosphere instability leads the equatorial Pacific to a never–ending warm state. During that time, Bjerknes did not know what causes a turnabout from a warm phase to a cold phase.

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Since Bjerknes' hypothesis, ENSO has not been intensively studied until the 1980s. The intense warm episode of the 1982–83 El Niño, which started without being noticed, galvanized the tropical climate research community to understand ENSO and finally predict ENSO. Besides the Bjerknes' positive feedback, the oscillatory nature of ENSO needs a negative feedback to turn the coupled ocean–atmosphere system from a warm (cold) phase to a cold (warm) phase. Four conceptual oscillators that include different negative feedbacks have been proposed: 1) the delayed oscillator (Suarez and Schopf, 1988; Graham and White, 1988; Battisti and Hirst, 1989; Cane et al., 1990); 2) the western Pacific oscillator (Weisberg and Wang, 1997b; Wang et al., 1999b); 3) the recharge–discharge oscillator (Jin, 1997a, b); 4) the advective–reflective oscillator (Picaut et al., 1997). The delayed oscillator assumes that the western Pacific is an inactive region and wave reflection at the western boundary provides a negative feedback for the coupled system to oscillate. The western Pacific oscillator emphasizes equatorial wind anomalies in the western Pacific that provide a negative feedback for the coupled system to oscillate. The recharge–discharge oscillator argues that discharge and recharge of equatorial heat content cause the coupled system to oscillate. The advective–reflective oscillator emphasizes the importance of zonal advection associated with wave reflection at both the western and eastern boundaries.

Section 2 of this paper will review and discuss all of the four ENSO oscillator models. A unified oscillator model, including the physics of these different ENSO oscillators, is presented in Section 3. Section 4 shows that all of the ENSO oscillators can be extracted as special cases of the unified ENSO oscillator. Section 5 gives a summary and discussion.

2. Different conceptual ENSO oscillators

2.1 *The delayed oscillator*

A mechanism for the oscillatory nature of ENSO was originally proposed by McCreary (1983), based on the reflection of subtropical oceanic upwelling Rossby waves at the western boundary. McCreary (1983) explored shallow water ocean dynamics coupled to wind stress patterns that are changed by discontinuous switch depending on thermocline depth and demonstrated how oceanic Rossby waves might be involved in generating the low-frequency oscillations associated with ENSO. Suarez and Schopf (1988) introduced the conceptual delayed oscillator as a candidate mechanism for ENSO, by considering the effects of equatorially trapped oceanic waves propagating in a closed basin through a delay term (Fig. 1). Based on the coupled ocean–atmosphere model of Zebiak and Cane (1987), Battisti and Hirst (1989) formulated and derived a version of the Suarez and Schopf (1988) conceptual delayed oscillator model and argued that this delayed oscillator model could account for important aspects of the numerical model of Zebiak and Cane (1987).

The conceptual delayed oscillator model is represented by a single ordinary differential delay equation with both positive and negative feedbacks

$$\frac{dT}{dt} = AT - BT(t-\eta) - \varepsilon T^3, \tag{1}$$

where T is the SST anomaly in the equatorial eastern Pacific, A, B, η, and ε are constant model parameters. The first term of the right hand side of Eq. (1) represents the positive feedback by local ocean–atmosphere coupling in the equatorial eastern Pacific. The second term is the delayed negative feedback by free Rossby waves generated in the eastern Pacific coupling region that propagate to and reflect from the western boundary, returning as Kelvin

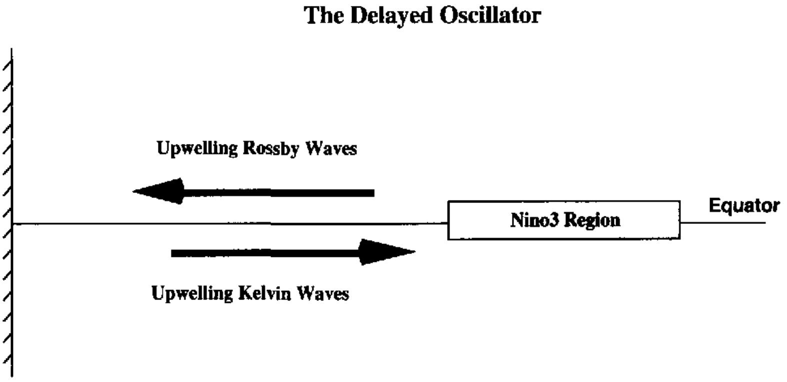

Fig. 1. Schematic diagram of the delayed oscillator for ENSO. The delayed oscillator model has a positive feedback and a negative feedback, assuming that the western Pacific is an inactive region. The positive feedback is represented by local ocean–atmosphere coupling in the equatorial eastern Pacific. The delayed negative feedback is represented by free Rossby waves generated in the eastern Pacific coupling region that propagate to and reflect from the western boundary, returning as Kelvin waves to reverse the anomalies in the eastern Pacific coupling region. Thus, the coupled ocean–atmosphere system oscillates on interannual time scales.

waves to reverse the anomalies in the eastern Pacific coupling region. The last term is a cubic damping term. The delayed oscillator model of Eq. (1) can oscillate on interannual time scales over a broad range of model parameters (McCreary and Anderson, 1991).

Graham and White (1988) presented observational evidence of off–equatorial Rossby waves and their reflection at the western boundary and then empirically constructed a conceptual oscillator model for ENSO. As shown in McCreary and Anderson (1991), the conceptual equations of the Graham and White model can be reduced to a single equation that has similar form to the delayed oscillator [also see the comments of Neelin et al. (1998)].

2.2 *The western Pacific oscillator*

Observations show that ENSO displays both eastern and western Pacific interannual anomaly patterns (e.g., Rasmusson and Carpenter, 1982; Wang et al., 1999b; Wang and Weisberg, 2000). During the warm phase of ENSO, warm SST and low sea level pressure (SLP) anomalies in the equatorial eastern Pacific and low outgoing longwave radiation (OLR) anomalies in the equatorial central Pacific are accompanied by cold SST and high SLP anomalies in the off–equatorial western Pacific and high OLR anomalies in the off–equatorial far western Pacific. Also, while the zonal wind anomalies over the equatorial central Pacific are westerly, those over the equatorial western Pacific are easterly. The nearly out–of–phase behavior between the eastern and western tropical Pacific is also observed during the cold phase of ENSO, but with anomalies of opposite sign.

The delayed oscillator only considers the ENSO eastern Pacific anomaly patterns and overlooks the western Pacific anomaly patterns. Consistent with observations and numerical modeling results (e.g., Weisberg and Wang, 1997a; Mayer and Weisberg, 1998; Wang et al., 1999b; Wang and Weisberg, 2000; Wang, 2000), Weisberg and Wang (1997b) developed a conceptual western Pacific oscillator model for ENSO (Fig. 2). This model emphasizes the role of the western Pacific in ENSO which has been overlooked in the delayed oscillator. In particular, off-equatorial SST and SLP anomalies in the western Pacific induce equatorial western Pacific wind anomalies that affect the evolution of ENSO. The western Pacific oscillator can be represented by the following equations

$$\frac{dT}{dt} = a\tau_1 + b_2\tau_2(t-\delta) - \varepsilon T^3, \tag{2}$$

$$\frac{dh}{dt} = -c\tau_1(t-\lambda) - R_h h, \tag{3}$$

$$\frac{d\tau_1}{dt} = dT - R_{\tau 1}\tau_1, \tag{4}$$

$$\frac{d\tau_2}{dt} = eh - R_{\tau 2}\tau_2, \tag{5}$$

where T is the SST anomaly in the equatorial eastern Pacific, h is the thermocline depth anomaly in the off-equatorial western Pacific, τ_1 and τ_2 are the equatorial zonal wind stress

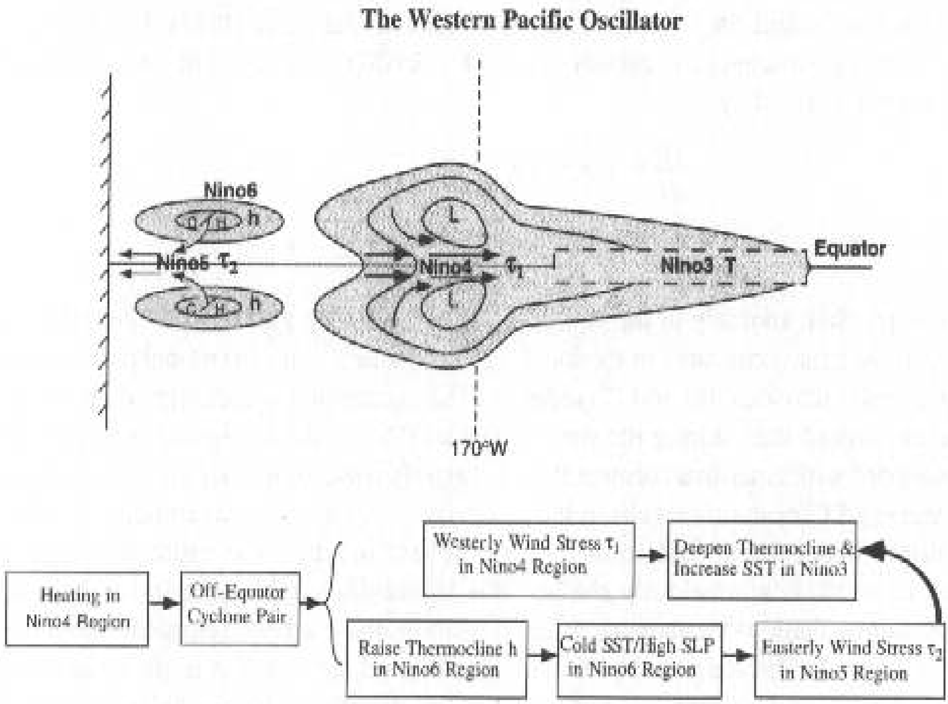

Fig. 2. Schematic diagram of the western Pacific oscillator for ENSO. Condensation heating in the central Pacific induces a pair of off-equatorial cyclones with westerly wind anomalies in the Nino4 region. The Nino4 westerly wind anomalies act to deepen the thermocline and increase SST in the Nino3 region, providing a positive feedback for anomaly growth. On the other hand, the off-equatorial cyclones raise the thermocline there via Ekman pumping. Thus, a shallow off-equatorial thermocline anomaly expands over the western Pacific leading to a decrease in SST and an increase in SLP in the Nino6 region. The Nino6 high SLP initiates equatorial easterly wind anomalies in the Nino5 region. The Nino5 easterly wind anomalies cause upwelling and cooling that proceed eastward, providing a negative feedback for the coupled ocean-atmosphere system to oscillate on interannual time scales.

anomalies in the central Pacific and the western Pacific, respectively. All of the model parameters are constant.

Arguing from the vantage point of a Gill (1980) atmosphere, condensation heating due to convection in the equatorial central Pacific (Deser and Wallace, 1990; Zebiak, 1990) induces a pair of off–equatorial cyclones with westerly wind anomalies on the equator. These equatorial westerly wind anomalies act to deepen the thermocline and increase SST in the equatorial eastern Pacific, thereby providing a positive feedback for anomaly growth. On the other hand, the off–equatorial cyclones raise the thermocline there via Ekman pumping. Thus, a shallow off–equatorial thermocline anomaly expands over the western Pacific leading to a decrease in SST and an increase in SLP in the off–equatorial western Pacific (e.g., Wang et al., 1999b). During the mature phase of El Niño, the off–equatorial high SLP anomalies initiate equatorial easterly wind anomalies in the western Pacific. These equatorial easterly wind anomalies cause upwelling and cooling that proceed eastward as a forced ocean response providing a negative feedback for the coupled ocean–atmosphere system to oscillate.

2.3 *The recharge–discharge oscillator*

Based on Wyrtki's (1975, 1986) suggestion on the buildup and relaxation of sea level over the western Pacific and the numerical model of Zebiak and Cane (1987), Jin (1997a, b) proposed a recharge–discharge oscillator model for ENSO (Fig. 3). The recharge–discharge oscillator is represented by

$$\frac{dT}{dt} = CT + Dh - \varepsilon T^3, \tag{6}$$

$$\frac{dh}{dt} = -ET - R_h h, \tag{7}$$

where T is the SST anomaly in the equatorial eastern Pacific, and h is the thermocline depth anomaly in the equatorial western Pacific. C, D ,ε, E, and R_h are the model parameters.

Jin claimed that Eqs. (6) and (7) represent the discharge and recharge of equatorial heat content. He argued that, during the warm phase of ENSO, the divergence of Sverdrup transport associated with equatorial central Pacific westerly wind anomalies and equatorial eastern Pacific warm SST anomalies results in the discharge of equatorial heat content. The discharge of equatorial heat content leads to a transition phase in which the entire equatorial Pacific thermocline depth is anomalously shallow due to the discharge of equatorial heat content. This anomalous shallow thermocline at the transition phase allows anomalous cold waters to be pumped into the surface layer by climatological upwelling and then leads to the cold phase. The converse occurs during the cold phase of ENSO. It is the recharge–discharge process that makes the coupled ocean–atmosphere system oscillate on interannual time scales.

2.4 *The advective–reflective oscillator*

Picaut et al. (1996) found that zonal displacement of the oceanic convergence zone at the eastern edge of the western Pacific warm pool is in phase with the Southern Oscillation Index. Based on this finding and the study of Picaut and Delcroix (1995) regarding wave reflection, Picaut et al. (1997) proposed a conceptual advective–reflective oscillator model for ENSO (Fig. 4). In this conceptual model, they emphasize a positive feedback of zonal currents that advect the western Pacific warm pool toward the east. Three negative feedbacks tending to push the warm pool back to its original position of the western Pacific are: anomalous zonal current associated with wave reflection at the western boundary, anomalous zonal current associated with wave reflection at the eastern boundary, and mean zonal current converging at the eastern edge of the warm pool. During the warm phase of ENSO, equatorial westerly wind anomalies in the central Pacific produce upwelling Rossby and downwelling Kelvin waves that propagate westward and eastward, respectively. The westward propagating upwelling Rossby waves reflect to upwelling Kelvin waves after they reach the western boundary, whereas the eastward propagating downwelling Kelvin waves reflect to downwelling Rossby waves at the eastern boundary. Since both the upwelling Kelvin and downwelling Rossby waves have westward zonal currents, they tend to push the warm pool back to its original position of the western Pacific. These negative feedbacks along with the negative feedback of the mean zonal current make the coupled ocean–atmosphere system to oscillate.

Unlike the delayed oscillator, the western Pacific oscillator, and the recharge–discharge oscillator, the advective–reflective oscillator does not have a set of simple and heuristic equations. Instead, using a linear wind–forced ocean numerical model that was restricted to the zonal current of the first baroclinic Kelvin and first meridional Rossby waves, Picaut et al.

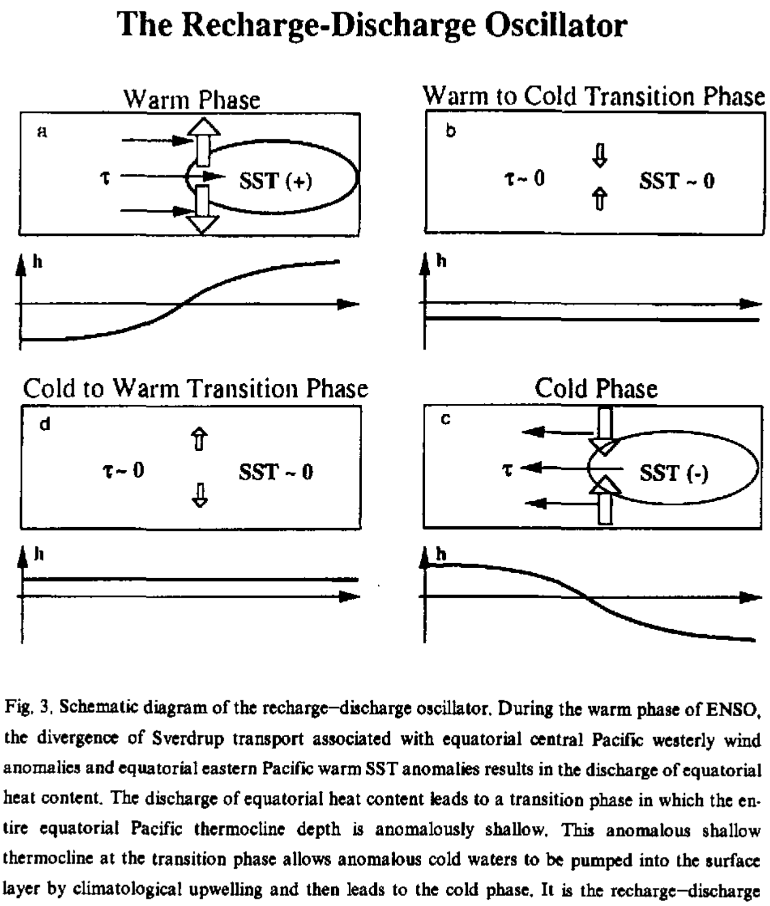

Fig. 3. Schematic diagram of the recharge–discharge oscillator. During the warm phase of ENSO, the divergence of Sverdrup transport associated with equatorial central Pacific westerly wind anomalies and equatorial eastern Pacific warm SST anomalies results in the discharge of equatorial heat content. The discharge of equatorial heat content leads to a transition phase in which the entire equatorial Pacific thermocline depth is anomalously shallow. This anomalous shallow thermocline at the transition phase allows anomalous cold waters to be pumped into the surface layer by climatological upwelling and then leads to the cold phase. It is the recharge–discharge process that makes the coupled ocean–atmosphere system oscillate on interannual time scales.

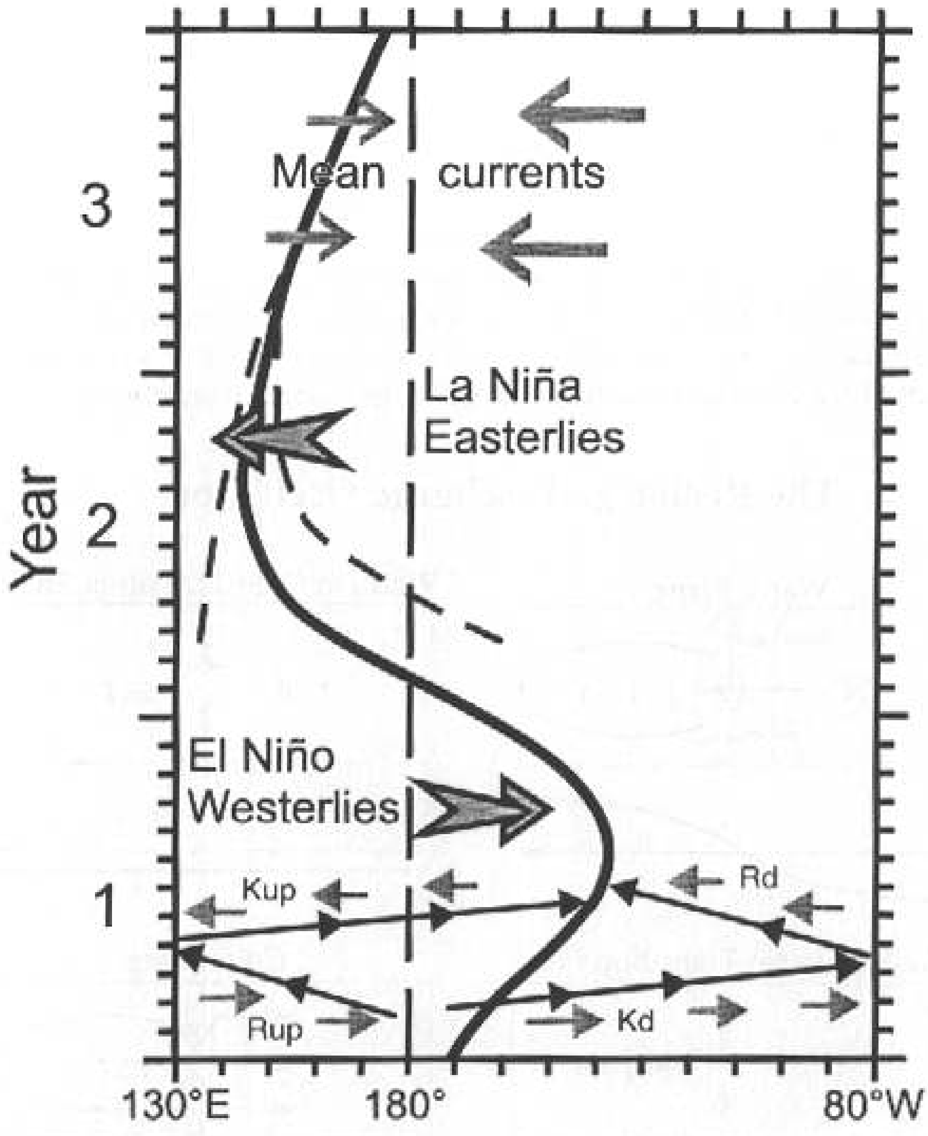

Fig. 4. Schematic diagram of the advective–reflective oscillator for ENSO. This oscillator emphasizes a positive feedback of zonal currents that advect the western Pacific warm pool toward the east. Three negative feedbacks tending to push the warm pool back to the western Pacific are: anomalous zonal current associated with wave reflection at the western boundary; anomalous zonal current associated with wave reflection at the eastern boundary; and mean zonal current converging at the eastern edge of the warm pool. These negative feedbacks make the coupled ocean–atmosphere system to oscillate.

(1997) showed an interannual oscillation with specified model parameters. They argued that anomalous zonal currents associated with equatorial wave reflection at both the western and eastern boundaries and mean zonal currents provide negative feedbacks for their model to oscillate.

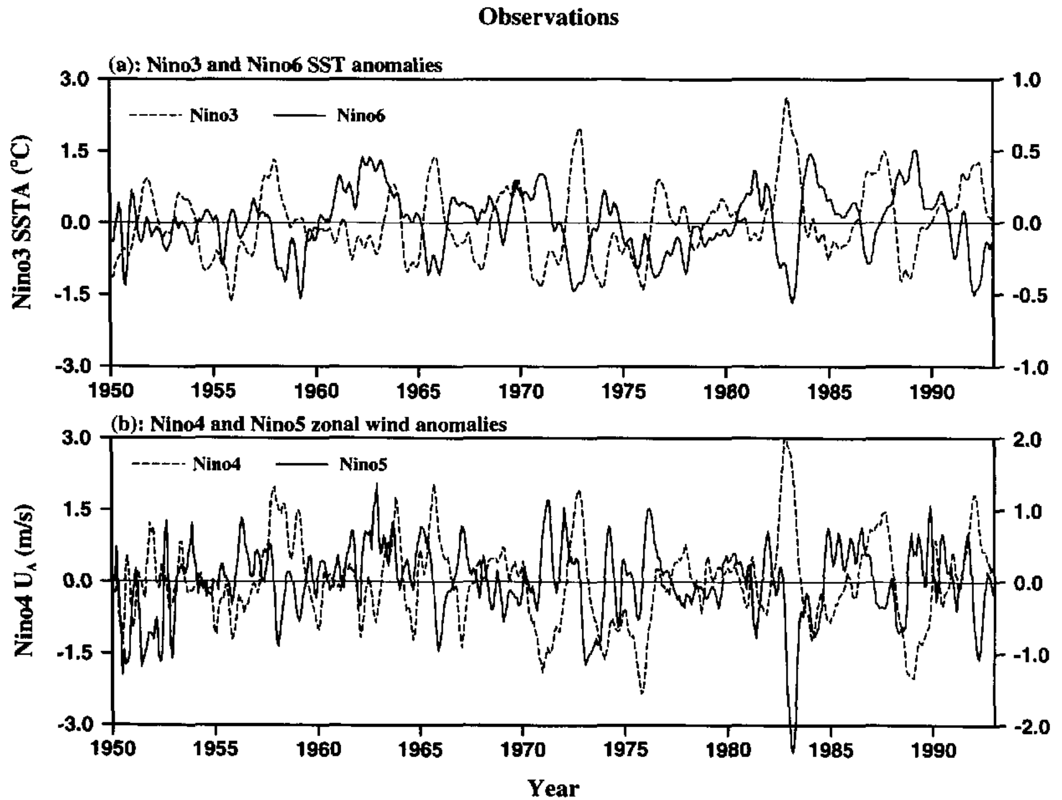

Fig. 5. Three-month running means of the observed (a) Nino3 (150°–90°W, 5°S–5°N) and Nino6 (140°–160°E, 8°–16°N) SST anomalies, and (b) Nino4 (160°E–150°W, 5°S–5°N) and Nino5 (120°–140°E, 5°S–5°N) zonal wind anomalies. The data are COADS data from January 1950 to December 1992.

3. A unified ENSO oscillator

As we stated in Section 2b, ENSO displays the western Pacific interannual anomaly patterns in nature, in addition to the eastern Pacific patterns (e.g., Wang et al., 1999b; Wang and Weisberg, 2000). These western Pacific interannual anomaly patterns are robust features of ENSO, independent of data sets. The western Pacific anomaly patterns also appear in other data sets and in other studies (e.g., Rasmusson and Carpenter, 1982; Rasmusson and Wallace, 1983; Graham and White, 1988, 1991; White et al., 1987, 1989; Kessler, 1990; Chao and Philander, 1993; Delcroix et al., 1994; Delecluse, a presentation in COARE98; Mestas–Nuñez and Enfield, 2001), but these patterns were not emphasized, probably due to the relatively small magnitude of SST anomalies when compared with those from the eastern Pacific patterns. However, the small SST anomalies in the off–equatorial western Pacific are sufficient to produce atmospheric responses of comparable amplitude to those in the equatorial eastern Pacific, due to the mean state of atmospheric convergence there associated with the western Pacific warm pool (Wang et al., 1999b; Wang, 2000).

For the purpose of comparison between eastern and western Pacific interannual variability, Wang et al. (1999b) defined two new regional ENSO indices in addition to the conventional ENSO indices, consistent with observations. The conventional eastern and central Pacific ENSO indices are: Nino1 over 90°–80°W, 10°–5°S; Nino2 over 90°–80°W, 5°S–0°; Nino3 over 150°–90°W, 5°S–5°N; and Nino4 over 160°E–150°W, 5°S–5°N. The new western Pacific ENSO indices are: Nino5 over 120°–140°E, 5°S–5°N and Nino6 over 140°–160°E, 8°–16°N. Comparisons of the SST anomalies between the Nino3 and Nino6 regions, and the zonal wind anomalies between the Nino4 and Nino5 regions are shown in Figs. 5a, and 5b, respectively. The SST anomalies in the Nino3 region are out–of–phase with those in the Nino6 region. That is, the warm (cold) SST anomalies in the equatorial eastern Pacific during El Niño (La Niña) are accompanied by the cold (warm) SST anomalies in the off–equatorial western Pacific. Similarly, zonal wind anomalies in the equatorial central Pacific tend to be out–of–phase with those in the equatorial western Pacific. During the mature warm (cold) phase of ENSO, equatorial westerly (easterly) wind anomalies in the central Pacific are accompanied by equatorial easterly (westerly) wind anomalies in the western Pacific.

The data show an interannual variability in both eastern and western Pacific anomaly patterns. The conceptual oscillator model should consider the variations of both the eastern and western anomaly patterns. Started from a coupled ocean–atmosphere system that is similar to the coupled model of Zebiak and Cane (1987), we can formulate and derive a unified oscillator model [The detailed formulation and derivation are given in Wang (2001)].

$$\frac{dT}{dt} = a\tau_1 - b_1\tau_1(t-\eta) + b_2\tau_2(t-\delta) - \varepsilon T^3, \tag{8}$$

$$\frac{dh}{dt} = -c\tau_1(t-\lambda) - R_h h, \tag{9}$$

$$\frac{d\tau_1}{dt} = dT - R_{\tau 1}\tau_1, \tag{10}$$

$$\frac{d\tau_2}{dt} = eh - R_{\tau 2}\tau_2. \tag{11}$$

Eqs. (8)–(11) control the variations of the Nino3 SST anomaly, T, the Nino6 thermocline anomaly, h, the Nino4 zonal wind stress anomaly, τ_1, and the Nino5 zonal wind stress anomaly, τ_2. The parameters a, b_1, b_2, c, d, and e are constant. The parameters η, δ, and λ represent the delay times. The parameters ε, R_h, $R_{\tau 1}$, and $R_{\tau 2}$ are damping coefficients.

The first term on the right hand side of Eq. (8) represents the positive feedback in the coupled system. The second term represents the negative feedback due to wave reflection at the western boundary. The third term represents the negative feedback due to the wind–forced wave contribution in the equatorial western Pacific. The last term is a cubic damping term that does not affect oscillatory behavior, but it limits anomaly growth (Battisti and Hirst, 1989; Wang, 2001). Eq. (9) states that the Nino6 thermocline anomaly is controlled by the wind stress in the Nino4 region, with a damping rate of R_h^{-1}. Eqs. (10) and (11) show that the Nino4 and Nino5 zonal wind stress anomalies are related to the Nino3 SST and Nino6 thermocline anomalies, respectively. As will be shown in next section, the unified oscillator model can be reduced to different oscillator models by further simplifications and assumptions. Since all of these different oscillators can oscillate on interannual time scales (Battisti and Hirst, 1989; Weisberg and Wang, 1997b; Jin, 1997a; Picaut et al., 1997), the model parameters are chosen to be consistent with those in these oscillator models: $a = 1.5 \times 10^{-2}$ °C m^2 N^{-1} yr^{-1}, $b_1 = 2.5 \times 10^2$ °C m^2 N^{-1} yr^{-1}, $b_2 = 7.5 \times 10^2$ °C m^2 N^{-1} yr^{-1}, $c = 1.5 \times 10^3$ m^3 N^{-1} yr^{-1}, $d = 3.6 \times 10^{-2}$ °C^{-1} N m^{-2} yr^{-1}, $e = 3.0 \times 10^{-3}$ N m^{-3} yr^{-1}, $\varepsilon = 1.2$°C^{-2} yr^{-1}, $R_h = 5.0$ yr^{-1}, $R_{\tau 1} = R_{\tau 2} = 2.0$ yr^{-1}, $\eta = 150$ days, $\delta = 30$ days, and $\lambda = 180$ days. These parameters give the unified oscillator model an interannual oscillation (Fig. 6). The model Nino3 SST anomalies and Nino6 thermocline depth anomalies are approximately in–phase with the model Nino4 zonal wind stress anomalies and Nino5 zonal wind stress anomalies, respectively. The model Nino3 SST anomalies tend to be out–of–phase with the model Nino6 thermocline depth anomalies. During the warm (cold) phase of the model ENSO, equatorial westerly (easterly) wind anomalies in the Nino4 region are accompanied by equatorial easterly (westerly) wind anomalies in the Nino5 region. All of these behaviors are consistent with observations.

4. Special cases of the unified oscillator

In this section, we will show that the ENSO conceptual models of the delayed oscillator, the western Pacific oscillator, and the recharge–discharge oscillator can be extracted as special cases of the unified oscillator model of Eqs. (8)–(11) by further simplifications and assumptions. Most of the physics of the advective–reflective oscillator are implicitly included in the unified oscillator, and the negative feedback of wave reflection at the eastern boundary can be added to the unified oscillator.

4.1 *The delayed oscillator*

The delayed oscillator model overlooks the ENSO western Pacific anomaly patterns and does not consider the effect of the western Pacific on ENSO. It is assumed that winds in the western Pacific do not affect the SST anomalies in the eastern Pacific. If we neglect the contribution of the wind–forced wave in the western Pacific, the unified oscillator of Eqs. (8)–(11) will exclude the role of the western Pacific in ENSO. By setting $b_2 = 0$ in Eq. (8), the western Pacific variables τ_2 and h are decoupled from the coupled system. The closed form of the coupled system requires only two equations

$$\frac{dT}{dt} = a\tau_1 - b_1\tau_1(t - \eta) - \varepsilon T^3, \tag{12}$$

$$\frac{d\tau_1}{dt} = dT - R_{\tau 1}\tau_1. \tag{13}$$

The physics of Eqs. (12) and (13) is the delayed oscillator, but the mathematical form is different from the original delayed oscillator. There are two equations and two variables in the coupled system of Eqs. (12) and (13). This system considers the variations of the Nino3 SST anomalies and the variations of the Nino4 zonal wind stress anomalies. The atmospheric zonal wind stress anomalies induce the variations of the SST anomalies that in turn

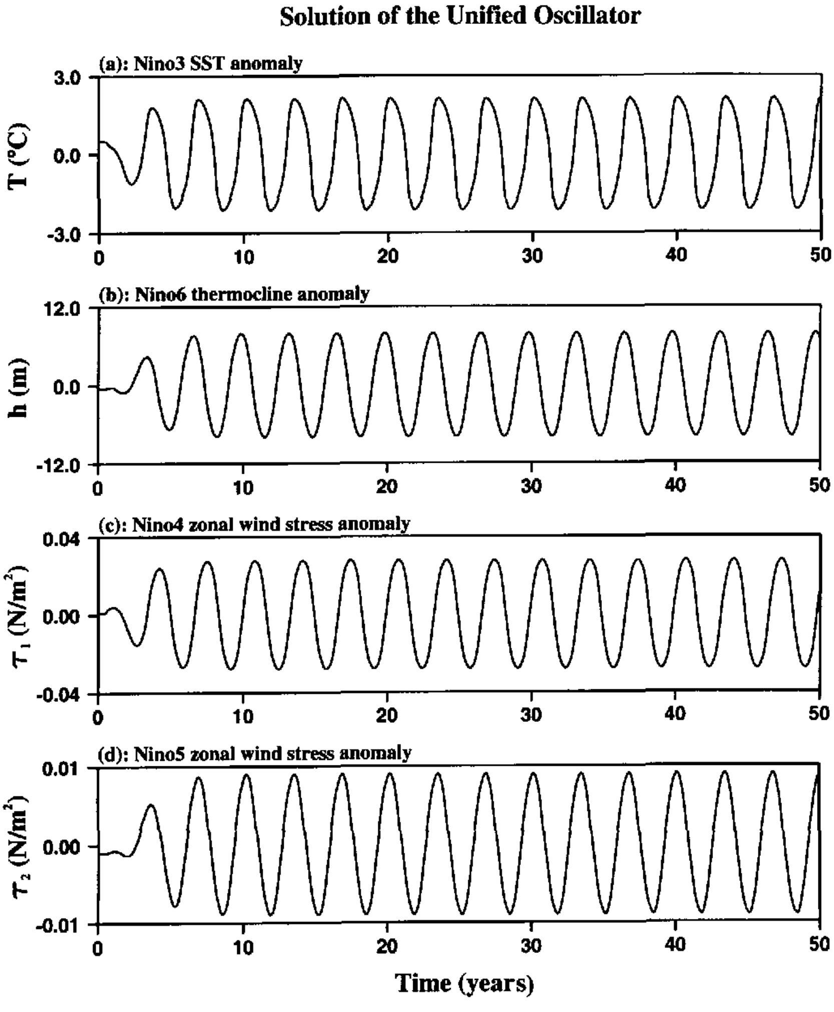

Fig. 6. Solution of the unified oscillator model of Eqs. (8)–(11).

affect the zonal wind stress anomalies. It is the interactions between the oceanic and atmospheric variables associated with equatorial wave dynamics that form the coupled system.

Further assumption or simplification can reduce Eqs. (12) and (13) into one equation of the original delayed oscillator. If we drop the time derivative of Eq. (13) (equivalent to assuming that the atmosphere is in a steady state), we obtain

$$\tau_1 = \frac{d}{R_{\tau 1}} T. \tag{14}$$

Substituting Eq. (14) into Eq. (12) results in

$$\frac{dT}{dt} = \frac{ad}{R_{\tau 1}} T - \frac{b_1 d}{R_{\tau 1}} T(t-\eta) - \varepsilon T^3. \tag{15}$$

Eq. (15) is the form of the delayed oscillator model of Eq. (1). This conceptual oscillator model emphasizes the ocean and atmosphere interactions in the equatorial eastern Pacific and considers anomaly variations only in this coupling region. There are a positive feedback and a negative feedback in the delayed oscillator model of Eq. (15). Both the positive feedback and the delayed negative feedback result from the equatorial eastern Pacific. Free Rossby waves generated there propagate westward and reflect from the western boundary as Kelvin waves, providing a negative feedback for the coupled ocean–atmosphere system to oscillate.

4.2 *The western Pacific oscillator*

The western Pacific oscillator of Weisberg and Wang (1997b) and Wang et al. (1999b) emphasizes the role of the western Pacific anomaly patterns in ENSO. Off–equatorial SST and SLP variations west of the date line initiate equatorial wind anomalies in the western Pacific. These wind anomalies force ocean responses that proceed eastward to affect anomalies in the equatorial eastern Pacific. This oscillator model does not necessarily require wave reflections at the western boundary. Neglecting the feedback due to equatorial Rossby wave reflection at the western boundary of the unified oscillator by setting $b_1 = 0$, Eqs. (8)–(11) reduce to the western Pacific oscillator

$$\frac{dT}{dt} = a\tau_1 + b_2\tau_2(t-\delta) - \varepsilon T^3, \tag{2}$$

$$\frac{dh}{dt} = -c\tau_1(t-\lambda) - R_h h, \tag{3}$$

$$\frac{d\tau_1}{dt} = dT - R_{\tau 1}\tau_1, \tag{4}$$

$$\frac{d\tau_2}{dt} = eh - R_{\tau 2}\tau_2. \tag{5}$$

During the warm phase of ENSO, with atmospheric convection extending eastward into the Nino4 region, westerly wind anomalies are maximum there, and which increase the Nino3 SST anomalies as represented by the first term in Eq. (2). During the mature phase of El Niño, equatorial easterly wind anomalies are produced in the western Pacific. Easterly wind anomalies force eastward propagating upwelling Kelvin waves to affect the Nino3 SST anomalies. The contribution of the wind–forced Kelvin waves is represented by the second term in Eq. (2). The corollary off–equatorial response to the same process that causes central Pacific equatorial westerly wind anomalies is off–equatorial Rossby waves induced by off–equatorial wind stress curl. Westward propagating Rossby waves raise the thermocline in the Nino6 region, as represented by Eq. (3). The coupled system does not necessarily involve off–equatorial Rossby wave reflection at the western boundary. This is consistent with the conclusions of Battisti (1989) who argued that reflection of off–equatorial Rossby waves (outside of 8° latitude) at the western boundary does not contribute to ENSO. Eq. (4) parameterizes the linear relationship between the Nino4 zonal wind stress anomalies and the Nino3 SST anomalies. Eq. (5) relates the initiation of the Nino5 easterly wind anomalies to the Nino6 thermocline depth variations.

4.3 *The recharge–discharge oscillator*

The recharge–discharge oscillator of Jin (1997a, b) has two equations that control the variations of two variables in the equatorial Pacific. The unified oscillator model of Eqs. (8)–(11) has four equations and four variables. To reduce the unified oscillator to the recharge–discharge oscillator, we first need to reduce the equations and variables by further assumptions and simplifications. If we drop the time derivative terms in Eqs. (10) and (11) (equivalent to assuming that the atmosphere is in a steady state), we obtain

$$\tau_1 = \frac{d}{R_{\tau 1}} T, \tag{16}$$

$$\tau_2 = \frac{e}{R_{\tau 2}} h. \tag{17}$$

Substituting Eqs. (16) and (17) into Eqs. (8) and (9), respectively, we can obtain

$$\frac{dT}{dt} = \frac{ad}{R_{\tau 1}} T - \frac{b_1 d}{R_{\tau 1}} T(t-\eta) + \frac{b_2 e}{R_{\tau 2}} h(t-\delta) - \varepsilon T^3, \tag{18}$$

$$\frac{dh}{dt} = -\frac{cd}{R_{\tau 1}} T(t-\lambda) - R_h h. \tag{19}$$

Eqs. (18) and (19) are the delayed version of the recharge–discharge oscillator. If all delay parameters are set to zero, i.e., $\eta = 0$, $\delta = 0$, and $\lambda = 0$ [Jin (1997a) argued that the slow ocean dynamical adjustment due to the recharge–discharge process does not necessarily need the explicit role of wave propagation], Eqs. (18) and (19) are reduced to

$$\frac{dT}{dt} = \frac{ad - b_1 d}{R_{\tau 1}} T + \frac{b_2 e}{R_{\tau 2}} h - \varepsilon T^3, \tag{20}$$

$$\frac{dh}{dt} = \frac{cd}{R_{\tau 1}} T - R_h h. \tag{21}$$

Eqs. (20) and (21) are the recharge–discharge oscillator model of Eqs. (6) and (7).

The mathematical form of Eqs. (20) and (21) is the same as the recharge–discharge oscillator of Jin (1997a) of Eqs. (6) and (7), but its interpretation may be different. In Eqs. (20) and (21), h represents western Pacific off–equatorial thermocline depth anomalies, whereas in Eqs. (6) and (7), Jin interpreted it as western Pacific equatorial thermocline depth anomalies. In the formulation of the recharge–discharge oscillator model, Jin (1997b) used a two–strip (one for equatorial strip and another for off–equatorial strip) approximation to the ocean dynamics. The equation controlling thermocline depth variations in the off–equatorial strip [his Eq. (2.5)] is the same equation that we used for deriving Eq. (9). The finite difference form of this equation also represents off–equatorial thermocline depth variations [see his Eq. (3.2b)]. He further assumed that off–equatorial thermocline depth anomalies are linearly related to equatorial thermocline depth anomalies by wave reflection at the western boundary [Note that, like the delayed oscillator, the effect of equatorial wave reflection at the western boundary has already been considered in the unified oscillator model as in the second term of Eq. (8)]. Therefore, the equation that controls thermocline depth variations in his recharge–discharge oscillator model becomes one associated with equatorial western Pacific thermocline depth anomalies rather than off–equatorial western Pacific thermocline depth anomalies. However, observations show that maximum interannual SST (e.g., Wang et al., 1999b), thermocline depth (White et al., 1987, 1989; Kessler, 1990; Delecluse, a presentation in COARE98), and sea level (Delcroix et al., 1994: Busalacchi, 1996, personal communication) variations are in the western Pacific off–equatorial region. Therefore, this raises a question how to interpret the mathematical form of the recharge–discharge oscillator. If we consider thermocline depth variations as being in the off–equatorial western Pacific region, the recharge–discharge oscillator can be exactly reduced from the unified oscillator.

4.4 *The advective–reflective oscillator*

In the advective–reflective oscillator, Picaut et al. (1997) emphasize a positive feedback of zonal currents that advect the western Pacific warm pool toward the east. Three negative feedbacks tending to push the warm pool back to its original position of the western Pacific are: anomalous zonal current associated with wave reflection at the western boundary, anomalous zonal current associated with wave reflection at the eastern boundary, and mean zonal current converging at the eastern edge of the warm pool. Picaut et al. (1997) argued that equatorial westerly wind anomalies in the central Pacific induce eastward anomalous zonal currents that advect the western Pacific warm pool eastward. The eastward warm pool displacement decreases the east–west SST gradient that further strengthens the equatorial westerly wind anomalies. This positive feedback leads El Niño to a growth phase. In terms of the unified oscillator model, the positive feedback of zonal advection is included in the first term of Eq. (8) (i.e., in $a\tau_1$). In the SST equation that was used to derive Eq. (8), zonal advection of the mean zonal SST gradient by the anomalous zonal current is $u\partial\overline{T}/\partial x$ (Wang 2001). If equatorial westerly wind anomalies produce eastward anomalous zonal current as argued by Picaut et al. (1997), then u is proportional to τ_1 (Battisti and Hirst (1989) made a similar assumpation in deriving the delayed oscillator). Zonal advection is proportional to τ_1. Thus, the positive feedback of zonal advection is already in the unified oscillator model of Eq. (8).

At the same time, equatorial westerly wind anomalies in the central Pacific produce upwelling Rossby waves and downwelling Kelvin waves that propagate westward and eastward, respectively. The westward propagating upwelling Rossby waves reflect to upwelling Kelvin waves after they reach the western boundary. Since the upwelling Kelvin waves have westward zonal currents, they tend to push the warm pool back to its original position of the western Pacific. Although the physics of this negative feedback is not exactly same as that of the delayed oscillator, it may be included in the term of $-b_1\tau_1(t-\eta)$ in Eq. (8).

The second negative feedback of the advective–reflective oscillator is wave reflection at the eastern boundary. The downwelling Kelvin waves, which are produced by the equatorial westerly wind anomalies in the central Pacific, propagate eastward and are reflected as downwelling Rossby waves at the eastern boundary. Associated with the downwelling Rossby waves are westward zonal currents that also tend to stop growth of El Niño. This negative feedback due to wave reflection at the eastern boundary is not considered in other conceptual oscillator models, but it can be added to the unified oscillator of Eq. (8)

$$\frac{dT}{dt} = a\tau_1 - b_1\tau_1(t-\eta) + b_2\tau_2(t-\delta) - b_3\tau_1(t-\mu) - \varepsilon T^3, \tag{22}$$

where $b_3\tau_1(t-\mu)$ represents the effect of wave reflection at the eastern boundary.

Picaut et al. (1997) also argued that the mean zonal current plays a negative feedback since it may help to push the warm pool in the central Pacific during the warm phase of ENSO back to the western Pacific. For their model to oscillate, they had to use the mean zonal current stronger than observed mean zonal current. This may be because other negative feedbacks are also operating in the coupled ocean–atmosphere system. For example, the negative feedback of easterly wind–forced ocean responses in the western Pacific cannot be ignored. In nature, the combination of different negative feedbacks makes the coupled system switch from a warm (cold) phase to a cold (warm) phase. Notice that zonal advection of the anomalous zonal SST gradient by the mean zonal current is also in the SST equation from which the unified oscillator is formulated and derived (Wang, 2001).

5. Summary and discussion

The delayed oscillator, the western Pacific oscillator, the recharge–discharge oscillator, and the advective–reflective oscillator have been proposed to interpret the oscillatory nature of ENSO. All of these oscillator models have a positive ocean–atmosphere feedback in the equatorial eastern and central Pacific hypothesized by Bjerknes (1969). Each, however, has different negative feedbacks that turn the warm (cold) phase into the cold (warm) phase. In the delayed oscillator, free Rossby waves generated in the equatorial eastern Pacific propagate westward and reflect from the western boundary as Kelvin waves. Since thermocline depth anomalies for the returning Kelvin waves have signs opposite to those in the equatorial eastern Pacific, these provide a negative feedback for the coupled ocean–atmosphere system to oscillate. In the western Pacific oscillator, equatorial easterly wind anomalies in the western Pacific, which are produced by western Pacific off–equatorial cold SST and high SLP anomalies, induce an ocean upwelling response that evolves eastward along the equator to provide a negative feedback. In the recharge–discharge oscillator, equatorial wind anomalies in the central Pacific induce the Sverdrup transport that recharges (or discharges) equatorial heat content. It is the recharge–discharge process that leaves an anomalously deep (or shallow) equatorial thermocline that serves as the phase transition for the coupled ocean–atmosphere system. The advective–reflective oscillator assumes that anomalous zonal currents associated with wave reflection at the ocean boundaries and mean zonal current tend to stop growth of El Niño.

The unified oscillator model includes all of the physics of the delayed oscillator, the western Pacific oscillator, the recharge–discharge oscillator, and the advective–reflective oscillator. Consistent with ENSO anomaly patterns observed in the tropical Pacific, the unified oscillator model considers SST anomalies in the Nino3 region, zonal wind stress anomalies in both the Nino4 and Nino5 regions, and thermocline depth anomalies in the Nino6 region. If the western Pacific wind–forced response is neglected, the Nino6 thermocline and Nino5 zonal wind stress anomalies are decoupled from the coupled system, and the unified oscillator reduces to the delayed oscillator. If wave reflection at the western boundary is neglected, the unified oscillator reduces to the western Pacific oscillator. The mathematical form of the recharge–discharge oscillator can also be derived from this unified oscillator. Most of the physics of the advective–reflective oscillator are implicitly included in the unified oscillator, and the negative feedback of wave reflection at the eastern boundary is added to the unified oscillator. As suggested by this unified oscillator, all of the previous different ENSO oscillator mechanisms may be operating in nature.

The delay times associated with negative feedbacks of wave propagation are not constant. The ocean–atmosphere coupling gives rise to slow modes and modifies the equatorial wave modes (Hirst, 1986; Neelin, 1991; Wang and Weisberg, 1994, 1996). As shown analytically by Wang and Weisberg, the modifications depend on frequency and strength of air–sea coupling. The primary modifications are in low frequency bands, with a decrease in phase speeds of Kelvin and Rossby waves. The slow mode and the phase speed decrease of equatorial waves are observed in previous studies (e.g., White and Tai, 1992; Chao and Philander, 1993). Thus, the ocean–atmosphere coupling changes negative feedbacks associated with wave propagation and reflection. In reality, different negative feedbacks may also interact one another.

The ENSO oscillator models produce periodic solutions, whereas ENSO variability in nature is known to be irregular. Introduction of stochastic atmospheric forcing of weather noise to an otherwise perfectly periodic oscillatory system can lead to irregular or chaotic oscillations (e.g., Graham and White 1988; Penland and Sardeshmukh 1995; Moore and Kleeman 1999).

Interactions between annual and interannual cycles (Jin et al., 1994; Tziperman et al., 1994; Chang et al., 1995; Wang et al., 1999a) can also produce irregular and chaotic oscillations. There may be many reasons for ENSO irregularity, but even for a simple system of linear equations the recognition that nature does not provide a constant parameter medium leads to irregularity. The temporal variations of parameters in nature may also determine the relative role of different ENSO mechanisms where the parameters may be more or less important for each evolving ENSO. For example, the western Pacific may have relatively more influence on strong El Niño events than weak El Niño events because the western Pacific shows strong equatorial wind anomalies for strong El Niño events. In terms of the unified oscillator, temporal variations of the parameters b_1 and b_2 may determine the relative importance between the delayed oscillator and the western Pacific oscillator.

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