Chapter 6. The El Niño-Southern Oscillation

Reference books:


6.1 The ENSO Phenomenon and Effects
6.1.1. The normal tropical Pacific and its annual cycle

The tropical Pacific extends from the coast of South America in the eastern Pacific to the various islands and land masses of Australia and Indonesia that form the so-called maritime continent – a collection of land masses and islands (Fig. 1). The equator runs from Ecuador in the east (at 80W) to Indonesia in the west, spanning 151 degrees (16,778km) from Ecuador to Halmahera Island at 129E.

![Fig. 1. The tropical Pacific, including the definition of the four Niño regions. Niño 1 and 2 (0-10S, 90W-80W); Niño 3 (5N-5S, 150W-90W); Niño 3.4 (5N-5S, 170W-120W); Niño 4 (5N-5S, 160E-150W). Image downloaded from http://www.hko.gov.hk/lrf/enso/enso-backgnd.htm.](image)

The normal climatic state in and over the tropical Pacific is given by a cartoon in Fig. 2. The surface of the western Pacific is warm and the atmosphere above it is rainy, with the rain coming from deep cumulonimbus clouds. The air rises in the region of the warm water and the rising air is characterized by low pressure at the surface. The winds across the surface of the tropical Pacific blow westward into the region of low pressure, consistent with the westward trade winds (also see left panel of Fig. 3 for observed surface wind and sea surface temperature (SST)). The rising motion in the warm region reaches the tropopause and returns eastward aloft and completes the circuit by descending in the eastern Pacific, leading to higher pressure at the surface. This tropical Pacific-wide circuit of air proceeding westward at the surface, rising over the (warm) region of persistent precipitation, returning eastward aloft, and descending over the cool eastern
Pacific, is called the *Walker circulation* (Fig. 2). Associated with the Walker circulation is the low surface pressure in the western Pacific and the high surface pressure in the east (see Fig. 4 for observed sea level pressure (SLP)). The warm pool precipitation extends eastward into the Pacific north of the equator at about 7N as a linear feature, the INertropical Convergence Zone (ITCZ), and south eastward into the South Pacific as the South Pacific Convergence Zone (SPCZ; see right panel of Fig. 3 and Fig. 5). The ITCZ lies over a band of warm water, which extends from the warm pool eastward into the eastern Pacific (Fig. 3) and coincides with the warm eastward ocean current, the north equatorial counter current (NECC) introduced earlier in this course.

Fig. 2. Schematic of the normal condition of the coupled atmosphere-ocean system in the tropical Pacific during boreal winter. The color contours on the surface represent sea-surface temperature (SST). Red: warm; blue: cold. Thick black arrows represent winds; surface white arrows represent zonal surface currents; the thin black arrows on the surface of the eastern basin show Ekman divergence (surface currents).

Fig. 3. (left) Annual mean SST (contours) and surface winds (arrows) in the tropical Pacific Ocean; the thick gray lines mark the locations of the INertropical Convergence
Zone (ITCZ) north of the equator and South Pacific Convergence Zone (SPCZ); (right) schematic diagram showing surface wind and ITCZ and SPCZ.

Fig. 4. Annual mean (1979-1995) global sea level pressure.

Fig. 5. Annual mean precipitation over the Pacific Ocean. The ITCZ and SPCZ are clearly shown by the strong precipitation bands.
Fig. 6 shows the temperature structure of the ocean near the equator beneath the surface of the ocean. Across the entire ocean, there is a region of sharp vertical temperature gradients (the thermocline) entered at the 20°C isotherm. The thermocline vertically divides the ocean into only two regions, one with warm temperatures and one where the temperatures are cold (Fig. 6). Consequently, a shoaling thermocline represents upwelling, when colder subsurface water comes up to approach the surface. The thermocline is a near-ubiquitous property of the oceans. The thermocline is deep in the west and shallow in the east (Figs 2 and 6). This tilted thermocline is driven by the westward surface winds (Fig. 3) – the surface expression of the Walker circulation. The westward surface winds (often referred to as surface easterlies) driven the Ekman divergence on the equator, shoal the thermocline and induce equatorial upwelling, causing cold SST in the eastern equatorial basin (Figs 2, 3 and 6).

The currents at the surface and below the surface are shown in a cross-sectional diagram in Fig. 7. Although the details will vary at different longitudes, the surface currents are generally in the direction of the wind near the equator (the south equatorial current, SEC) and against the winds north of the equator (the NECC). The NECC is within the ITCZ. Below the surface, and under the westward-moving SEC, lies a rapidly moving current (~1m/s) to the east, opposite to the direction of the winds – the equatorial undercurrent (EUC). The EUC is in the thermocline and is about 200km wide and 100m deep. The heavy arrows below the surface of the ocean show upwelling on the equator, poleward Ekman divergence on either side of the equator, and equatorward replenishment of the surface-diverging water at a few hundred meters depth.
The annual cycle of tropical SST is shown in Fig. 8. While the midlatitude ocean is warmest in (northern) summer and coldest in winter, the tropics has March-April as its warmest period and September-October as its coolest. Along the equator, the deviations from the annual mean clearly propagate westward (not shown) with the majority of annual amplitude confined to the eastern third of the tropical Pacific. The maximum seasonal SST variations occur in the eastern Pacific, and the SST becomes the coldest in northern fall-winter.
6.6.2. The phases and evolution of ENSO

**ENSO Phases:** Superimposed on the normal state of the tropical Pacific is an irregular cycle of warming and cooling of the eastern Pacific with attendant atmospheric and oceanic effects, which is referred to as ENSO. Fig. 9 shows conditions in and over the tropical Pacific during warm phases of ENSO – El Niño (top panel) and during cold phases of ENSO – La Niña (bottom panel). A measure of the strength of the Walker circulation is the difference in the surface pressure between the east (Tahiti) and west (Darwin) – this difference is conventionally called the Southern Oscillation Index (SOI; see Fig. 1 for locations of Tahiti and Darwin). Fig. 10 shows the SLP anomalies and SOI from 1982-2003.

![Diagram of ENSO phases](image)

The eastern Pacific warms, and can warm to such an extent that the temperature across the entire tropical Pacific becomes almost uniform to the temperature of the western Pacific. The warm phase of ENSO is due to a failure of the eastern Pacific to stay cold.
Consistent with this point of view is the relaxation of the westward surface winds (weaker than normal easterly winds which implies westerly anomalous winds), which produces less upwelling and therefore less cooling (Fig. 9, top). Consistent with weaker westward winds, the thermocline is not as tilted, and any upwelling in the eastern Pacific would bring warmer water to the surface. When the cooling in the eastern Pacific is totally gone and the westward surface winds relaxed to almost zero, the warm phase of ENSO is as strong as it can be, and the temperature over the entire tropical Pacific is uniform and assumes the approximate temperature of the western Pacific. This happened in the strong warm phases of ENSO during 1982-83 and 1997-98.

As the eastern Pacific becomes less cold, the region of persistent precipitation that lies over the warmest water expands eastward into the central Pacific. The normally high SLP of the eastern Pacific becomes lower and the sea-level pressure difference between the western and eastern Pacific decreases. Consistent with this decrease is the weakening of the Walker circulation and the relaxation of the normally westward surface winds. As the central and eastern tropical Pacific becomes warm, the ITCZ moves onto the equator and the line of deep convection assumes its southernmost position and the Hadley circulation becomes stronger.

Fig. 10. SLP anomalies at Tahiti and Darwin (top), and the SOI (bottom) for 1982-2003.

During cold phases of ENSO, the normal cooling of the eastern Pacific becomes even stronger, the SOI and thus the Walker circulation, in general, become stronger. Consistent with this, the surface westward winds become stronger, the tilt of the thermocline becomes greater, the stronger westward winds in the eastern Pacific produce even more
upwelling and, because the thermocline is closer to the surface, the water upwelled is colder. The regions of warmest water in the western Pacific contract westward under the encroachment of cold water in the east and, with the warm water, the region of persistent precipitation contracts westward onto the maritime continent. Excess rainfall in Indonesian and western Australian becomes far more common during cold phases of ENSO (Fig. 9, bottom).

**ENSO evolution:** The phases of ENSO evolve differently each time they appear. The general recurrence time for warm and cold phases is around 4 years, with large variations around this mean. The literature often speaks of an ENSO band from 2 to 7 years (see Fig. 10b).

![Fig. 11. a) Nino3.4 SST Anomaly (°C) Time series](image)

One way of describing ENSO evolution with time is to examine the SST anomalies in various regions of the Pacific defined by Fig. 1. Fig. 11a shows the SST anomalies since 1900 in Nino3.4 region; Fig. 10 shows a measure of the strength of the Walker circulation, the SOI for 1982-2003; and Fig. 12 shows the evolution of El Nino in the Pacific basin. We can infer a number of important properties of the warm and cold phases of ENSO by examining SST data shown in figures 10-12. First, we see that the major
phases of ENSO tend to have expression all the way across the Pacific (Fig. 12), from the coast of South America (Nino 1+2 region) to the western Pacific (Nino 4; see Fig. 1 for the regions). Second that the larger events seem to start around summer, peak near the end of the year, and end before the next summer (Figs 11a and 11c and 12), so that the warm and cold phases last about a year. The El Nino peaks near the end of the year, which is “phase locked” with the annual cycle. Thirdly, that there are stretches of time in which not much is happening in the tropical Pacific (the entire 1930s were noted for having no major warm or cold events and that these times are punctuated by the appearance of large phases of ENSO. The warm phases in 1982-83 and 1997-98 were the largest of the century.

**Global ENSO evolution (warm phase)**

Associated with the SST change, atmospheric fields – surface wind, SLP and convection (represented by Outgoing Longwave Radiation (OLR) – also vary (Figs 13A-13B). The lowest OLR (negative OLR anomaly) represents “deep convection” in the tropics where storm clouds reach the top of the troposphere. The high cloud top is cold and thus has low outgoing longwave radiation. The anomalously low OLR (negative anomaly) shows that this region is more cloudy than normal.

Note that in the past few years, some researchers found that for some El Nino events, the strongest warming occur in the central equatorial Pacific basin, rather than in the eastern equatorial basin. These events were referred them to as “central Pacific El Nino” or El
Nino Modoki. However, it is not clear whether the central Pacific El Nino is different from the conventional El Nino or it is simply different phase of the El Nino evolution.

Fig. 13A. Adapted from Fig. 5 of Deser and Wallace (1990).
(a) Regressions of July-November SLP and surface wind upon the equatorial SST index (defined as July-November SST averaged over 6S-6N, 180-80W), based on the period 1946-85. Contour interval=0.25mb/C of the SST index; the zero contour is darkened. Wind vectors are shown only for those grid points whose u or v correlations with the SST index exceed 0.4 in absolute value. (b) As in (a) but for SST and surface wind. Contour interval=0.5C/C of the SST index; the zero contour is darkened and regression coefficients>1C/C are shaded. (c) As in (a) but for OLR and surface wind. OLR regressions are based on the period 1974-89 (1978 missing). Contour interval=10W/m2/C of the SST index; the zero contour is darkened and the positive contour is dashed. Values < -20W/m2/C are shaded. (d) Divergence of surface wind
regressions shown in (a) in units of $10^{-6}/\text{s}/\text{C}$. Solid (dashed) contours indicate anomalous wind convergence (divergence) during warm episodes.

Fig. 13B. Adapted from Fig. 7 of Deser and Wallace (1990). Regressions of December-February (a) SLP and surface wind, (b) SST and surface wind, (c) OLR and surface wind, and (d) surface wind divergence upon the equatorial SST index (defined as December-Feb. SST averaged over 6S-6N, 180-80W). Ploting convection as in Fig. 12A.
6.1.3. ENSO Impacts
The effect of the warming of the eastern Pacific due to El Nino, and the consequent eastward movement of the region of persistent precipitation, is felt throughout the world (Fig. 14). In the tropics, the normally rainy western Pacific becomes drier as the region of persistent precipitation moves eastward into the central Pacific. Droughts in Indonesia and in eastern Australia become far more common during the warm phases of ENSO. Rainfall in the normally arid coastal plains of Peru becomes far more likely and warm water spreads north and south along the western coasts of the North and South American continents. The temperature and rainfall in other selected areas of the world (e.g., Zimbabwe, Madagascar) are similarly affected, and usually Colorado has a good snow (ski) year during El Nino, even though the reasons are either difficult to explain or unknown (likely through atmospheric bridge).

Fig. 14a. El NINO impacts.
Fig. 14b. La Nina impacts. Not necessarily a simple “opposite” with the El Nino impacts due to the complexity of the climate system (especially precipitation).
Warming associated with El Nino, especially with the maximum warming in the central Pacific Ocean, can cause large impacts on Indian summer monsoon rainfall and often cause severe draught in India (Figure 15).

ENSO can also have direct impacts on marine life: corals, fishery, birds (see Clarke 2008), etc.. For example, during normal years, upwelling along the west coast of South America bring the nutrients to euphotic zone, and thus favor marine life and fishery. During El Nino, however, upwelling is suppressed and thus the marine food chain is significantly affected. Fishery declines.
6.2 ENSO Mechanisms

Observations show two opposite phases of ENSO: warm phase (El Nino) and cold phase (La Nina). How does the turnabout between the two phases take place in the coupled ocean-atmosphere system? Below we introduce a few existing theories to elaborate this oscillatory phenomenon. These theories tend to produce ENSO like oscillations.

6.2.1 The delayed-oscillator theory


Fig. 16. Zonal displacement of the (left) 28.5°C SST averaged over 4S-4N and (right) SOI from 1950-1998. The SOI axis is inverted. The monthly 28.5C zonal displacement has been smoothed with a 3-point running mean filter and the SOI has been filtered three times with this filter (from Clarke et al. 2000).

Fig. 16 shows that the SOI and the zonal displacement of the eastern edge of the warm pool (28.5C isotherm) are highly correlated. When the warm pool moves we expect the deep atmospheric convection and zonal winds it generates to move with it. Thus we expect the cloudiness at the edge of the warm pool (as measured by the OLR 240 W/m2 isoline) and the -4m/s eastward wind isotach to track closely with the 28.5C isotherm. Fig. 17 shows that these isolines do track each other quite closely. Due to the rain falling on it, the surface waters of the warm pool are fresher than waters to the east of it and consequently there is a sharp salinity front, which also closely tracks the other isolines (not shown).
Fig. 17. Zonal displacement of the equatorial (4S-4N average) 28.5°C SST isotherm (heavy solid line: a proxy of the east edge of the warm pool), zonal displacement of the equatorial -4m/s westerly surface isotach (dash-dot line) and zonal displacement of the 240 W/m² equatorial OLR isoline (thin solid line; 5S-5N average). The vertical dotted line is the mean position of the 28.5°C isotherm (172.2W). Monthly values of SST, wind and OLR have been smoothed with a double 5-month running mean filter. From Shu and Clarke 2002.

The delayed oscillator theory often emphasizes SST anomalies (SSTAs) in the eastern equatorial Pacific, because in many models air-sea coupling is strongest there. But observations suggest that the main coupling occurs at the edge of the western equatorial Pacific warm pool where SLP, surface wind and OLR obtain their maximum amplitudes (Figs 13A and 13B). Model error occurs because, to a first approximation, the anomalous heating that typically takes place in the atmospheric component of the model depends primarily on the SSTAs which are biggest in the eastern equatorial Pacific (Clarke 2008, section 7.2). But the anomalous heating of the atmosphere depends not only on the SSTA but also on whether the total SST is high enough for deep atmospheric convection; although SSTAs are largest in the eastern equatorial Pacific, the total SST is usually too low for deep atmospheric convection. Consequently the most important SSTAs in ENSO
dynamics are over the warm water in the western equatorial Pacific, particularly in the west-central equatorial Pacific at the edge of the western equatorial Pacific warm pool.

Because interannual displacement of the eastern edge of the western equatorial Pacific warm pool rather than eastern equatorial Pacific SSTA seems central to ENSO dynamics, Clarke (2008) discussed delayed oscillator theory for zonal equatorial displacements of the eastern edge of the western equatorial Pacific warm pool.

**a) Formulation of a warm pool displacement/delayed oscillator model of ENSO**

Let’s consider the zonal displacement of the eastern part of the equatorial western Pacific warm pool: the displacement of 28.5°C isotherm (Figs 16 and 17). Let

\[ x = x(t) \quad (1) \]

be the **anomalous eastward displacement of the 28.5°C isotherm on the eastern side of the warm pool**. Existing studies suggest that the warm pool and 28.5°C isotherm displacement are mainly due to zonal advection by an anomalous zonal ocean current \( u' \), with relatively minor contributions from other processes (such as anomalous net solar surface heat flux and evaporation, vertical advection and mixing). To simplify the demonstration of the warm pool displacement/delayed oscillation mechanism, it is assumed that anomalous zonal advection is the only mechanism operating and write (Clarke 2008):

\[ \frac{dx}{dt} = u'(t). \quad (2) \]

The anomalous zonal ocean current \( u' \) consists of three parts which are denoted \( u'_1, u'_2, u'_3 \). When \( x \) is positive, the warm pool will be eastward of its normal position and anomalous deep atmospheric convection will generate westerly surface equatorial winds (see Figs 13A and 13B). These anomalous eastward winds generate an anomalous eastward **local flow** \( u'_1 \) and **two delayed negative feedback flows** \( u'_2 \) and \( u'_3 \). Note that we omitted the advection effect of mean zonal surface current in equation (2), which is close to zero at the convergence region near the date line (Clarke et al 2000). The mean flow effects will be discussed later. The local anomalous flow is essentially in phase with the anomalous wind stress and \( x(t) \) and we write:

\[ u'_1(t) = ax(t). \quad (3) \]

If the anomalous zonal equatorial flow \( u' \) were entirely due to the local current \( u'_1 \), then the right-hand side of (2) would be given by (3). The solution to this equation would be an exponentially growing solution (the first order partial differential equation). Physically, this is because a small eastward displacement of the warm pool (small \( x(t) \)) will cause anomalous deep atmospheric convection, westerly wind anomalies (Figs 13 and 17) and thus an eastward surface current. This eastward current displaces the warm pool further to the east, generating increased deep atmospheric convection, a stronger
westerly wind anomaly, an increased eastward surface current and therefore further eastward warm pool displacement, etc.

The second current anomaly \( u'_2 \) results from wave propagation (a delayed part). The anomalous westerly winds generate “upwelling” Rossby waves that propagate westward: with negative sea level anomalies “off the equator” and eastward surface currents on the equator (see Figs 18 and 19). The Rossby waves reflect at the western ocean boundary and return as equatorial Kelvin waves with westward surface currents (Fig. 19) in order to satisfy the western equatorial boundary condition of “no normal flow \((u=0)\)”. When these equatorial Kelvin waves reach the eastern edge of the equatorial warm pool, the westward equatorial currents will try to return the 28.5°C isotherm to its original position, i.e., \( u'_2 \) provides a “negative” feedback. Mathematically, since the currents are of opposite sign to the locally driven current \( u'_1 \) we may write

\[
u'_2 = -bx(t - \Delta)
\]

where \( \Delta \) is the time it takes for Rossby waves to propagate from the region of strong ocean-atmosphere coupling (near the date line) to the western boundary, reflect as an equatorial Kelvin wave and return to the region of strong ocean-atmosphere coupling. Both observations and theory suggest that it should take Rossby waves about 3 months from the central equatorial basin to the western boundary and, theoretically, it should take the equatorial Kelvin waves 1 month to get back so \( \Delta = 4 \) months. But when the warm pool moves these travel times will change. Based on the result that the first vertical mode (baroclinic) Kelvin wave (speed c) and the first baroclinic mode, first meridional mode Rossby wave (speed c/3) dominate the flow averaged over 4S-4N, we write:

\[
\Delta = 4 \text{months} + \frac{3x(t - \Delta)}{c} + \frac{x(t)}{c}.
\]

The “4-month” represents the time for the Rossby wave (generated by wind at the mean location of 28.5°C) to cross the basin to the western boundary and the reflected Kelvin wave to return to the same location. The 2nd term of equation (5) represents the time it takes the first meridional mode Rossby wave to propagate for distance \( x(t-\Delta) \) because the Rossby wave starts from the previous time \((t-\Delta)\) when the displacement of the 28.5°C line is at \((t-\Delta)\); the 3rd term represents the time it takes the reflected Kelvin wave to propagate for distance \( x(t) \), the displacement of the 28.5°C line at time \( t \). Note that equation (5) defines a non-linear problem for \( \Delta \) which is rapidly solved by iteration. At each time step \( t \) during the integration of equation (2) the initial guess for \( \Delta(t) \) is \( \Delta \) at the previous time step.

The third contribution to the current anomaly of \( u'_1, u'_3 \), is of similar form to the second. The second contribution \( u'_2 \) involves generation of Rossby waves and reflection of these from the western Pacific Ocean boundary while the third contribution involves the generation of equatorial Kelvin waves and their reflection from the eastern Pacific boundary as Rossby waves (Fig. 19). Both contributions are qualitatively the same, exerting a delayed negative feedback on the flow (Fig. 19). To keep the mathematics
simple, we ignore $u'$. Thus the governing equation (2), using (3) and (4), can be written as the non-linear equation:
\[
\frac{dx}{dt} = ax(t) - bx(t - \Delta).
\] (6)

Fig. 18. (left) A westerly wind stress is applied in the middle of the Tropical Pacific Ocean. The white areas represent the coasts. The stronger wind stress is red and the weaker wind stress is blue; (right) Ocean response to the wind stress on the left. The red signal is the Kelvin wave which propagates with a positive sea surface height anomaly and thus deepens the thermocline in the eastern Pacific. The blue-green signal is the Rossby wave which propagates with a negative sea surface height anomaly and thus rises the thermocline in the western Pacific. The dark arrows represent zonal surface currents forced by the westerly winds; Source: IRI.

Fig. 19. Propagation of the Kelvin and Rossby waves in the tropical Pacific. Pink arrows show reflected surface currents at the eastern and western boundaries after the Kelvin and Rossby waves impinge onto the eastern and western boundaries, respectively. Source IRI.
b) Model solutions

We begin our analysis of (6) by considering the simple case when the delay \( \Delta \) is constant. This equation has the same form as the delayed oscillator models from earlier studies (say Battisti and Hirst 1989; Suarez and Schopf 1988 except for their cubic damping term). These earlier delayed oscillator models (see Sarachik and Cane 2010 for summary) differ physically from equation (6) in that they are not based on warm pool displacement as the main component of ENSO physics.

Now with the local and delayed processes included in equation (6), we wish to find out under what conditions solutions to (6) can “oscillate” at a period range of observed ENSO (2-7 years).

Looking for solutions (for \( \Delta = \text{constant} \)) in the form

\[
 x(t) = Ae^{\sigma t} \cos \omega t \quad (7)
\]

Equation (7) has growing (or decaying) amplitude \( e^{\sigma t} \) of an oscillation \( \cos \omega t \), with growth (decay) rate of \( \sigma \), and oscillating frequency of \( \omega \). Substituting \( Ae^{(\sigma + i\omega) t} \) for \( x(t) \) in equation (6), we obtain:

\[
 \sigma + i\omega = a - be^{-(\sigma + i\omega) \Delta} \quad (8)
\]

Here we are interested in the solutions with low frequencies relevant to ENSO. Therefore, we look for the solutions with

\[
 \zeta = \omega \Delta \leq 1. \quad (9)
\]

Equation (9) means that we are looking for low frequency, long period (T) cases with period \( T \geq 2\pi \Delta \). For \( \Delta = 4\) months, \( T \) is longer than 25 months or 2 years, which is within the observed period range of ENSO. By manipulating the above equations, we finally obtain the solutions to the period of the oscillation (T) and the e-folding growth rate time scale \( \sigma^{-1} \):

\[
 T = \sqrt{2\pi \Delta} / (\ln b \Delta - a \Delta + 1)^{1/2} \quad (10)
\]

and

\[
 \sigma^{-1} = 3\Delta / (2\ln b \Delta + a \Delta - 1). \quad (11)
\]

The formulae (10) and (11) indicate that the model oscillations can be of ENSO periodicity for reasonable parameter values. To show this, let’s first estimate \( a \) and \( b \). For typical current speeds of \(-0.2\) m/s associated with the equatorial Kelvin waves (see Clarke 2008 for a detailed discussion), equation (4) yields:

\[
 \sim bx \sim 0.2\text{m/s}. \quad (12)
\]
Since typical amplitudes for $x$ are $\sim 2000$ km, for $\Delta = 4$ months, $b\Delta \sim 1$.  

But the Rossby wave mass flux onto the western boundary, generated remotely through the ENSO air-sea interaction, is comparable to the reflected equatorial Kelvin wave mass flux. Therefore we have

$$ax \sim bx,$$  \hspace{1cm} (14)

and hence

$$b\Delta \sim a\Delta \sim 1.$$  \hspace{1cm} (15)

For such a parameter range, equations (10), (11) and (7) show that growing, neutral or decaying long-period solutions are possible.

Notice that the period $T$ of the coupled oscillation (see equation (10)) depends both on the wave transit time $\Delta$ and the time scales $a^{-1}$ and $b^{-1}$ associated with the anomalous advecting flow. Physically, if the negative feedback is sufficiently delayed, there is time for the instability to grow before being damped out by the negative feedback. There is also time, once the damping has overpowered the instability, for the negative feedback to change the sign of $x(t)$, e.g., from positive to negative. Mathematically, in equation (6) when $x(t) = 0$ and $x(t-\Delta) > 0$, the negative feedback $-bx(t-\Delta) < 0$ so that $dx/dt < 0$ and a small negative $x(t)$ subsequently results. This small negative perturbation can then grow and be restrained by delayed negative feedback so that eventually a small positive (eastward) perturbation of the warm pool results. In this way a sequence of warm and cold ENSO episodes occurs. The oscillation periodicity, which is about 2-7 years, is much longer than the 4-month wave transit delay time; although the delay time is crucial to the existence of the period of the oscillation, the oscillation period also depends critically on the relative importance of the growth rate of the instability and the negative feedback.

c) **The influence of other negative feedbacks and non-constant $\Delta$**

There exist other negative feedbacks. One is associated with the easterly (westerly) wind anomalies which form in the far western equatorial Pacific during warm (cold) ENSO episodes (Fig. 13B). These far western equatorial Pacific wind anomalies generate equatorial Kelvin waves with zonal flow opposite to the flow of the growing instability near the date line. Since the equatorial Kelvin waves only take about 1 month to propagate from the western equatorial Pacific to the date line, this negative feedback is essentially in phase with $-x(t)$, and thus its main effect will be to reduce the parameter $a$ in the growth term $ax(t)$. This effect is likely small or negligible (see Shu and Clarke 2002).

Another negative feedback is due to the surface heat flux. Specifically, increased eastward warm pool displacement implies increased deep atmospheric convection, more clouds, more shade, less heating by the sun’s incoming short wave radiation and hence cooling. Consequently, the original SST anomaly tends to be damped out. This effect tends to reduce the growth term $ax(t)$. For realistic interannual periodicity (2-7 years),
however, this feedback mechanism requires that parameter values must occur only over a narrow range.

So far the delay $\Delta$ has been taken to be constant when we discuss solutions rather than the more accurate time-varying $x$-dependent value shown in equation (5). One way ENSO irregularity can arise is via random wind-forced currents affecting the eastward displacement of the warm pool (also see section 6.2.3 for stochastic forcing). When $\Delta$ depends on $x$, the simple model $x(t)$ is irregular in both amplitude and $T$ but with dominant interannual periodicity.

6.2.2 The discharge-recharge oscillator theory
Clarke (2008) section 7.3;
Sarachik and Cane (2010), section 7.5.
Please see Jin 1997a,b (Journal Atmospheric Science) for original concepts of recharge oscillator theory.

Here, instead of showing detailed mathematics, we primarily provide physical descriptions of the discharge-recharge oscillator theory. The thermocline depth of the equatorial ocean is one of the main variables that need to be considered for understanding ENSO dynamics. On the ENSO timescale, the leading ocean dynamical balance is between the pressure gradient force and wind stress over the equatorial band, for instance, within a couple of oceanic Rossby radii of deformation. In other words, the zonal pressure gradient force accompanying the thermocline depth tilt along the equator is largely in a Sverdrup balance with the equatorial wind stress force. A numerical ocean model forced with a slowly changing wind stress anomaly can be used to verify this quasi-equilibrium relation. Using a shallow-water model, we have $He=Hw+\text{int}_t\alpha_x$, where $Hw$ denotes the D20 anomaly in the western Pacific, within one oceanic Rossby radius from the equator; $He$ is the D20 anomaly in the EQ eastern Pacific, and $\text{int}_t\alpha_x$ is proportional to the zonally integrated wind stress in this band.

However, this leading balance of forces only constrains the east-west contrast of the thermocline depth. The absolute depth at the western Pacific or the mean thermocline depth over the equatorial band is not constrained by this balance. The mean D20 depends on the mass adjustment of the entire tropical Pacific ocean and may not be in equilibrium with the slowly varying wind forcing. In the nonequilibrium between the zonal mean thermocline depth and the wind stress forcing resides the memory of the subsurface ocean dynamics and makes the fast wave limit approximation inappropriate for an ENSO-like mode that relies on the subsurface ocean memory. There are two equivalent qualitative approaches to this adjustment process. One is from the wave propagation point of view. The ocean mass adjustment is completed through the oceanic Kelvin and Rossby waves, which propagate in opposite directions and are forced by the basin boundaries to change into opposite wave characteristics through reflections. The damping effect due to the leakage of energy via the boundaries and other damping processes allow this adjustment to finally settle into a quasi-equilibrium in the equatorial region. In this view, one of the two unknowns in the $He$ equation can be related to a boundary condition that determines wave reflections as shown in a number of theoretical studies. The other
approach is to consider that the equatorial wave propagation process is relatively fast for establishing this thermocline slope that extends to the off-equatorial region as a result of the broadness of the atmospheric wind system. The Coriolis force becomes important off the equatorial band, and therefore there will be Sverdrup transport either pumping the mass in or out of the equatorial region depending on wind forcing. Under linear shallow-water dynamics, this Sverdrup transport is accomplished by the Rossby waves. The zonally integrated effect of this Sverdrup transport of mass or, equivalently, heat content will result in the deepening or shoaling of the western Pacific thermocline depth. Thus, although the thermocline tilt along the equator is set up quickly to balance the equatorial wind stress, the D20 of the warm pool takes time to adjust to the zonal integrated meridional transport, which is related to both the wind stress and its curl off the equatorial band.

The physics of the oscillation is summarized in Fig. 20. We begin our discussion of the oscillation at the height of an El Nino (panel 1 of Fig. 20). An initial positive SSTA induces a westerly wind forcing over the central to western Pacific. When the SSTAs are maximally positive, there is anomalous equatorial deep atmospheric convection and the wind anomalies are maximally eastward (westerlies). The wind anomalies cause a twofold ocean response. One part of the response consists of a tilt of the thermocline in phase with the westerly winds as the wind stress forcing is in quasi-steady balance with the zonal pressure gradient (Sverdrup balance). While this tilt affects the thermocline depth in the eastern and western equatorial Pacific, the displacement in the central equatorial Pacific is nearly zero. This means that the zonal mean depth of the 20C isotherm (D20) is not changed. However, two processes can affect the zonal mean D20, or D20 in the central equatorial Pacific basin: (a) Zonal pressure gradient force associated with the tilt of the thermocline causes geostrophic current divergence at the equator (Jin 1997a; thick arrows of Fig. 20). This is because on the equator, geostrophic current v satisfies: $\beta v_\| = \frac{1}{\rho} \frac{\partial \rho}{\partial x}$. When the thermocline anomaly tilts downward toward the east (panel 1 of Fig. 20), sea level tilts upward and $\frac{1}{\rho} \frac{\partial \rho}{\partial x} > 0$. Since $\beta > 0$, v is positive (northward) in northern hemisphere ($y>0$) and negative (southward) in southern hemisphere ($y<0$), causing equatorial heat divergence, shallowing the zonal mean D20, discharging the equatorial heat and reducing the equatorial upper-ocean heat content. (b) The anomalous wind stress curl associated with the equatorial westerlies causes northward (southward) current north (south) of the equator (Clarke et al. 2007; Clarke 2008), due to Sverdrup relation: $\beta v = \text{curl}(\vec{\tau})$. When $\text{curl}(\vec{\tau}) > 0$ (see north of equator in panel 1), v is positive northward; when $\text{curl}(\vec{\tau}) < 0$ (see south of equator in panel 1), v<0 which is negative southward. This will also cause equatorial mass and heat divergence and shoal the mean D20. This discharge gradually reduces the thermocline depth in both the western and eastern Pacific and eventually leads to a cooling trend for the SST anomaly when the warming due to the still positive thermocline depth anomaly in the eastern Pacific is balanced off by the damping process of the SST (when SST increases in the east, evaporation increases and solar shortwave radiation decreases due to enhanced convection). Thus the warm phase of ENSO evolves to the transition phase as shown in panel 2 of Fig. 20. At this time, the SST anomaly cools to zero. The east–west
thermocline tilt anomaly diminishes because it is always in a quasi-equilibrium balance with the zonal wind stress, which has disappeared following the SST anomaly. However, the entire equatorial Pacific thermocline depth and thus the eastern Pacific thermocline depth is anomalously shallow because of the discharge of the equatorial heat content during the warm phase. When the D20 is shallow, normal climatology winds will entrain colder water into the surface layer, and thus cools the SST. It is this anomalous shallow thermocline depth at the transition that allows anomalous cold water to be pumped into the surface layer by climatological upwelling; the SST anomaly then slides into a negative phase. Once the SST anomaly becomes negative, the cooling trend proceeds because the negative SST anomaly will be further amplified through the positive thermocline feedback. That is, the enhanced trades, in response to the cold SST anomaly, deepen the thermocline depth in the western Pacific and lift the thermocline depth up in the east. Thus the oscillation develops into its mature cold phase as shown in panel 3 of Fig. 20. At the same time, the zonal mean thermocline depth over the equatorial Pacific is deepening, as a result of the recharging of the equatorial heat content due to the strengthened trades. This reverses the cooling trend after the mature cold phase and brings it to another transition phase as shown in panel 4. When the cold SST anomaly reduces to zero, the positive zonal mean thermocline depth generated by the recharging process will lead the SST anomaly evolving back to another warm phase.

It has been shown that this oscillator can be either self-excited or stochastically sustained. Its period is robust in the range of 3–5 years. This recharge oscillator model depicts the slow physics of ENSO and also embodies the delayed oscillator without requiring an explicit wave delay.

This simple paradigm does not capture the changes in the western equatorial Pacific preceding the warming in the east. In addition, the idealized oscillator is not phase-locked to the seasonal cycle. The analog model of C. Wang (2001) includes specific processes in the western Pacific, in particular parameterizing changes in western Pacific wind stresses on the equator in terms of thermocline depth. He includes delayed effects and shows that the delayed oscillator and the discharge-recharge equations of Jin are obtained as special cases. Nevertheless, the idealized model does illustrate the idea of oscillation on the observed interannual timescales of ENSO.

Note also that Kessler (2002, GRL) found phases 1-3 by analyzing data; but could not find the transition from phase 4 to phase 1. This suggests that this re-charge oscillator may not always occur, and other mechanisms may exist to kick the system from phase 4 to El Nino phase 1.
6.2.3 Stochastically forced models

Sarachik and Cane (2010), section 7.6.

In linearly unstable model systems, a perturbation grows exponentially without limit. Something must, in reality, equilibrate the system at finite amplitude. Either some nonlinearity limits growth, or the system was not unstable to begin with. Here, we examine non-normal growth in stable, coupled atmosphere-ocean models.
The issue can be stated succinctly as follow: for a coupled linear model of ENSO of the form

\[
\frac{du}{dt} = Au + f, \quad (16)
\]

where \(u\) is the state vector of quantities in the atmosphere and ocean, \(A\) is the linear evolution operator, and \(f\) some combination of nonlinearity and random noise; the linear stability properties of the system are determined by the nature of the eigenvalues of \(A\). If the eigenvalues of \(A\) have a positive real part, then there are exponentially growing normal modes and \(f\) must contain some nonlinear term in order to limit the amplitude of these modes. If the matrix \(A\) is non-normal (due to strains and shears of background flow, etc.), then even if the real parts of eigenvalues of \(A\) are negative, so that all the normal modes decay, there may be transient disturbances that can grow and then decay. If this is the case, then a purely random forcing \(f\) in equation (16) may be sufficient to excite such a set of disturbances. These are the ideas of generalized linear stability theory.

Moore and Kleeman (1999) examined the potential role that tropical variability on synoptic–intraseasonal timescales (especially the Madden-Julian Oscillation (MJO), which is the strongest in the tropical Indian and western Pacific Oceans) can play in controlling variability on seasonal–interannual timescales. The ideas of generalized stability theory are investigated using an intermediate coupled ocean–atmosphere model of the ENSO. The variability on synoptic–intraseasonal timescales is treated as stochastic noise that acts as a forcing function for variability at ENSO timescales. The spatial structure is computed that the stochastic noise forcing must have in order to enhance the variability of the system on seasonal–interannual timescales. These structures are the so-called stochastic optimals of the coupled system, and they bear a good resemblance to variability that is observed in the real atmosphere on synoptic and intraseasonal timescales.

When the coupled model is subjected to a stochastic noise forcing composed of the stochastic optimals, variability on seasonal–interannual timescales develops that has spectral characteristics qualitatively similar to those seen in nature. The stochastic noise forcing produces perturbations in the system that can grow rapidly. The response of the system to the stochastic optimals is to induce perturbations that bear a strong resemblance to westerly and easterly wind bursts frequently observed in the western tropical Pacific. (Fig. 21). In the model, these “wind bursts” can act as efficient precursors for ENSO episodes if conditions are favorable. The response of the system to noise-induced perturbations depends on a number of factors that include 1) the phase of the seasonal cycle, 2) the presence of nonlinearities in the system, 3) the past history of the stochastic noise forcing and its integrated effect, and 4) the stability of the coupled ocean–atmosphere system. Based on their findings, they concur with the view adopted by other investigators that ENSO may be explained, at least partially, as a stochastically forced phenomena, the source of the noise in the Tropics being synoptic–intraseasonal variability, which includes the MJO, and westerly/easterly wind bursts. These ideas fit well with the observed onset and development of various ENSO episodes, including the 1997–98 El Nin˜o event (Fig. 22).
Fig. 21. Adapted from Fig. 7 of Moore and Kleeman (1999). The cyclonic winds on either side of the equatorial western Pacific (panel b) agree well with the observed convective clouds (panel c).

Fig. 22. Time versus longitude sections of surface zonal wind (left), SST (middle), and outgoing longwave radiation (OLR) (right) from Sep. 1996-Aug. 1998. Analyses are based on 5-day averages for 2S-2N for the TAO data, and 2.5S-2.5N for OLR. Black
squares on abscissas of the wind and SST plots indicate longitudes of data availability at the start (top) and end (bottom) of the time-series record. Positive winds are westerly, negative winds are easterly. OLR values below about 235 W/M2 indicate an increased likelihood of deep cumulus cloudiness and heavy convective precipitation. OLR data are from the National Center for Environmental Prediction. From McPhaden (1999).

This suggests that the large-scale intraseasonal forcing, in particular the MJO, is most efficient at producing interannual variance. It is likely that including the MJO forcing, the coupled model may describe the phases and irregularity of ENSO more accurately. Indeed, observations reveal that no two ENSO episodes are completely alike, and, in fact, the episodes of the 1990s are quite different from those of the 1970s and 1980s. This perhaps suggests that no single mechanism is solely responsible for ENSO, and that the relative importance of different physical processes may vary from episode to episode. Even though there is fairly strong observational and modeling evidence that supports the idea that ENSO is, by and large, governed by an oscillator, individual events tend to be episodic rather than part of a regular oscillatory cycle.