

GEOSCIENCES

Special Topic: Advances in El Niño Research

A review of ENSO theories

Chunzai Wang

ABSTRACT

The El Niño and the Southern Oscillation (ENSO) occurrence can be usually explained by two views of (i) a self-sustained oscillatory mode and (ii) a stable mode interacting with high-frequency forcing such as westerly wind bursts and Madden-Julian Oscillation events. The positive ocean–atmosphere feedback in the tropical Pacific hypothesized by Bjerknes leads the ENSO event to a mature phase. After ENSO event matures, negative feedbacks are needed to cease the ENSO anomaly growth. Four negative feedbacks have been proposed: (i) reflected Kelvin waves at the ocean western boundary, (ii) a discharge process due to Sverdrup transport, (iii) western-Pacific wind-forced Kelvin waves and (iv) anomalous zonal advections and wave reflection at the ocean eastern boundary. These four ENSO mechanisms are respectively called the delayed oscillator, the recharge–discharge oscillator, the western-Pacific oscillator and the advective–reflective oscillator. The unified oscillator is developed by including all ENSO mechanisms, i.e. all four ENSO oscillators are special cases of the unified oscillator. The tropical Pacific Ocean and atmosphere interaction can also induce coupled slow westward- and eastward-propagating modes. An advantage of the coupled slow modes is that they can be used to explain the propagating property of interannual anomalies, whereas the oscillatory modes produce a standing oscillation. The research community has recently paid attention to different types of ENSO events by focusing on the central-Pacific El Niño. All of the ENSO mechanisms may work for the central-Pacific El Niño events, with an addition that the central-Pacific El Niño may be related to forcing or processes in the extra-tropical Pacific.

Keywords: ENSO, ocean–atmosphere interaction, climate variability

INTRODUCTION

El Niño represents oceanic warming in the tropical Pacific Ocean and the Southern Oscillation is a saw-saw of sea-level pressure (SLP) between the tropical western and eastern Pacific. Bjerknes [1] first recognized that El Niño and the Southern Oscillation (ENSO) are linked together and are oceanic and atmospheric aspects of the same interannual climate phenomenon. Bjerknes [1] hypothesized that a positive ocean–atmosphere feedback process causes ENSO. Given an initial warm sea-surface temperature (SST) anomaly in the equatorial eastern Pacific, the warm SST anomaly reduces the east–west SST gradient and hence weakens the Walker circulation [2,3], producing the westerly wind anomaly in the equatorial central Pacific. The westerly wind anomaly in turn drives the ocean circulation change that further enhances the SST anomaly. As a result

of the positive feedback, the tropical Pacific reaches a warm state, i.e. El Niño. After El Niño matures, negative feedbacks are required to turn El Niño from a warm phase to a cold phase, which was called La Niña [4].

Bjerknes's landmark paper in 1969 was not paid attention to until the 1980s. The occurrence of the intensive 1982–83 El Niño event was not realized by the community. This motivated the international research community to heavily study ENSO during the past three decades by focusing on interaction between the tropical Pacific Ocean and atmosphere [4–10]. ENSO is one of the natural climate phenomena whose understanding, observations and prediction are relatively successful and complete due to hard and excellent work by the international research community during the past three decades.

State Key Laboratory of Tropical Oceanography, South China Sea Institute of Oceanology, Chinese Academy of Sciences, Guangzhou 510301, China

E-mail:
cwang@scsio.ac.cn

Received 6 March 2018; Revised 5 August 2018; Accepted 17 September 2018

The present paper summarizes and reviews ENSO theories associated with physical processes. The ‘Self-sustained ENSO oscillators’ section summarizes the view of ENSO as self-sustained oscillators. The ‘A stable mode interacting with high-frequency forcing’ and ‘Eastward- and westward-propagating slow modes’ sections discuss ENSO as a stable mode interacting with high-frequency forcing and as a coupled slow (or SST) mode, respectively. The ‘Dynamics of the central-Pacific El Niño’ section briefly discusses the dynamics of the central-Pacific El Niño. Finally, summary and future work are given in the ‘Summary and future work’ section.

SELF-SUSTAINED ENSO OSCILLATORS

Bjerknes [1] first hypothesized that El Niño is a product of the tropical Pacific Ocean–atmosphere interaction—El Niño is caused by a positive ocean–atmosphere feedback process. The positive ocean–atmosphere feedback leads the tropical Pacific to a never-ending warm state. For the coupled ocean–atmosphere system to oscillate, negative feedbacks are required. The previous studies have proposed four negative feedbacks. These negative feedbacks can explain the oscillatory nature of ENSO and are called the delayed oscillator [11,12], the recharge–discharge oscillator [13,14], the western-Pacific oscillator [15,16] and the advective–reflective oscillator [17]. The physics of these four oscillator models are the negative feedbacks of (i) reflected Kelvin waves at the ocean western boundary, (ii) a discharge process due to Sverdrup transport, (iii) western-Pacific wind-forced Kelvin waves and (iv) anomalous zonal advection and reflected Rossby waves at the ocean eastern boundary, respectively. The unified oscillator suggested that all four ENSO mechanisms may operate in nature, and they are special cases of the unified oscillator [18].

Four ENSO oscillators

The delayed oscillator

The early oscillatory mechanism of ENSO, built on Rossby wave reflection at the ocean western boundary, was originally hypothesized by McCreary [19]. McCreary demonstrated that reflection of oceanic Rossby waves might help to generate the low-frequency interannual oscillations of ENSO by using shallow-water ocean dynamics. By emphasizing the delayed effects of oceanic wave reflection at the ocean western boundary, Suarez and Schopf [11] introduced and presented the delayed oscillator explaining the oscillatory feature of ENSO. Zebiak and Cane’s [20] coupled model—a coupled ocean–

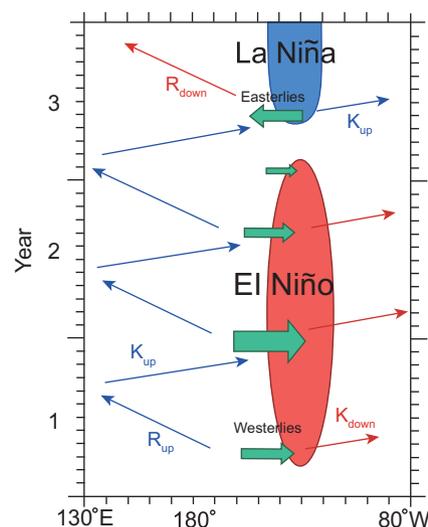


Figure 1. Schematic diagram of the delayed oscillator. Red shading represents positive SST anomalies and green arrows represent wind anomalies. K_{down} and K_{up} stand for downwelling and upwelling Kelvin waves that propagate eastward. R_{up} and R_{down} represent upwelling and downwelling Rossby waves that propagate westward.

atmosphere model of intermediate complexity with the specified mean states—was the first model to successfully predict ENSO. Battisti and Hirst [12] used this intermediate complexity model to develop the Suarez and Schopf’s delayed oscillator model:

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

where T represents the SST anomaly in the equatorial eastern Pacific and the model parameters A , B , η and ε are constant. The first term on the right-hand side (RHS) of Equation (1) stands for the Bjerknes positive feedback between the ocean and atmosphere. The second term on the RHS of Equation (1) represents the delayed negative feedback of wave reflection at the ocean western boundary. The warm SST anomalies in the equatorial eastern Pacific according to Gill’s [2] physics produce the equatorial westerly wind anomalies in the central Pacific that force westward-propagating Rossby waves. After the Rossby waves reach the ocean western boundary, they are reflected into Kelvin waves and the reflected Kelvin waves propagate eastward to the eastern Pacific, switching the sign of the anomalies in the eastern Pacific (Fig. 1). The cubic term in Equation (1) is a damping term that is to limit SST anomaly growth, but does not change the oscillation feature of Equation (1) [12,18].

The delayed oscillator overlooks the role of the ocean–atmosphere interaction in the western Pacific and also assumes that wave reflection at the ocean

eastern boundary is not important. The delayed oscillator only considers the physics of wave reflection at the ocean western boundary. With a broad range of model parameters, Equation (1) is able to oscillate on interannual time scales [21]. It was shown by Battisti and Hirst [12] that the delayed oscillator model of Equation (1) was able to account for major oscillatory features of Zebiak and Cane’s coupled model.

The recharge–discharge oscillator

Wyrtki [22,23] indicated that the growth and decrease in sea level over the western Pacific Ocean are related to ENSO. With these ideas and Zebiak and Cane’s coupled model, Jin [13,14] developed a recharge–discharge oscillator model for ENSO that is represented by:

$$\frac{dT}{dt} = CT + Dh - \epsilon T^3, \tag{2}$$

$$\frac{dh}{dt} = -ET - R_h h, \tag{3}$$

where T is the SST anomaly in the equatorial eastern Pacific and h is the thermocline-depth anomaly

in the equatorial western Pacific. The model parameters C, D, ϵ, E and R_h are constant.

As explained by Jin, Equations (2) and (3) represent the discharge and recharge of tropical Pacific Ocean heat content. The warm phase of ENSO is associated with the equatorial westerly wind anomalies in the central Pacific and the equatorial warm SST anomalies in the eastern Pacific, which result in and produce the divergence of Sverdrup transport and thus the discharge of tropical Pacific Ocean heat content (Fig. 2). The discharge of tropical Pacific Ocean heat content leads the tropical Pacific to a transition phase in which the entire thermocline depth is anomalously shallow due to the discharge of tropical ocean heat content. This anomalously shallow thermocline in the tropical Pacific permits anomalous cold waters to be pumped into the surface layer by climatological mean upwelling (green arrows in Fig. 2). The whole process then leads the warm phase of ENSO to the cold phase. The same process but with the opposite sign can lead the cold phase to the warm phase. Thus, the recharge–discharge process makes the coupled system oscillate on interannual time scales.

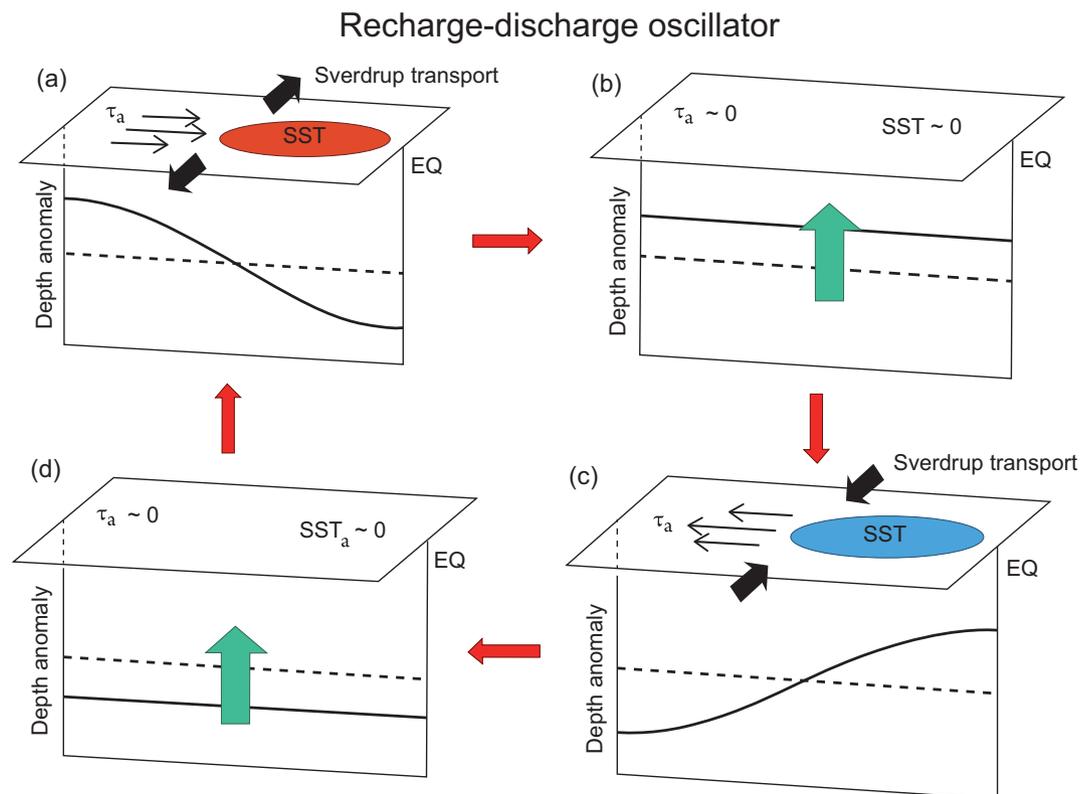


Figure 2. Schematic diagram of the recharge–discharge oscillator. Shown are (a) the warm phase, (b) the warm-to-cold transition phase, (c) the cold phase and (d) the cold-to-warm transition phase. Red (blue) SST represents warm (cold) SST anomalies and thin black arrows stand for wind anomalies. Dashed lines represent zero of the thermocline-depth anomalies and black lines are the thermocline-depth anomalies. Heavy black arrows mean the divergence and convergence of Sverdrup transport. Green arrows represent climatological mean upwelling.

The western-Pacific oscillator

When an ENSO event occurs, both the tropical eastern and western Pacific show interannual anomaly patterns [24,16,25]. For example, during the warm phase of ENSO, the warm SST and low SLP anomalies in the equatorial eastern Pacific and the low outgoing longwave radiation (OLR) anomalies in the equatorial central Pacific are accompanied by the cold SST and high SLP anomalies in the off-equatorial western Pacific and the high OLR anomalies in the off-equatorial far western Pacific. Additionally, the westerly wind anomalies over the equatorial central Pacific are associated with the easterly wind anomalies in the equatorial western Pacific. During the cold phase of ENSO, the relationship is also held, but with anomalies of opposite signs.

The delayed oscillator only considers the ENSO eastern-Pacific anomaly pattern and overlooks or does not consider the western-Pacific anomaly pattern. Consistently with and supported by observational and modeling results [15,16,25,26,27], Weisberg and Wang [15] developed and formulated a western-Pacific oscillator model for ENSO. This oscillator model stresses the role of the ocean-atmosphere coupling over the western Pacific. Particularly, the equatorial wind anomalies in the far western Pacific play an important role in the evolution of ENSO. The western-Pacific oscillator model is represented by the following equations:

$$\frac{dT}{dt} = a\tau_1 + b_2\tau_2(t - \delta) - \varepsilon T^3, \quad (4)$$

$$\frac{dh}{dt} = -c\tau_1(t - \lambda) - R_h h, \quad (5)$$

$$\frac{d\tau_1}{dt} = dT - R_{\tau_1}\tau_1, \quad (6)$$

$$\frac{d\tau_2}{dt} = eh - R_{\tau_2}\tau_2, \quad (7)$$

where T is the SST anomaly in the equatorial eastern Pacific, h is the thermocline-depth anomaly in the off-equatorial western Pacific, and τ_1 and τ_2 are the equatorial zonal wind-stress anomalies in the central Pacific and the western Pacific, respectively. All model parameters are constant.

According to Gill's [2] atmosphere, heating in the equatorial central Pacific [28,29] produces a pair of cyclones in the off-equatorial region, which results in the equatorial westerly wind anomalies (Fig. 3). The wind anomalies in the Nino4 region increase the Nino3 SST anomalies, and this process is represented by the first term on the RHS of Equation (4). At the same time, a pair of cyclones in the off-equatorial region raises the thermocline via Ekman pumping. As a result, a shallow thermocline anomaly in the off-equatorial region expands over the Nino6 region of the western Pacific. The process is represented by Equation (5) with a delay time of λ . The shallow off-equatorial thermocline results in the cold SST anomalies and the high SLP anomalies in the Nino6 region [16]. These off-equatorial high SLP anomalies initiate and induce the equatorial easterly wind anomalies in the Nino5 region.

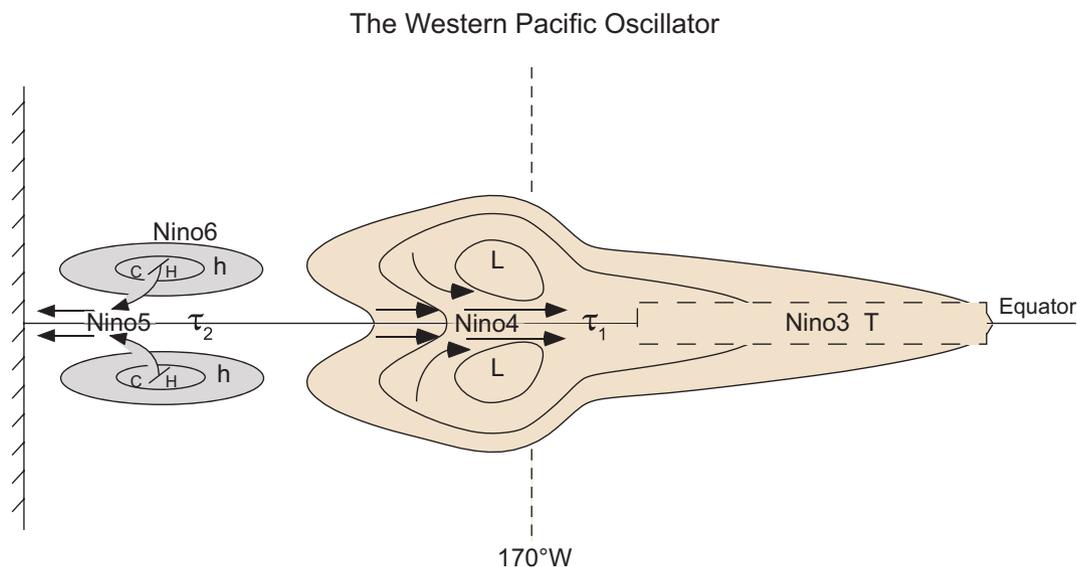


Figure 3. Schematic diagram of the western-Pacific oscillator. The ENSO index regions of Nino3, Nino4, Nino5 and Nino6 are defined in Wang *et al.* [16]. L, C and H represent low SLP, cold SST and high SLP, respectively. τ_1 and τ_2 stand for zonal wind-stress anomalies in the Nino4 and Nino5 regions, respectively. T is the SST anomaly in the Nino3 region and h is the thermocline-depth anomaly in the Nino6 region.

The equatorial easterly wind anomalies force an upwelling Kelvin wave that propagates eastward and serves as a negative feedback for the coupled system to oscillate on interannual time scales. This delayed negative feedback is represented by the second term on the RHS of Equation (4). Equation (6) relates the Nino4 zonal wind-stress anomalies to the Nino3 SST anomalies. Equation (7) states that the initiation of the Nino5 easterly wind anomalies is related to the Nino6 thermocline anomalies.

The western-Pacific oscillator emphasizes the oceanic processes that relate to the westerly wind anomalies in the central Pacific, the anomalous anticyclone in the off-equatorial western Pacific and the equatorial easterly wind anomalies in the western Pacific during ENSO. The thermocline in the western Pacific shows a large variation around 10° north of the equator, although the mean thermocline in the western Pacific is generally deep. Wang *et al.* [16] showed that the off-equatorial shallow thermocline anomalies could increase the SLP anomalies, which in turn initiate the equatorial easterly wind anomalies in the western Pacific. Subsequently, several studies emphasized the atmospheric processes and mechanisms for maintaining and developing the anomalous anticyclone and the equatorial easterly wind anomalies in the western Pacific: local wind–evaporation–SST feedback [30], seasonal modulations of atmospheric response to El Niño [31,32] and the Indian Ocean’s effect [33,34].

The advective–reflective oscillator

Picaut *et al.* [17] found an in-phase relationship between the eastern edge of the western-Pacific warm pool (WPWP) and the Southern Oscillation index, and Picaut and Delcroix [35] studied the wave reflections in both the ocean western and eastern boundaries. Based on the results in these two studies, Picaut *et al.* [36] proposed a conceptual model of the advective–reflective oscillator for ENSO (Fig. 4). In this oscillator, they argued that the positive feedback results from ocean zonal currents that advect the WPWP toward the east. The three negative feedbacks, all of which tend to push the WPWP back to the western Pacific, include: (i) anomalous zonal current associated with wave reflection at the ocean western boundary, (ii) anomalous zonal current associated with wave reflection at the ocean eastern boundary and (iii) mean zonal current converging at the WPWP’s eastern edge. When an El Niño event occurs, the equatorial westerly wind anomalies are located in the central Pacific and the westerly wind anomalies force westward-propagating upwelling Rossby and eastward-propagating downwelling Kelvin waves. The westward-propagating upwelling

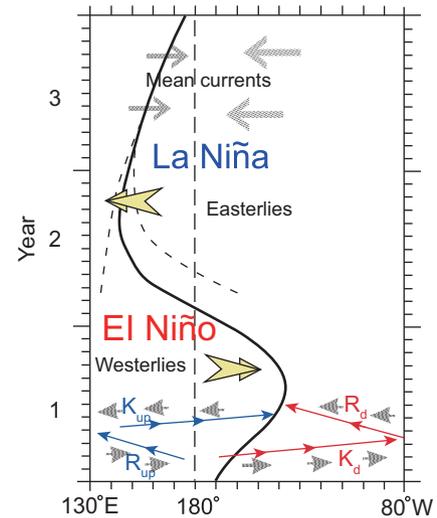


Figure 4. Schematic diagram of the advective–reflective oscillator. Yellow arrows represent zonal wind anomalies and gray arrows represent oceanic currents. The black thick solid line indicates the eastern edge of the western-Pacific warm pool. K_d and K_{up} stand for downwelling and upwelling Kelvin waves that propagate eastward. R_{up} and R_d represent upwelling and downwelling Rossby waves that propagate westward.

Rossby waves are reflected to upwelling Kelvin waves after they reach the ocean western boundary, whereas the eastward-propagating downwelling Kelvin waves are reflected to downwelling Rossby waves at the ocean eastern boundary. Because both the upwelling Kelvin and downwelling Rossby waves have westward zonal currents, their effects are to push the WPWP back to the western Pacific. These negative feedbacks along with the one associated with the mean zonal current make the coupled system to oscillate on interannual time scales.

Picaut *et al.* [36] did not give a set of heuristic equations for the advective–reflective oscillator, unlike the other three oscillator models. However, Picaut *et al.* demonstrated an interannual model oscillation by using a linear ocean numerical model forced by wind anomalies, which was associated with the zonal current of the first baroclinic Kelvin and first meridional Rossby waves. The physics for the model to oscillate is due to the anomalous zonal currents due to wave reflections at both the ocean western and eastern boundaries and the mean zonal currents. As will be shown later in the section on ‘Special Case: The Advective–Reflective Oscillator’, the advective–reflective oscillator can have a set of heuristic equations and is a special case of the unified oscillator.

The unified ENSO oscillator

As shown in the ‘Four ENSO oscillators’ section, four ENSO oscillators have been proposed and they are all capable of oscillating on interannual time scales. Given these four oscillators, Wang [18] developed the unified ENSO oscillator based on the dynamics and thermodynamics of Zebiak and Cane’s coupled ocean–atmosphere model. The motivation is to include the physics of all previous ENSO oscillator models, with a hypothesis that ENSO may be a multi-mechanism phenomenon and their relative importance may depend on time. Considering ENSO interannual anomalies in both the eastern and western Pacific, Wang [18] formulated and derived the unified oscillator:

$$\frac{dT}{dt} = a\tau_1 - b_1\tau_1(t - \eta) + b_2\tau_2(t - \delta) - b_3\tau_1(t - \mu) - \varepsilon T^3, \quad (8)$$

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

$$\frac{d\tau_1}{dt} = dT - R_{\tau_1}\tau_1, \quad (10)$$

$$\frac{d\tau_2}{dt} = eh - R_{\tau_2}\tau_2, \quad (11)$$

where T , h , τ_1 and τ_2 are four variables that represent the SST anomalies in the equatorial eastern Pacific, the thermocline-depth anomalies in the off-equatorial western Pacific, the zonal wind-stress

anomalies in the equatorial central Pacific and the zonal wind-stress anomalies in the equatorial western Pacific, respectively. All of the model parameters are constants. For a given set of parameters, Equations (8)–(11) can oscillate on interannual time scales.

The first term on the RHS of Equation (8) represents Bjerknes’s positive feedback. The second and third terms represent the negative feedbacks as a result of ocean western boundary wave reflection and the western Pacific wind-forced wave effect, respectively (Fig. 5). The fourth term is the wave reflection contribution at the ocean eastern boundary. Equation (9) shows that the off-equatorial thermocline anomalies in the western Pacific are related to the zonal wind-stress anomalies in the equatorial central Pacific. Equation (10) states that the zonal wind-stress anomalies in the equatorial central Pacific are controlled by the SST anomalies in the eastern Pacific and Equation (11) indicates that the zonal wind-stress anomalies in the equatorial western Pacific are controlled by the off-equatorial thermocline anomalies in the western Pacific. As shown next, the unified oscillator of Equations (8)–(11) can reduce to the four previous ENSO oscillators by making further simplifications and assumptions. In other words, the four ENSO oscillators—the delayed oscillator, the recharge–discharge oscillator, the western-Pacific oscillator and the advective–reflective oscillator—are special cases of the unified oscillator.

Special case: the delayed oscillator

The delayed oscillator assumes that ocean–atmosphere interaction in the western Pacific and wave reflection at the ocean eastern boundary

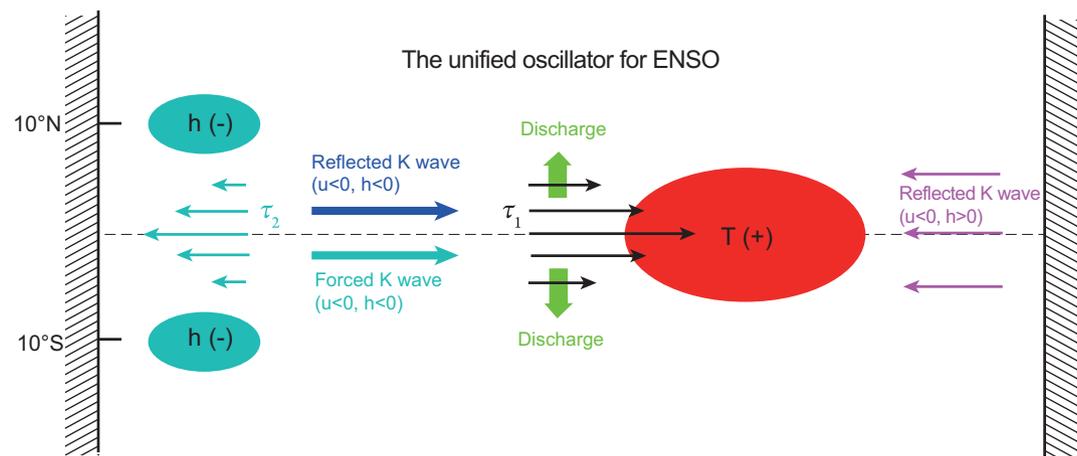


Figure 5. Schematic diagram of the unified oscillator. The unified oscillator includes all negative feedbacks of the previous four ENSO oscillators, which include the reflected Kelvin wave at the ocean western boundary, the discharge process due to Sverdrup transport, the western-Pacific wind-forced Kelvin wave and the reflected Rossby wave at the ocean eastern boundary.

are not important. By setting $b_2 = 0$ and $b_3 = 0$ in Equation (8), two variables of τ_2 and h in the western Pacific are decoupled from the coupled system. If the time derivative in Equation (10) is further dropped, the unified oscillator reduces to:

$$\frac{dT}{dt} = \frac{ad}{R_{\tau_1}} T - \frac{b_1 d}{R_{\tau_1}} T(t - \eta) - \varepsilon T^3. \quad (12)$$

Equation (12) is the delayed oscillator of Equation (1) with $A \equiv ad/R_{\tau_1}$ and $B \equiv b_1 d/R_{\tau_1}$.

Special case: the recharge–discharge oscillator

The recharge–discharge oscillator considers two variables: the SST anomalies in the equatorial eastern Pacific and the thermocline anomalies in the equatorial western Pacific. Jin [13] argued that equatorial-wave propagations do not explicitly appear in the recharge–discharge oscillator, although the tropical Pacific Ocean is adjusted by equatorial-wave dynamics. If we drop the time derivatives in Equations (10) and (11), set all delayed parameters to zero (i.e. $\eta = 0$, $\delta = 0$ and $\lambda = 0$) and ignore wave reflection at the ocean eastern boundary ($b_3 = 0$), the unified oscillator reduces to:

$$\frac{dT}{dt} = \frac{ad - b_1 d}{R_{\tau_1}} T + \frac{b_2 e}{R_{\tau_2}} h - \varepsilon T^3, \quad (13)$$

$$\frac{dh}{dt} = -\frac{cd}{R_{\tau_1}} T - R_h h. \quad (14)$$

The mathematical form of Equations (13) and (14) is the same as the recharge–discharge oscillator of Equations (2) and (3), with $C \equiv (ad - b_1 d)/R_{\tau_1}$, $D \equiv b_2 e/R_{\tau_2}$, and $E \equiv cd/R_{\tau_1}$. However, in Jin’s recharge–discharge oscillator, h is the thermocline anomaly in the equatorial western Pacific (instead of the off-equatorial western Pacific here).

Special case: the western-Pacific oscillator

The western-Pacific oscillator emphasizes the importance of western-Pacific Ocean–atmosphere interaction in ENSO. For the western-Pacific oscillator model to oscillate, wave reflections at the ocean western and eastern boundaries are not necessarily required. If we neglect the negative feedbacks from wave reflections by setting $b_1 = 0$ and $b_3 = 0$, Equations (8)–(11) reduce to:

$$\frac{dT}{dt} = a\tau_1 + b_2\tau_2(t - \delta) - \varepsilon T^3, \quad (15)$$

$$\frac{dh}{dt} = -c\tau_1(t - \lambda) - R_h h, \quad (16)$$

$$\frac{d\tau_1}{dt} = dT - R_{\tau_1}\tau_1, \quad (17)$$

$$\frac{d\tau_2}{dt} = eh - R_{\tau_2}\tau_2. \quad (18)$$

Equations (15)–(18) are the western-Pacific oscillator of Equations (4)–(7).

Special case: the advective–reflective oscillator

When Picaut *et al.* [36] proposed the advective–reflective oscillator for ENSO, they did not provide a set of heuristic equations for the advective–reflective oscillator. However, if we set $b_2 = 0$ in Equation (8), the unified oscillator model is reduced to:

$$\frac{dT}{dt} = a\tau_1 - b_1\tau_1(t - \eta) - b_3\tau_1(t - \mu) - \varepsilon T^3, \quad (19)$$

$$\frac{d\tau_1}{dt} = dT - R_{\tau_1}\tau_1. \quad (20)$$

As shown by Wang [18] and Battisti and Hirst [12], two advection terms $u\partial\bar{T}/\partial x$ and $\bar{u}\partial T/\partial x$ appear in the first term of $a\tau_1$ in Equation (19). Therefore, the contributions of zonal current are in the first term on the RHS of Equation (19), i.e. in $a\tau_1$. The negative feedback of the anomalous zonal current corresponding to wave reflection at the ocean western boundary is in the term $-b_1\tau_1(t - \eta)$ in Equation (19). The contribution of wave reflection at the ocean eastern boundary is presented by the third term on the RHS of Equation (19).

Non-linear oscillatory models

All ENSO oscillator models in the ‘Four ENSO oscillators’ and ‘The unified ENSO oscillator’ sections above are linear, producing periodic oscillations. However, when a noise or high-frequency forcing is added to these linear models, irregular oscillations appear. On the other hand, non-linear oscillatory models for ENSO were also built and developed [37–43]. One of advantages for these non-linear simple models is that models themselves can produce irregular oscillations without the need for a noise or high-frequency forcing. Additionally, these non-linear models also address some important features associated with ENSO. For example, Guckenhimer *et al.* [42] investigated the predictability of strong El Niño events that seem to occur on decadal time scales, such as the events in 1982–83,

1997–98 and 2015–16. Using a non-linear simple model, they argued that ENSO can be in a regime of irregular switching between an oscillatory state that has strong El Niño events and a chaotic state that lacks strong events, and that, in this regime, the timing of strong El Niño events on decadal time scales is unpredictable. Liang *et al.* [40] employed an analytical but non-linear model to study the ENSO asymmetry—the strongest El Niño is stronger than the strongest La Niña. Their model showed that the ENSO asymmetry occurs when the two adjacent strongest warm events are spaced farther apart and more small events occur in between, consistently with the recent result of Guckenheimer *et al.* [42]. Kohayama and Hartmann [43] used a non-linear recharge oscillator model to suggest that the non-linear ENSO warming suppression causes extreme El Niños to dissipate but La Niñas to remain almost unchanged, causing a La Niña-like mean state warming in the tropical Pacific, which is consistent with observations and GFDLES2M model.

A STABLE MODE INTERACTING WITH HIGH-FREQUENCY FORCING

ENSO is also viewed as a stable mode interacting with high-frequency forcing or triggering by stochastic atmospheric/oceanic forcing [44–48]. This view and hypothesis suppose that high-frequency forcing or random forcing external to the ENSO system is a cause of ENSO. The beautiful and attractive part of this view is that it provides a natural explanation in terms of noise for the irregular feature of ENSO. Since this view of ENSO requires the presence of ‘noise’, it can easily explain why no El Niño is exactly same and El Niño events are so hard to predict [47,49]. The external atmospheric variability or forcing may include the Madden-Julian Oscillation and westerly wind bursts [50–56], and the oceanic noise may involve oceanic high-frequency variability such as the tropical instability waves [57].

El Niño can be considered to be either a self-sustained oscillator mode or a stable mode interacting with high-frequency forcing. In either case, when an El Niño event occurs, the tropical central and eastern Pacific Ocean is warmer than the normal. However, after an El Niño event peaks in the winter, negative feedbacks are needed to terminate the further growth of the interannual anomalies. In other words, the negative feedbacks associated with the delayed oscillator, the recharge–discharge oscillator, the western-Pacific oscillator and the advective–reflective oscillator may be still valid for demise of an El Niño, even if El Niño is regarded as a stable mode interacting with high-frequency forcing. Mantua and

Battisti [58] showed that independent warm events still need the negative feedback due to wave reflection at the ocean western boundary. In summary, the physics of the previous ENSO oscillators are fundamental elements of ENSO, regardless of the views of ENSO theories.

EASTWARD- AND WESTWARD-PROPAGATING SLOW MODES

Tropical Pacific Ocean and atmosphere interaction is able to produce coupled slow modes, which can be a slow westward-propagating unstable mode [59,60] or a slow eastward-propagating unstable mode [60–67]. These coupled slow modes were further investigated numerically by Hirst [68] and analytically by Wang and Weisberg [69], demonstrating that they are related to ENSO anomaly propagations on interannual time scales. The delayed oscillator physics are not relevant to these coupled ocean–atmosphere slow modes. For example, Wang and Weisberg [70] performed two groups of model experiments with the closed and open ocean western boundary conditions. Their model results demonstrated that the evolutions of the eastward-propagating slow modes in these two experiments are nearly identical, indicating that wave reflection at the ocean western boundary does not play an important role in the slow modes because the open ocean western boundary does not allow waves to be reflected.

Neelin [71] introduced a concept of slow SST modes that are similar to the coupled slow modes due to air–sea coupling. They argued that the SST modes are largely distinct from the ocean dynamics modes that are relevant to the delayed oscillator. The existence of these modes in the coupled ocean–atmosphere system depends on the ocean adjustment processes. Two key adjustments on interannual time scales are the dynamical adjustment of the equatorial ocean and the thermodynamical change in SST due to air–sea coupling. The relative time length of these two adjustments determines which modes appear in the coupled system. If the thermodynamical change in SST is slow, the coupled system favors the SST mode instead of the ocean-wave dynamics. However, if the dynamical adjustment of the ocean is slow, the equatorial-wave dynamics plays an important role in the interannual oscillation in the coupled ocean–atmosphere system. Jin and Neelin [72] and Neelin and Jin [73] further argued that, in most of the model parameter space, the coupled modes are the mixed SST–ocean dynamics modes. An advantage of the slow modes is that they can explain the propagating property

of interannual anomalies, whereas the delayed oscillator mode produces a standing oscillation.

The direction of the mode's eastward or westward propagation is determined by physical processes. We first examine a case with a positive SST anomaly in the equatorial eastern and central Pacific. The atmospheric wind responses are westerly and easterly wind anomalies in the west and east of the positive SST anomaly, respectively [2]. This wind pattern produces anomalous zonal currents that advect warm (cold) water to the west (east) of the region. At the same time, the westerly (easterly) wind anomalies to the west (east) cause anomalous downwelling (upwelling). Because both zonal advection and downwelling increase the SST in its western side, the SST anomaly then propagates westward. On the other hand, the westerly wind anomalies deepen the thermocline in the east, which increases the SST in the east by mean upwelling. Furthermore, the non-linear term of the anomalous vertical temperature gradient by the anomalous upwelling can also increase the SST in the east [74]. Thus, both these processes cause the SST anomaly to propagate eastward.

Next, we consider the interannual anomaly propagation in the western Pacific Ocean [47,75]. We first discuss dynamical processes. The westerly wind bursts or the Madden-Julian Oscillation events in the western Pacific generate eastward current, which advects the warm water eastward, thus reducing the east–west temperature gradient. The reduction of the east–west temperature gradient weakens the easterly trade winds, which further increases the SST in the east, resulting in an eastward propagation. Second, we consider thermodynamical processes. The mean zonal wind pattern in the western Pacific during the boreal winter and spring is a weak westerly from 130°E to 150°E and an easterly from 150°E to the Date Line. An equatorial westerly anomaly superimposing the above mean zonal wind will cause different SST changes. Because the region of 160°E to 170°E is covered by the mean easterly wind, the westerly wind anomalies reduce the total wind speed and decrease evaporation, thus causing an increase in SST. However, due to the mean westerly wind west of 150°E, the westerly wind anomalies enhance the total wind speed, resulting in a decrease in SST through increased evaporation. As a result, an eastward SST zonal gradient is induced, which in turn reinforces the equatorial westerly wind anomalies in the western Pacific [3]. The interaction between the eastward SST gradient and westerly anomalies causes an eastward propagation of interannual anomalies when an El Niño event occurs in the tropical Pacific.

DYNAMICS OF THE CENTRAL-PACIFIC EL NIÑO

Two different types of ENSO events have recently been paid attention to, although they were not new phenomena in the tropical Pacific and were mentioned in the literature a long time ago [76,25,77–82]. The two types of ENSO events are defined based on locations of their maximum SST anomalies: the eastern-Pacific (EP) type over the eastern tropical Pacific and the central-Pacific (CP) type near the International Date Line [79,81]. The CP El Niño is also referred to as Date Line El Niño [78], El Niño Modoki [80] or Warm Pool El Niño [82]. Does the CP El Niño have different mechanisms from the EP El Niño?

In spite of the maximum anomaly location difference, the CP and EP El Niño all have a common pattern that the equatorial westerly (easterly) anomalies are always located to the west (east) of the positive SST anomalies. Therefore, all of the ENSO mechanisms discussed previously may work on the CP El Niño. Ashok *et al.* [80] stated that the wind-anomaly pattern induces the thermocline changes that are responsible for generating the CP ENSO event. The equatorial westerly and easterly wind anomalies force downwelling Kelvin and Rossby waves that propagate eastward and westward, respectively. These propagating waves work together to deepen the thermocline in the central Pacific, thus inducing the CP El Niño. Kug *et al.* [82] showed that the equatorial easterly anomalies in the eastern Pacific increase upwelling and surface evaporation and then decrease the warming of the CP El Niño event. Because the thermocline depth in the central Pacific is relatively deep, the wind-induced thermocline variations may not be efficient in inducing the CP SST anomalies. Instead, they suggested that ocean advections are responsible for the development of the CP El Niño event.

Yu *et al.* [83] performed a mixed-layer heat-budget analysis and also concluded that ocean-advection processes are responsible for the rapid increase in the SST anomalies during the CP ENSO event. However, they suggested that the origin of the warm SST anomalies is not local, but from atmospheric forcing in the extratropics and subsequent air–sea interactive processes in the subtropics. They indicated that the warm SST anomalies first appear in the northeastern subtropical Pacific and then expand toward the equatorial central Pacific. The air–sea interactive processes in the subtropics, which make the warm SST anomalies spread equatorward, are similar to those depicted by the seasonal footprinting mechanism [84–86]. The footprinting mechanism provides an explanation for how

mid-latitude atmospheric variations in the winter can produce subtropical SST anomalies, maintain them from the winter to the next summer and at the same time spread them toward the equatorial Pacific. In fact, many studies showed that the CP type of El Niño event is attributed to the North Pacific meridional mode (NPMM) variability [87–89]. The NPMM is generated in the subtropical northeastern Pacific through the wind–evaporation–SST feedback of Xie [90]. The associated NPMM variability expands toward the equatorial central Pacific and then the Bjerknes feedback comes in for the further development of the CP El Niño event.

By emphasizing the role of the zonal advective feedback in the CP El Niño event, Fang and Mu [91] recently extended the work of the recharge–discharge oscillator. They developed a three-region conceptual model including the central Pacific in addition to the two-region (the eastern and western Pacific) model of the recharge–discharge oscillator. The three-region oscillatory model is able to depict the different variations between the CP and EP El Niño events. Fang and Zheng [92] further showed that the main characteristics of the CP type of ENSO can be reproduced if the thermocline feedback is switched off and the zonal advective feedback is retained as the only positive feedback, confirming the dominant role played by zonal advective feedback in the development of the CP type of ENSO event.

The study by Chen *et al.* [93] suggested that El Niño diversity largely depends on westerly wind bursts in the equatorial western and central Pacific. They showed that the CP type of El Niño will occur if westerly wind bursts are relatively weak and are confined to the equatorial western Pacific. However, if westerly wind bursts are strong and move across the Date Line, the extreme EP type of El Niño will appear in the eastern Pacific. Their heat-budget analysis also revealed that, for the extreme EP El Niño, the vertical advection, i.e. thermocline feedback, is the dominant factor; whereas the horizontal advections are important for the CP El Niño. This study also concluded that ENSO is likely to be a combination of the self-sustaining oscillatory dynamics (as described by various oscillators in the ‘Self-sustained ENSO oscillators’ section) and the forcing of westerly wind bursts. Recently, Levine *et al.* [94] emphasized that multiplicative (i.e. state-dependent) westerly wind burst forcing can cause extreme El Niño events to occur in the eastern Pacific.

SUMMARY AND FUTURE WORK

The ENSO occurrence can usually be explained by two views: (i) a self-sustained oscillatory mode

and (ii) a stable mode interacting with high-frequency forcing such as westerly wind bursts and Madden-Julian Oscillation events. The positive ocean–atmosphere feedback hypothesized by Bjerknes leads ENSO event to a mature phase. After an ENSO event matures, negative feedbacks are needed to cease ENSO anomaly growth in the tropical Pacific. The previous studies have proposed four negative feedbacks: (i) reflected Kelvin waves at the ocean western boundary, (ii) a discharge process due to Sverdrup transport, (iii) western-Pacific wind-forced Kelvin waves and (iv) anomalous zonal advections and wave reflection at the ocean eastern boundary. These four ENSO mechanisms are respectively called the delayed oscillator, the recharge–discharge oscillator, the western-Pacific oscillator and the advective–reflective oscillator. Given the existence of four ENSO oscillators, the unified oscillator is developed by including all ENSO mechanisms, i.e. all four ENSO oscillators are special cases of the unified oscillator. The unified oscillator suggested that all physics associated with the four oscillators may work in nature, and their relative importance is time-dependent. The issue of ENSO as a self-sustained oscillatory mode or a stable mode interacting with high-frequency forcing is not settled yet. ENSO may be a self-sustained oscillatory mode during some periods, a stable mode during others or a mode that is intermediate or mixed between the former and the latter. The predictability of ENSO is more limited if ENSO is a stable mode interacting with high-frequency forcing than if ENSO is a self-sustained oscillatory mode because the former depends on random disturbances.

On interannual time scales, oceanic and atmospheric variables are also observed to propagate eastward or westward. The propagating property of interannual anomalies can be explained by the coupled slow or SST modes due to the ocean–atmosphere coupling. A number of dynamical and thermodynamical processes compete to determine the direction of modes’ eastward or westward propagation. An advantage of the coupled slow modes is that they can explain the propagating property of interannual ENSO anomalies, whereas the oscillator modes produce a standing oscillation.

ENSO studies have been focused on two types of ENSO events: the EP and CP ENSO events. These two types of ENSO events have different locations of maximum SST anomalies and heating anomalies, resulting in different climate- and weather-related impacts on the globe. In spite of the location difference of maximum SST anomalies, both the EP and CP El Niño events show the westerly wind anomalies in the west of maximum SST anomalies. Thus, it is not surprising that the physical processes (e.g. wave

reflections at the ocean western and eastern boundaries and ocean zonal advections) associated with the different oscillator mechanisms involve the CP El Niño events. Unlike the EP El Niño, the CP El Niño may more involve the ocean zonal advection feedback than the thermocline feedback. In addition to the processes in the tropical ocean–atmosphere system, the CP El Niño events may originate processes in the extra-tropical Pacific. In other words, the CP El Niño events may be initiated and related to forcing or processes in the extra-tropical Pacific.

Compared with other climate variability, ENSO is the most successful climate phenomenon in terms of its complete understanding due to the ocean–atmosphere interaction. However, there are at least several issues regarding the ENSO understanding that require the research community to focus on in the future. First, inter-ocean interactions can modify or change Pacific ENSO events, and thus interactions among the Pacific, Indian and Atlantic Oceans should be paid attention to in future research. Several studies have recently demonstrated the influences of the Indian Ocean and Atlantic Ocean on Pacific variability. For example, tropical Indian Ocean warming can prohibit the development of Pacific El Niño [95] and can strengthen the tropical Pacific trade winds observed during the last two decades [96]. The Atlantic multidecadal oscillation can increase the occurrence of the CP type of ENSO events [97] and can affect multidecadal ENSO variability [98]. The tropical Atlantic warming induces a cold tropical eastern Pacific and a warm Indo-Pacific on longer time scales [99]. Second, interactions among intra-seasonal variability (e.g. westerly wind bursts and Madden-Julian Oscillation events), interannual ENSO variability and decadal–multidecadal variability need to be further studied [100,101]. Third, the relationship between ENSO and global warming is uncertain. We are not even sure whether greenhouse warming will produce an El Niño-like or La Niña-like pattern in the tropical Pacific due to climate model uncertainty. Zheng *et al.* [102] concluded that the spatial pattern of tropical Pacific surface warming is the major source of inter-model uncertainty in ENSO amplitude change. On the other hand, Cai *et al.* [103] claimed that the extreme El Niño occurrence in the future in response to greenhouse warming would double. To advance and improve our understanding of anthropogenic and natural ENSO variability, global coupled climate models must be greatly improved and be able to simulate both ENSO and the response to greenhouse warming. Fourth, it is unknown whether the ENSO mechanisms discussed in this paper still hold in future global-warming scenarios. It is hoped that these

issues and problems will be understood, or at least partially understood, in the near future.

ACKNOWLEDGEMENTS

I would like to thank Mu Mu, Renhe Zhang, Chongyin Li, Huijun Wang, Xiu-Qun Yang, Dongxiao Wang, Xin Wang and Huaxia Liao for their comments and/or for helping get some references. Four reviewers' comments and suggestions helped to improve the manuscript.

FUNDING

This research is supported by the National Natural Science Foundation of China (41731173), the Pioneer Hundred Talents Program of the Chinese Academy of Sciences, the Leading Talents of Guangdong Province Program, the Strategic Priority Research Program of the Chinese Academy of Sciences (XDA20060502) and the National Program on Global Change and Air-Sea Interaction (GASI-IPOVAI-04).

REFERENCES

1. Bjerknes J. Atmospheric teleconnections from the equatorial Pacific. *Mon Wea Rev* 1969; **97**: 163–72.
2. Gill AE. Some simple solutions for heat-induced tropical circulation. *QJ Royal Met Soc* 1980; **106**: 447–62.
3. Lindzen RS and Nigam S. On the role of sea surface temperature gradients in forcing low-level winds and convergence in the Tropics. *J Atmos Sci* 1987; **44**: 2418–36.
4. Philander SG. *El Niño, La Niña, and the Southern Oscillation*. London: Academic Press, 1990, 289.
5. Neelin JD, Battisti DS and Hirst AC *et al.* ENSO theory. *J Geophys Res* 1998; **103**: 14261–90.
6. Wang C and Picaud J. Understanding ENSO physics—a review. In: Wang C, Xie S-P and Carton J (eds). *Earth's Climate: The Ocean-Atmosphere Interaction*. Washington, DC: American Geophysical Union, 2004, 21–48.
7. Wang C, Deser C and Yu J-Y *et al.* El Niño-Southern Oscillation (ENSO): a review. In: Glynn P, Manzello D and Enochs I (eds). *Coral Reefs of the Eastern Pacific*. Berlin/Heidelberg: Springer Science Publisher, 2016, 85–106.
8. Chao JP. *El Niño and Southern Oscillation Dynamics*. Beijing: China Meteorological Press, 1993.
9. Li CY. *Introduction to Climate Dynamics*. Beijing: China Meteorological Press, 2000, 515.
10. Liu QY, Xie SP and Zheng XT. *Tropical Ocean-Atmosphere Interaction*. Beijing: Higher Education Press, 2013, 145.
11. Suarez MJ and Schopf PS. A delayed action oscillator for ENSO. *J Atmos Sci* 1988; **45**: 3283–7.
12. Battisti DS and Hirst AC. Interannual variability in the tropical atmosphere-ocean model: influence of the basic state, ocean geometry and nonlinearity. *J Atmos Sci* 1989; **45**: 1687–712.
13. Jin FF. An equatorial ocean recharge paradigm for ENSO. Part I: Conceptual model. *J Atmos Sci* 1997; **54**: 811–29.

14. Jin FF. An equatorial ocean recharge paradigm for ENSO. Part II: A stripped-down coupled model. *J Atmos Sci* 1997; **54**: 830–47.
15. Weisberg RH and Wang C. A western Pacific oscillator paradigm for the El Niño-Southern Oscillation. *Geophys Res Lett* 1997; **24**: 779–82.
16. Wang C, Weisberg RH and Virmani JI. Western Pacific interannual variability associated with the El Niño-Southern Oscillation. *J Geophys Res* 1999; **104**: 5131–49.
17. Picaut J, Ioualalen M and Menkes C *et al*. Mechanism of the zonal displacements of the Pacific warm pool: implications for ENSO. *Science* 1996; **274**: 1486–9.
18. Wang C. A unified oscillator model for the El Niño-Southern Oscillation. *J Climate* 2001; **14**: 98–115.
19. McCreary JP. A model of tropical ocean-atmosphere interaction. *Mon Wea Rev* 1983; **111**: 370–87.
20. Zebiak SE and Cane MA. A model El Niño-Southern Oscillation. *Mon Wea Rev* 1987; **115**: 2262–78.
21. McCreary JP and Anderson DLT. An overview of coupled ocean-atmosphere models of El Niño and the Southern Oscillation. *J Geophys Res* 1991; **96**: 3125–50.
22. Wyrtki K. El Niño—the dynamic response of the equatorial Pacific Ocean to atmospheric forcing. *J Phys Oceanogr* 1975; **5**: 572–84.
23. Wyrtki K. Water displacements in the Pacific and the genesis of El Niño cycles. *J Geophys Res* 1985; **90**: 7129–32.
24. Rasmusson EM and Carpenter TH. Variations in tropical sea surface temperature and surface wind fields associated with the Southern Oscillation/El Niño. *Mon Wea Rev* 1982; **110**: 354–84.
25. Wang C and Weisberg RH. The 1997–98 El Niño evolution relative to previous El Niño events. *J Climate* 2000; **13**: 488–501.
26. Mayer DA and Weisberg RH. El Niño-Southern Oscillation-related ocean-atmosphere coupling in the western equatorial Pacific. *J Geophys Res* 1998; **103**: 18635–48.
27. Wang C. On the atmospheric responses to tropical Pacific heating during the mature phase of El Niño. *J Atmos Sci* 2000; **57**: 3767–81.
28. Deser C and Wallace JM. Large-scale atmospheric circulation features of warm and cold episodes in the tropical Pacific. *J Climate* 1990; **3**: 1254–81.
29. Zebiak SE. Diagnostic studies of Pacific surface winds. *J Climate* 1990; **3**: 1016–31.
30. Wang B, Wu R and Fu X. Pacific–East Asian teleconnection: how does ENSO affect East Asian climate? *J Climate* 2000; **13**: 1517–36.
31. Vecchi GA. The termination of the 1997–98 El Niño. Part II: Mechanisms of atmospheric change. *J Climate* 2006; **19**: 2647–64.
32. Xie SP and Zhou ZQ. Seasonal modulations of El Niño-related atmospheric variability: Indo–western Pacific Ocean feedback. *J Climate* 2017; **30**: 3461–72.
33. Kug JS and Kang IS. Interactive feedback between ENSO and the Indian Ocean. *J Climate* 2006; **19**: 1784–801.
34. Xie SP, Hu KM and Hafner J *et al*. Indian Ocean capacitor effect on Indo–western Pacific climate during the summer following El Niño. *J Climate* 2009; **22**: 730–47.
35. Picaut J and Delcroix T. Equatorial wave sequence associated with the warm pool displacement during the 1986–1989 El Niño and La Niña. *J Geophys Res* 1995; **100**: 18398–408.
36. Picaut J, Masia F and du Penhoat Y. An advective-reflective conceptual model for the oscillatory nature of the ENSO. *Science* 1997; **277**: 663–6.
37. Wang B and Fang Z. Chaotic oscillations of tropical climate: a dynamic system theory for ENSO. *J Atmos Sci* 1996; **19**: 2786–802.
38. Sun DZ. El Niño: a coupled response to radiative heating? *Geophys Res Lett* 1997; **24**: 2031–4.
39. Liang J, Yang XQ and Sun DZ. The effect of ENSO events on the tropical Pacific mean climate: insights from an analytical model. *J Climate* 2012; **25**: 7590–606.
40. Liang J, Yang XQ and Sun DZ. Factors determining the asymmetry of ENSO. *J Climate* 2017; **30**: 6097–106.
41. Takahashi K and Dewitte B. Strong and moderate nonlinear El Niño regimes. *Clim Dyn* 2016; **46**: 1627–45.
42. Guckenheimer J, Timmermann A and Dijkstra H *et al*. (Un)predictability of strong El Niño events. *Dynamics and Statistics of the Climate System* 2017; **2**: 1–12.
43. Kohyama T and Hartmann DL. Nonlinear ENSO warming suppression (NEWS). *J Climate* 2017; **30**: 4227–51.
44. Lau KM. Elements of a stochastic-dynamical theory of long-term variability of the El Niño-Southern Oscillation. *J Atmos Sci* 1985; **42**: 1552–8.
45. Penland C and Sardeshmukh PD. The optimal growth of tropical sea surface temperature anomalies. *J Climate* 1995; **8**: 1999–2024.
46. Moore AM and Kleeman R. Stochastic forcing of ENSO by the intraseasonal oscillation. *J Climate* 1999; **12**: 1199–220.
47. Philander SG and Fedorov A. Is El Niño sporadic or cyclic? *Annu Rev Earth Planet Sci* 2003; **31**: 579–94.
48. Kessler WS. Is ENSO a cycle or a series of events? *Geophys Res Lett* 2003; **29**: 2125.
49. Landsea CW and Knaff JA. How much skill was there in forecasting the very strong 1997–98 El Niño. *Bull Amer Meteor Soc* 2000; **81**: 2107–19.
50. Gebbie G, Eisenman I and Wittenberg AT *et al*. Modulation of westerly wind bursts by sea surface temperature: a semi-stochastic feedback for ENSO. *J Atmos Sci* 2007; **64**: 3281–95.
51. Li CY and Liao Q. The exciting mechanism of tropical intraseasonal oscillation to El Niño event. *J Tropical Meteor* 1998; **4**: 113–21.
52. Wang HJ, Chen H and Xue F *et al*. The Intraseasonal oscillation and its inter-annual variability—a simulation study. *Acta Meteorologica Sinica* 2001; **15**: 49–58.
53. Li CY. Frequent activities of stronger aerotroughs in East Asia in wintertime and the occurrence of the El Niño event. *Sci China Ser B* 1989; **32**: 976–85.
54. Li CY. Interaction between anomalous winter monsoon in East Asia and El Niño events. *Adv Atmos Sci* 1990; **7**: 36–46.
55. Zheng YX, Zhang RH and Bourassa MA. Impact of East Asian winter and Australian summer monsoons on the enhanced surface westerlies over the western tropical Pacific Ocean preceding the El Niño onset. *J Climate* 2014; **27**: 1928–44.
56. Wang T, Guo D and Gao Y *et al*. Modulation of ENSO evolution by strong tropical volcanic eruptions. *Clim Dyn* 2017; **51**: 2433–53.
57. An S-I. Interannual variations of the tropical ocean instability wave and ENSO. *J Climate* 2008; **21**: 3680–6.
58. Mantua NJ and Battisti DS. Evidence for the delayed oscillator mechanism for ENSO: The ‘observed’ oceanic Kelvin mode in the far western Pacific. *J Phys Oceanogr* 1994; **24**: 691–9.
59. Gill AE. Elements of coupled ocean-atmosphere models for the tropics. In: Nihoul JC (ed.). *Coupled Ocean-Atmosphere Models*. Elsevier Oceanography Series. Amsterdam, Elsevier, 1985, 303–27.

60. Hirst AC. Unstable and damped equatorial modes in simple coupled ocean-atmosphere models. *J Atmos Sci* 1986; **43**: 606–32.
61. Philander SGH, Yamagata T and Pacanowski RC. Unstable air-sea interactions in the tropics. *J Atmos Sci* 1984; **41**: 604–13.
62. Yamagata T. Stability of a simple air-sea coupled model in the tropics. In: Nihoul JCJ (ed.). *Coupled Ocean-Atmosphere Models*. Elsevier Oceanography Series. Amsterdam, Elsevier, 1985, 637–57.
63. Chao J and Zhang R. The air-sea interaction waves in the tropics and their instabilities. *Acta Meteor Sinica* 1988; **2**: 275–87.
64. Yang XQ and Huang SS. Intraseasonal oscillation mode in a linear air-sea interaction system. *J Tropical Meteorology* 1993; **9**: 202–10 (in Chinese).
65. Zhang RH and Huang RH. Dynamical roles of zonal wind stresses over the tropical Pacific on the occurring and vanishing of El Niño Part I: Diagnostic and theoretical analyses. *Scientia Atmospherica Sinica* 1998; **22**: 587–99 (in Chinese).
66. Zhang RH and Chao JP. Some new aspects in understanding of ENSO cycle. *Climatic and Environmental Research* 2002; **7**: 175–83 (in Chinese).
67. Zhang R and Chao J. Unstable tropical air-sea interaction waves and their physical mechanisms. *Adv Atmos Sci* 1993; **10**: 61–70.
68. Hirst AC. Slow instabilities in tropical ocean basin-global atmosphere models. *J Atmos Sci* 1988; **45**: 830–52.
69. Wang C and Weisberg RH. Stability of equatorial modes in a simplified coupled ocean-atmosphere model. *J Climate* 1996; **9**: 3132–48.
70. Wang C and Weisberg RH. On the ‘slow mode’ mechanism in ENSO-related coupled ocean-atmosphere models. *J Climate* 1994; **7**: 1657–67.
71. Neelin JD. The slow sea surface temperature mode and the fast-wave limit: analytic theory for tropical interannual oscillations and experiments in a hybrid coupled models. *J Atmos Sci* 1991; **48**: 584–606.
72. Jin FF and Neelin JD. Modes of interannual tropical ocean-atmosphere interaction—a unified view. Part I: Numerical results. *J Atmos Sci* 1993; **50**: 3477–503.
73. Neelin JD and Jin FF. Modes of interannual tropical ocean-atmosphere interaction—a unified view. Part II: Analytical results in the weak-coupling limit. *J Atmos Sci* 1993; **50**: 3504–22.
74. Jin FF, An S-I and Timmermann A *et al*. Strong El Niño events and nonlinear dynamical heating. *Geophys Res Lett* 2003; **30**: 1120.
75. Wang B. Transition from a cold to a warm state of the El Niño-Southern Oscillation cycle. *Meteorol Atmos Phys* 1995; **56**: 17–32.
76. Fu CB and Fletcher J. Two types of warming over equator during El Niño [J]. *Chin Sci Bull* 1985; **8**: 596–9 (in Chinese).
77. Trenberth KE and Stepaniak DP. Indices of El Niño evolution. *J Climate* 2001; **14**: 1697–701.
78. Larkin NK and Harrison DE. Global seasonal temperature and precipitation anomalies during El Niño autumn and winter. *Geophys Res Lett* 2005; **32**: L16705.
79. Yu J-Y and Kao H-Y. Decadal changes of ENSO persistence barrier in SST and ocean heat content indices: 1958–2001. *J Geophys Res* 2007; **112**: n/a.
80. Ashok K, Behera SK and Rao SA *et al*. El Niño Modoki and its possible teleconnection. *J Geophys Res* 2007; **112**: C11007.
81. Kao HY and Yu JY. Contrasting eastern-Pacific and central-Pacific types of ENSO. *J Climate* 2009; **22**: 615–32.
82. Kug J-S, Jin F-F and An S-I. Two types of El Niño events: cold tongue El Niño and warm pool El Niño. *J Climate* 2009; **22**: 1499–515.
83. Yu J-Y, Kao H-Y and Lee T. Subtropics-related interannual sea surface temperature variability in the central equatorial Pacific. *J Climate* 2010; **23**: 2869–84.
84. Vimont D, Battisti D and Hirst A. Footprinting: a seasonal connection between the tropics and mid-latitudes. *Geophys Res Lett* 2001; **28**: 3923–6.
85. Vimont DJ, Wallace JM and Battisti DS. The seasonal footprinting mechanism in the Pacific: implications for ENSO. *J Climate* 2003; **16**: 2668–75.
86. Vimont DJ, Alexander M and Fontaine A. Midlatitude excitation of tropical variability in the Pacific: the role of thermodynamic coupling and seasonality. *J Climate* 2009; **22**: 518–34.
87. Chiang JCH and Vimont DJ. Analogous Pacific and Atlantic meridional modes of tropical atmosphere–ocean variability. *J Climate* 2004; **17**: 4143–58.
88. Chang P, Zhang L and Saravanan R *et al*. Pacific meridional mode and El Niño–Southern oscillation. *Geophys Res Lett* 2007; **34**: L16608.
89. Ma J, Xie SP and Hu HM. Contributions of the North Pacific meridional mode to ensemble spread of ENSO prediction. *J Climate* 2017; **30**: 9167–81.
90. Xie SP. A dynamic ocean–atmosphere model of the tropical Atlantic decadal variability. *J Climate* 1999; **12**: 64–70.
91. Fang XH and Mu M. A three-region conceptual model for central Pacific El Niño including zonal advective feedback. *J Climate* 2018; **31**: 4965–79.
92. Fang XH and Zheng F. Simulating eastern- and central-Pacific type ENSO using a simple coupled model. *Adv Atmos Sci* 2018; **35**: 671–81.
93. Chen D, Lian T and Fu CB *et al*. Strong influence of westerly wind bursts on El Niño diversity. *Nat Geosci* 2015; **8**: 339–45.
94. Levine A, Jin FF and McPhaden MJ. Extreme noise–extreme El Niño: how state-dependent noise forcing creates El Niño–La Niña asymmetry. *J Climate* 2016; **29**: 5483–99.
95. Dong L and McPhaden MJ. Unusually warm Indian Ocean sea surface temperatures help to arrest development of El Niño in 2014. *Sci Rep* 2018; **8**: 2249–58.
96. Luo JJ., Sasaki W and Masumoto Y. Indian Ocean warming modulates Pacific climate change. *Proc Natl Acad Sci USA* 2012; **109**: 18701–6.
97. Yu JY, Kao PK and Paek H *et al*. Linking emergence of the central Pacific El Niño to the Atlantic multidecadal oscillation. *J Climate* 2015; **28**: 651–62.
98. Levine AF, McPhaden MJ and Frierson DM. The impact of the AMO on multidecadal ENSO variability. *Geophys Res Lett* 2017; **44**: 3877–86.
99. Li XC, Xie SP and Gille ST *et al*. Atlantic-induced pan-tropical climate change over the past three decades. *Nat Geosci* 2016; **6**: 275–9.
100. Li F, Wang HJ and Liu JP. The strengthening relationship between Arctic oscillation and ENSO after the mid-1990s. *Int J Climatol* 2014; **34**: 2515–21.
101. Chen D, Wang HJ and Sun J *et al*. Pacific multi-decadal oscillation modulates the effect of Arctic oscillation and El Niño southern oscillation on the East Asian winter monsoon. *Int J Climatol* 2018; **38**: 2808–18.
102. Zheng XT, Xie SP and Lv LH *et al*. Intermodel uncertainty in ENSO amplitude change tied to Pacific Ocean warming pattern. *J Climate* 2016; **29**: 7265–79.
103. Cai WJ, Borlace S and Lengaigne M *et al*. Increasing frequency of extreme El Niño events due to greenhouse warming. *Nat Clim Change* 2014; **4**: 111–6.