Current theories on El Niño-Southern Oscillation: A review

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RESUMEN

Se presenta el estado actual del conocimento sobre la física del fenómeno El Niño-Oscilación del Sur (ENOS). Se describen las principales características basado en observaciones, y se discute la física básica que controla el fenómeno con resultados de modelos de diversa complejidad. ENOS es un fenómeno acoplado del sistema Océano-Atmósfera en el cual los mecanismos de retroalimentación son muy importantes; por lo tanto, es difícil establecer relaciones causales, y a pesar de 20 años de grandes avances en su observación, explicación y predicción, aún quedan preguntas importantes por responder.

PALABRAS CLAVE: El Niño, retroalimentación Océano-Atmósfera, modelos.

ABSTRACT

The current state of knowledge on the physics of the El Niño-Southern Oscillation (ENSO) phenomenon is presented. The main observed features are described, and the basic physics that control the phenomenon are discussed with results from models of various levels of complexity. ENSO is a coupled ocean-atmosphere phenomenon in which feedback mechanisms are very important; therefore clear causality relationships are difficult to establish, and despite 20 years of great advances in observing, explaining and predicting it, important questions still remain.

KEY WORDS: El Niño, Ocean-Atmosphere feedback, models.

1. INTRODUCTION

The purpose of this short contribution to the Special Issue on El Niño impacts in Mexico is to review the air-sea interaction phenomena and other physics involved in El Niño-Southern Oscillation (ENSO). We highlight some of the outstanding and basic features that characterize ENSO, such as its irregularity, duration, and phase locking to the seasonal cycle which are not yet fully understood, and the theories put forward to explain them. In preparing this short review I have benefited particularly from Wang (2001) and Tziperman (2001), where the interested reader can find further information and discussions on the subject.

2. ENSO MAIN PLAYERS

Figure 1a below[1] shows the normal atmospheric and oceanic conditions in the equatorial Pacific ocean, and illustrates what are believed to be the principal players involved in the ENSO phenomenon. The sea-surface temperature (SST) shows warm water (in red) located toward the west, and the cool equatorial waters in the eastern Pacific. The white arrows represent the normal easterly winds (the Trades) blowing along the equator toward the west. The blue sheet below the surface represents the equatorial thermocline, or interface between the warm surface waters and colder deep waters, which has an average depth of about 200 m across

the Pacific but with a characteristic tilt along the equator, being shallow in the east and deep in the west (where the warmest surface waters are located). The thin black poleward arrows in both hemispheres represent the off-equatorial surface Ekman flow driven by the Trades, whose equatorial divergence provides the physical mechanism for the upwelling of deeper water into the surface layer and induces the "cold tongue" of eastern Pacific cool surface waters along the equator. In the atmosphere, the zonal convective cell also called the Walker circulation cell (Walker and Bliss, 1932), results in low level convergence over the warm waters of the western Pacific, low sea-level pressure, upward motion and high precipitation. On the eastern side, where the colder surface waters reside, there is subsidence (downward motion), higher surface pressure and precipitation is inhibited. As will be discussed below, a recently included player in the ENSO game is the meridional subsurface geostrophic flow, which is associated with the tilt of the equatorial thermocline, but is not shown in Figure 1.

During El Niño, the pattern described above changes dramatically (Figure 1b). The warmest surface waters move from the west toward the central Pacific due to advective and sub-surface thermocline processes, and the equatorial cold tongue nearly disappears. The region of deep convection in the atmosphere usually located above the warmest surface waters of the west also shifts eastward. The strong

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[1] All figures in this section were obtained from the very illustrative web page http://www.pmel.noaa.gov/tao/elnino

Normal Conditions
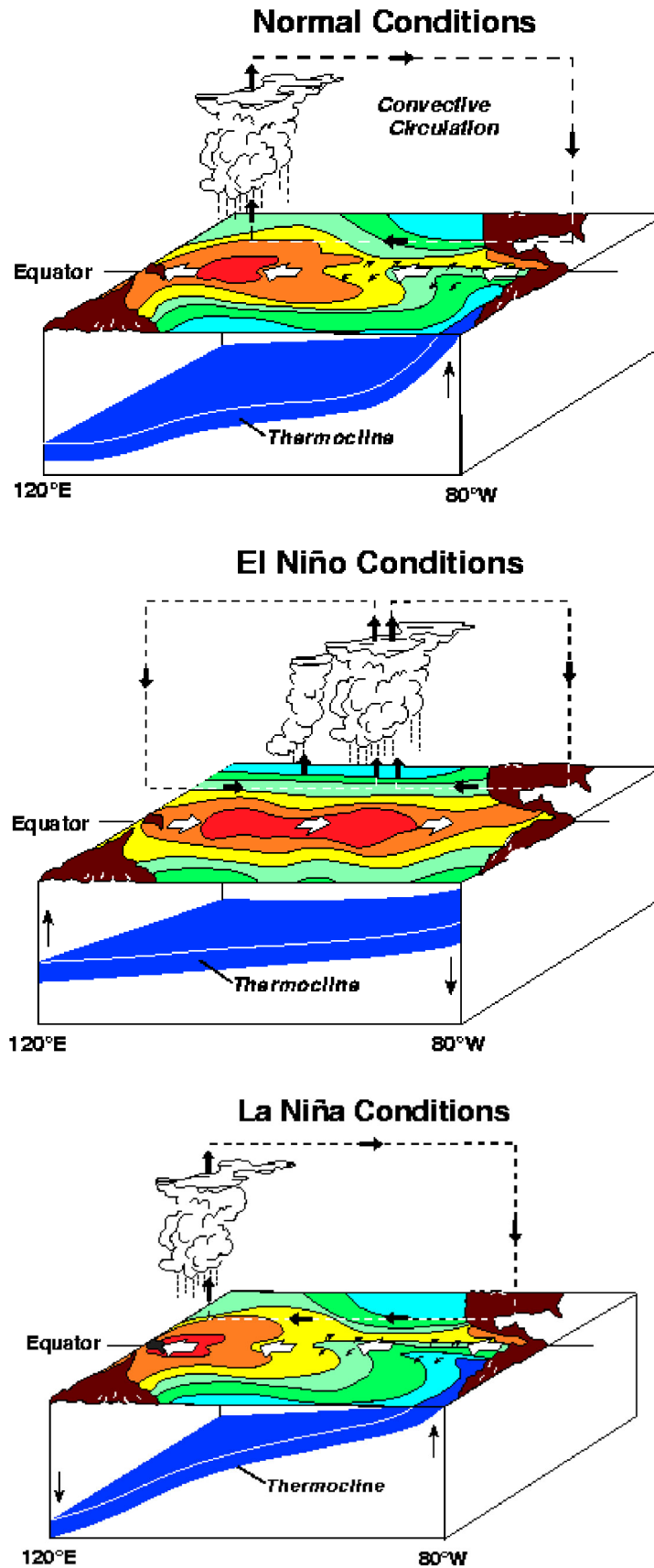
El Niño Conditions

La Niña Conditions

Fig. 1. (a) Ocean-atmosphere conditions in the tropical Pacific during normal conditions. (b) Conditions during El Niño. (c) Conditions during La Niña.

SST contrast between the eastern and western equatorial Pacific that drives the Walker circulation diminishes considerably and gives rise to westerly wind anomalies. These anomalous winds have a considerable impact on the upper layer thermal structure of the ocean (above the thermocline) via local mixing processes, air-sea heat exchange and wind-driven dynamical processes such as equatorial waves. The latter can very rapidly (in a matter of months) transfer information about anomalous oceanic conditions to regions far away from the forcing region, flattening the thermocline along the entire equatorial Pacific, which eventually becomes deeper in the east and shallower in the west. The key to understanding these processes is a strong air-sea interaction mechanism, which provides a positive feedback between atmospheric and oceanic anomalies and can lead to instability of the climatological state. These feedback mechanisms are discussed in the next section.

El Niño is now understood to be the warm phase of an irregular cycle of a coupled ocean-atmosphere mode of climate variability. The cold phase of this mode, named La Niña is, in a broad sense, a sign-reversed El Niño, with some asymmetry probably introduced by the nonlinearity of the system. Figure (1c) depicts the oceanic and atmospheric conditions during a cold La Niña event. The features that characterize the normal conditions (Figure 1a) are strengthened: There is an increased SST contrast between eastern and western regions with stronger than normal easterly winds. The slope of the thermocline increases, becoming shallower in the east and deeper in the west, which brings colder water to the surface in the east and concentrates the warmest surface waters farther to the west. The complexity of the coupled ocean-atmosphere processes involved in these phenomena do not allow the establishment of clear cause-effect relationships.

In the last 20 years considerable progress has been made in observing, explaining and even predicting this phenomena. Nonetheless, many fundamental features of the ENSO cycle are still poorly understood, and many ideas and theories have been proposed to explain them. Some of the key questions that remain are:

- Why is the mean period of ENSO about 4 years?

- What are the dominant feedback mechanisms that control the oscillations?

- What causes the irregularity in the cycle? is it due to chaotic processes intrinsic to the system? or is the intermittency forced by external stochastic processes (weather) unrelated to the ENSO mechanism?

- Is ENSO a self-sustained chaotic oscillation or a damped one, requiring external stochastic forcing to be excited?

- Why and how is ENSO phase-locked to the seasonal cycle?

Now follows a very brief and schematic description of some observations and the intermediate models that have been used to explain them. From these intermediate models one can derive (e.g. Suarez and Schopf, 1988; McCreary and Anderson, 1991; Jin, 1997; Weisberg and Wang, 1997; Picaut *et al.*, 1997) even more simplified models. These "toy models" are believed to contain the basic physics required to explain the observations and the behavior of more sophisticated numerical models of ENSO.

3. AIR-SEA COUPLING: THE BJERKNES HYPOTHESIS

Bjerknes (1969) was the first to realize the strong coupling between atmospheric and oceanic phenomena in the tropical Pacific, and the positive feedback mechanisms that take place among the main ENSO players. His ideas are the basis of our current understanding of ENSO, further developed among others by Gill (1980), Rasmusson and Carpenter (1982), Rasmusson and Wallace (1983), McCreary (1983), Wyrtki (1975, 1986), Anderson and McCreary (1985), Cane and Zebiak (1985), Zebiak and Cane (1987), Graham and White (1988), to mention just a few. The Bjerknes coupling paradigm goes as follows: consider (for the sake of argument) a warm SST anomaly in the central or eastern Pacific, which reduces the zonal SST gradient, weakens the Walker circulation and therefore generates weaker Trade winds or westerly wind anomalies. These anomalies result in a deeper thermocline, which in turn induces positive SST anomalies in the cold tongue region, since the upwelled water is warmer than normal. This reduces even more the zonal SST contrast, which weakens the trades, etc., and so on. This positive feedback mechanism can lead to an instability of the coupled system (Hirst, 1986). ENSO is nonetheless an irregular oscillation, and a negative feedback mechanism must exist to return the system either to normal conditions or, further, to its cold phase. This negative feedback is provided by oceanic subsurface (thermocline) adjustment processes.

Figures 2, 3 and 4 provide interesting clues about the ENSO phenomenon. Figure 2 shows time series of two indices that are commonly used to identify it, and demonstrate quite clearly some of its characteristics. The top curve is the Southern Oscillation Index (SOI), the difference in sea-level pressure anomalies between eastern and western Pacific (Tahiti and Darwin, Australia). Negative (positive) values indicate El Niño (La Niña) conditions, i.e. weakening (strengthening) of the Trades. The second curve is a time-series of SST anomalies in the eastern equatorial Pacific. There is an obvious (negative) correlation between the two series, a clear indication that ENSO is a coupled ocean-atmosphere mode of climate variability. There is an evident irregularity in the period between positive and/or negative extrema in both series and also in their duration.

Southern Oscillation Index and TAO/TRITON SST Anomaly Time Series
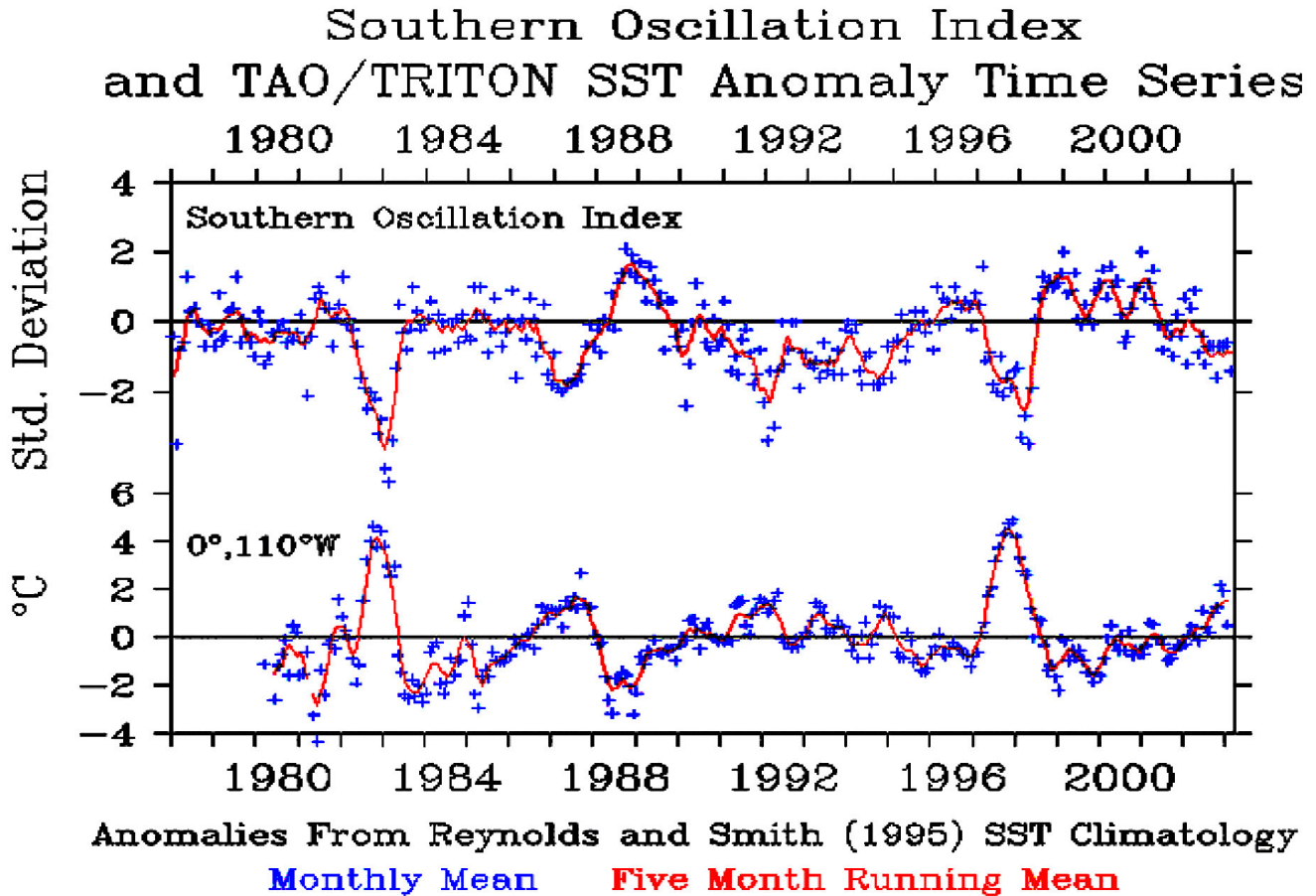
Fig. 2. Time series of the Southern Oscillation Index (SOI) (top panel) and eastern equatorial SST's (lower panel). Negative SOI values indicate El Niño conditions; positive values are associated with La Niña.

Figure 3 gives additional details about the spatial characteristics of ENSO processes; it depicts longitude-time plots of three key ENSO elements along the equator: zonal wind anomalies, SST anomalies and thermocline-depth anomalies. Yellow and red indicate El Niño conditions. Blue contours La Niña conditions. A contour slope different from zero, is an indication of zonal propagation, whereas zero slopes indicate standing variations. The middle panel shows the SST anomalies of the central-eastern equatorial Pacific. Notice the irregularity in ENSO variability, and also that most contours have zero slope: eastern SST anomalies show a pattern consistent with a standing oscillation. By contrast, the contours of thermocline-depth anomalies (right hand panel) have a nonzero slope, indicative of zonal propagation and oceanic adjustment processes. Zonal propagation of wind anomalies are also evident in the left-hand panel. Relatively simple physical models of the tropical ocean atmosphere system can explain the connection between these anomalies. It is known (e.g.: Anderson and McCreary, 1985;

Cane and Zebiak, 1985; Battisti and Hirst, 1989; Neelin *et al*., 1998), that SST anomalies in the eastern Pacific are very much determined by the thermocline depth: climatological upwelling brings warmer or colder water to the surface depending on whether the thermocline is deep or shallow, although advection may be locally important. In turn, the thermocline depth anomalies are to a large extent remotely generated by wind anomalies in the western Pacific. A circular argument arises when one tries to determine the origin of the anomaly in one field, an indication of the coupled nature of these phenomena.

Another important characteristic feature of ENSO is its phase locking to the seasonal cycle. Figure 4 shows seven time-series of SST anomalies in the eastern Pacific from the Niño3 region (a rectangular area that runs from 5°N-5°S, and from 90°W-150°E), corresponding to different El Niño events. The horizontal axis begins in July of the year prior to a warm ENSO event. The label yr0 is the year in

Monthly Zonal Wind, SST, and 20°C Isotherm Depth Anomalies 2°S to 2°N Average

Zonal Wind (m s⁻¹)　　　SST (°C)　　　20°C Isotherm Depth (m)
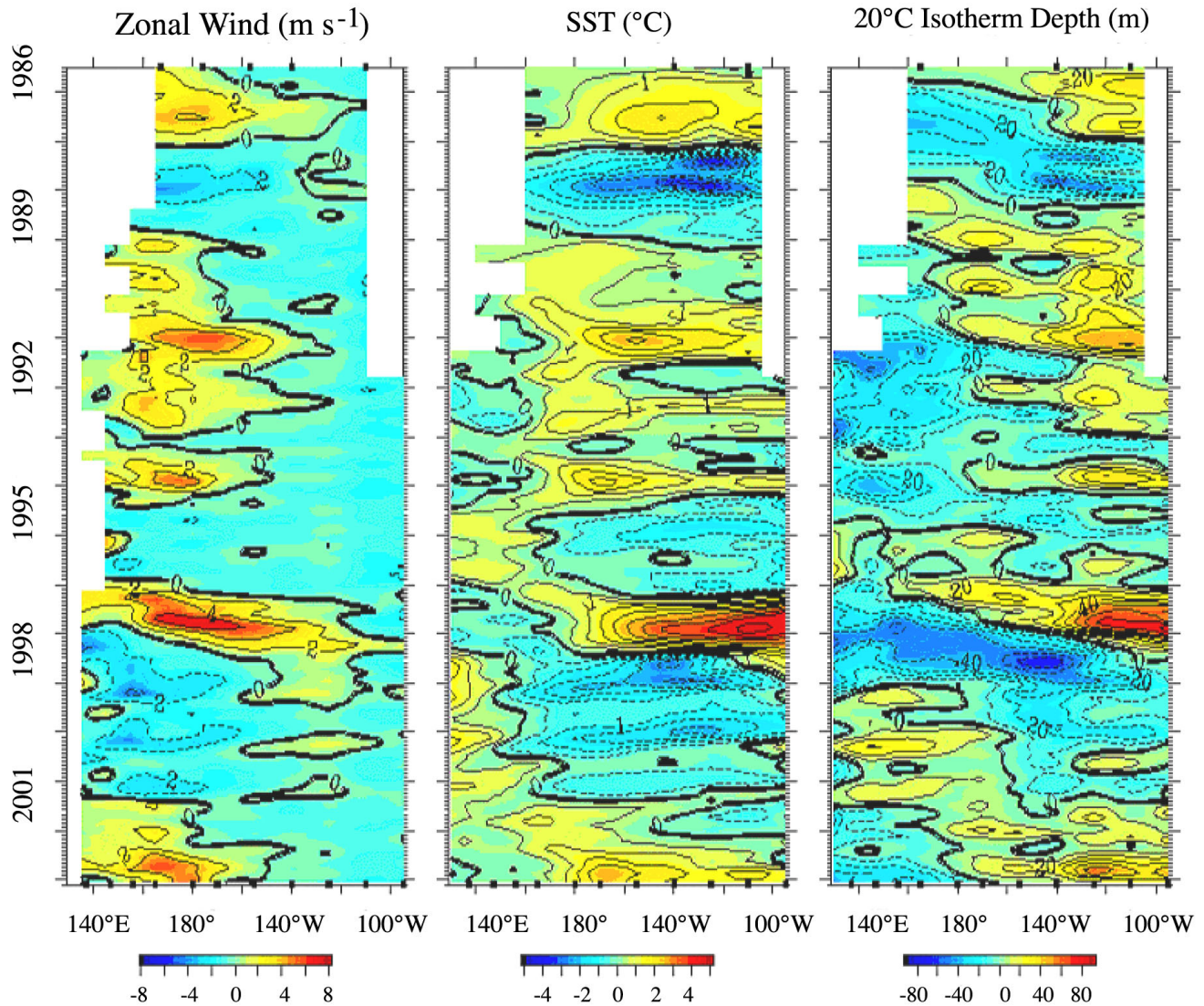
Fig. 3. Longitude-time plots along the Equator of a) zonal wind anomalies, b) SST anomalies and c) thermocline depth anomalies.

which the event starts. With one exception, all events reach their maximum in the boreal winter between December/February of yr0/yr+1. This is a clear indication of the close relation between ENSO and the seasonal cycle. Tziperman *et. al*. (1994), Chang *et al*. (1995) and Galanti and Tziperman (2000) have developed theories to explain this behavior.

4. MODELS DEVELOPED TO UNDERSTAND ENSO

A wide spectrum of models has been used to understand and predict ENSO. They range from very sophisticated coupled general circulation models (GCM) of the ocean and atmosphere, to very simple models (known as "toy models")

that attempt to extract the very basic physics of the phenomena, based on educated guesses or limits of the ocean-atmosphere interaction process. Statistical models based only on data have also been developed with success, but here they will be mentioned only briefly.

A set of intermediate models appear to capture the essential features of ENSO (Cane and Zebiak, 1985; Anderson and McCreary, 1985; Battisti and Hirst, 1989; Jin and Neelin, 1993); the "toy models" are reduced versions of these intermediate ones. The paradigm of the intermediate ocean models is a shallow-water reduced-gravity layer model, in which only the top layer is dynamically active (also known as a $1\frac{1}{2}$

layer model). The layer interface between the upper and the deep ocean represents the thermocline, with the deep ocean assumed motionless and infinitely deep. Since SST is a key player of ENSO, a thermodynamic equation is also included, in which temperature variations are determined by mixing, advection and surface heat fluxes. Temperature variations affect neither the pressure gradients nor the dynamics of the system. These models are also formulated as anomaly models, that is, basic state conditions of all variables are given (e.g. a seasonal cycle) and so the variables only represent anomalies with respect to the known basic state (Zebiak and Cane, 1987; Battisti and Hirst, 1989; Jin and Neelin, 1993).

A schematic of the $1\frac{1}{2}$ layer model is shown in Figure 5 and described more thoroughly in the next section.

4.1 The Ocean

Starting from the momentum, volume conservation and heat equations, the following linearized equations for the layer model can be derived (see Gill, 1982; Philander *et al.*, 1984):

$$\frac{\partial u}{\partial t} - \beta y v = -g'\frac{\partial h}{\partial x} - \varepsilon u + \frac{\tau_x}{\rho_0 H}$$

$$\frac{\partial v}{\partial t} + \beta y u = -g'\frac{\partial h}{\partial y} - \varepsilon v + \frac{\tau_y}{\rho_0 H} \qquad (1a,b,c)$$

$$\frac{\partial h}{\partial t} + H(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}) = -\varepsilon h,$$

where h is the layer depth (of mean value H); u and v are the zonal (x) and meridional (y) velocities, β is the derivative of the Coriolis parameter with respect to latitude, τ_x and τ_y are the zonal and meridional wind-stresses, g' is de reduced gravity and ε is a damping coefficient for a crude representation of momentum and heat mixing. The layer temperature equation (heat equation) takes the form

$$\frac{\partial T}{\partial t} + u\frac{\partial T}{\partial x} + v\frac{\partial T}{\partial y} + Hev(w)w(T - T_{sub}) = -\varepsilon_T T + Q, \quad (1d)$$
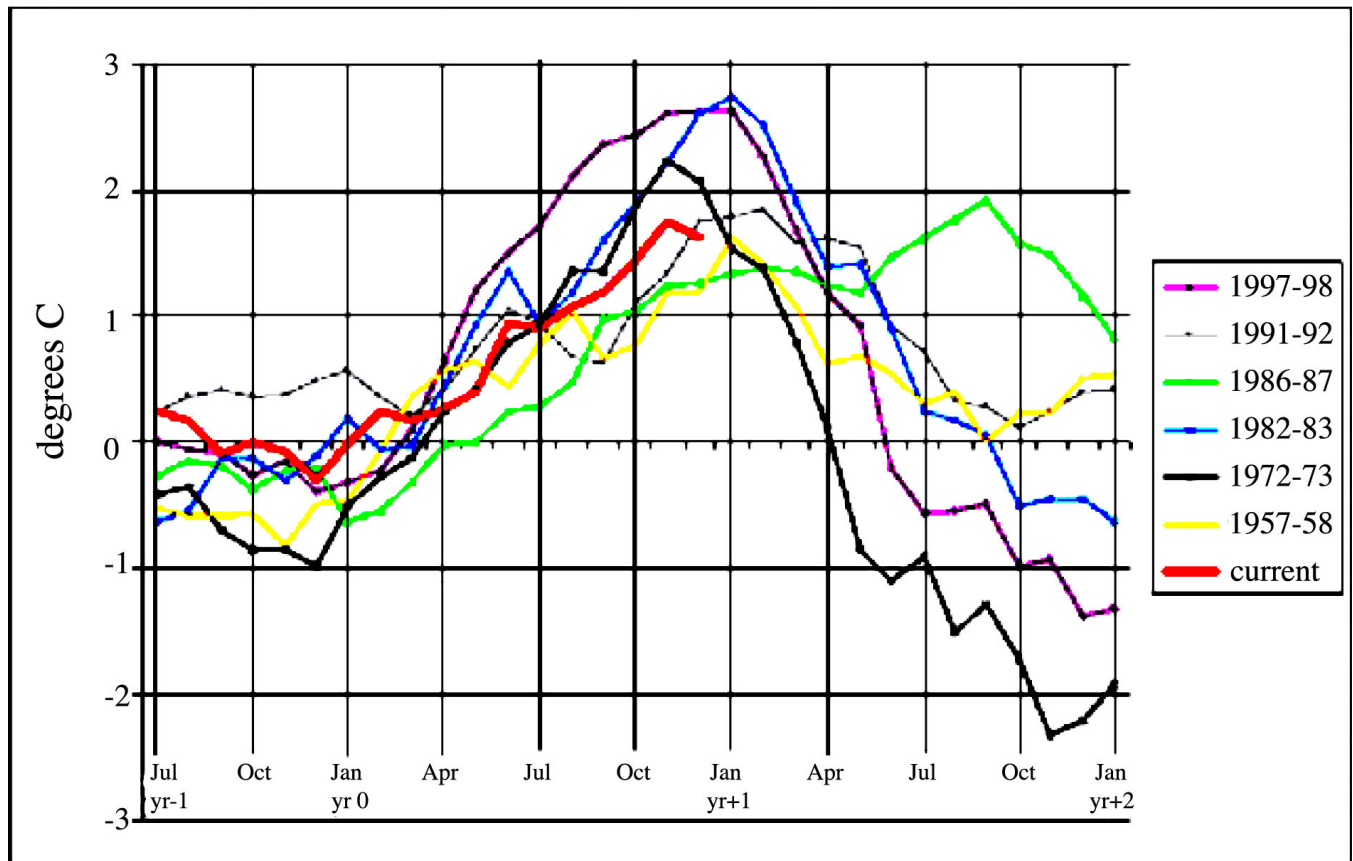

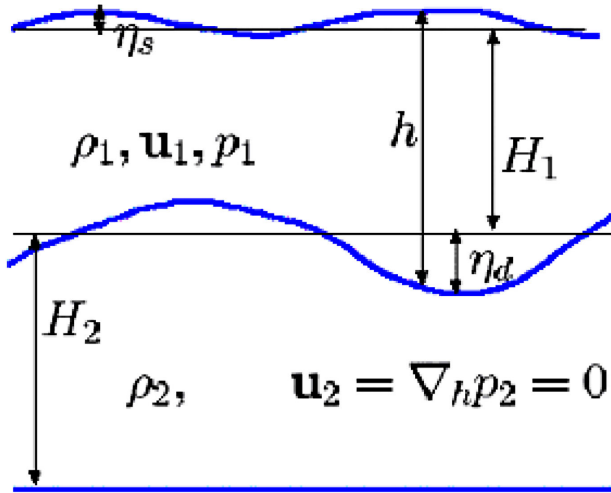Fig. 4. Time evolution of SST anomaly time-series from Niño3 region for seven ENSO events.

Fig. 5. Characteristics of the reduced-gravity ocean model.

where Q is the surface heat-flux, w is the vertical velocity, Hev(w) is a step function, T_{sub} is the subsurface temperature and ε_t parameterizes mixing.

4.2 The Atmosphere

The atmospheric model was developed by Gill (1980, 1982), and represents a "first baroclinic mode" atmosphere whose equations are given by:

$$-\beta y V = -\frac{\partial \Theta}{\partial x} - \varepsilon_a U$$

$$+\beta y U = -\frac{\partial \Theta}{\partial y} - \varepsilon_a V$$

$$c_a^2 \left(\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} \right) = -\varepsilon_a \Theta + Q \ ,$$

(2 a,b,c)

where *U,V,* stand for the zonal and meridional surface winds, Θ is the geopotential (Gill, 1982), c_a is the phase speed of gravity waves in the atmosphere, β is the meridional variation of the Coriolis parameter and ε_a is a damping parameter. To connect the atmosphere model with the ocean, the heating term in the equations (Q) can be parameterized to be proportional to the SST:

$$Q = \alpha_T T.$$

(3)

Note that the atmospheric equations are diagnostic, i.e. there are no time-derivatives. This is because the time-scale for atmospheric adjustment (a few days) is much shorter than that of the ocean (months). Recall also that in both oceanic and atmospheric models the variables represent anomalies with respect to a known, pre-defined mean state.

4.3 Further approximations

Additional approximations can be made to the dynamical ocean model; see Tziperman (2001) for a discussion. Dimensional analysis based on scales of variability of the system (e.g. on ENSO time-scales (i.e. months), zonal scales are much larger than meridional variations, and meridional wind-stress is also less important than the zonal component) indicate that the following "long wave" approximation to system (1) can be used:

$$\frac{\partial u}{\partial t} - \beta y v + g \frac{\partial h}{\partial x} = -\varepsilon_m u + \frac{\tau_x}{\rho H},$$

$$\beta y u + g \frac{\partial h}{\partial y} = 0,$$

$$\frac{\partial h}{\partial t} + H \left[\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right] = -\varepsilon_m h.$$

(4)

Eliminating *u* and *v* from (4), a single equation for *h* may be obtained

$$\beta y^2 (\partial_t + \varepsilon_m) h + \frac{g H}{\beta} \left[\frac{2}{y} \partial_y - \partial_{yy} \right] (\partial_t + \varepsilon_m) h - g H \partial_x h + \frac{1}{\rho} (\tau_x - y \partial_y \tau_x) = 0.$$

(5)

The thermodynamic equation (1d) can also be simplified:

$$\partial_t T = -\varepsilon_T T - \gamma \frac{\overline{w}}{H_1} (T - T_{sub}(h)),$$

(6)

where T_h is the temperature anomaly at some specified constant depth H_1, and is a function of the thermocline depth anomaly h. The parameter $0 < \gamma < 1$ relates the temperature anomalies entrained into the surface layer to the non-local deeper temperature variations due to T_h. $\overline{w}$ is the mean vertical velocity and ε_T is a damping term.

From the atmospheric model (Gill's model, equations 2) the following non-local relation between the zonal wind stress and equatorial SST can be derived (Jin and Neelin, 1993; Jin, 1997; Tziperman, 2001):

$$\tau_x(x,y) = \mu A(T_e,x,y).$$

(7)

In this last formula, A(T_e,x,y) is a non-local function that relates the equatorial SST (T_e) to the zonal wind stress τ_x, and μ serves as a relative coupling coefficient. Equations 4 (or 5), 6 and 7 represent a coupled ocean-atmosphere model. These equations are further simplified in the construction of the conceptual "toy models" discussed later.

4.4 Equatorial waves

At the heart of the dynamical adjustment of the equatorial ocean and atmosphere are the wave motions supported by the previous equations. Of particular importance to ENSO are the long oceanic equatorial waves (Gill, 1982; Philander *et al.,* 1984) for which a clear distinction can be made between an eastern propagating wave (Kelvin wave) and a set of western propagating Rossby waves. The phase speed of the gravest Rossby wave is about 1/3 of the speed of a Kelvin wave (which is close to 3 m/sec), which can traverse the Pacific in 2-3 months. These waves can be forced (i.e. directly generated by wind perturbations) or can be generated by wave reflection of other waves at the solid boundaries of the basin, and are then labeled free waves. Kelvin and Rossby waves can be clearly identified in thermocline depth perturbations (TAO array, PMEL, http://www.pmel.noaa.gov/tao) and also imprint their signal in sea-level variations (See AVISO web pages on observing ENSO from space, http://www-aviso.cls.fr). These waves carry the information from earlier perturbations of the ocean-atmosphere system in the Pacific and are of paramount importance to explain ENSO.

4.5 Ocean response to a wind-stress anomaly

The fact that Kelvin waves can travel only to the east and long Rossby waves can travel only to the west, has a tremendous importance in the wave reflection problem, because a Kelvin wave impinging on a eastern boundary can only reflect its energy as a set of Rossby waves, whereas Rossby waves impinging on a western boundary can only reflect as Kelvin waves. Figure 6 shows an idealized numerical experiment using equations (1) (Philander *et al.,* 1984; figures taken from http://iri.columbia.edu/climate/ENSO/theory/ in which a westerly wind perturbation of finite geographical and temporal extent (Figure 6a), at the center of the equator, is applied to an ocean in a state of rest. The sequence shows the ocean response to this forcing. Figure 6b shows a downwelling Kelvin wave (red contours) propagating to the east and an upwelling Rossby wave with two off-equatorial cyclones. By day 50 (Figure 6c) the Kelvin wave has reached the eastern boundary and its reflection as a downwelling Rossby wave can be clearly seen at day 75 (Figure 6d). After 100 days (Figure 6e) the Rossby wave reaches the western boundary, and its reflected upwelling Kelvin wave can be identified after 125 days along the equator. By day 175 (Figure 6g) this upwelling Kelvin wave has reached the eastern boundary and an upwelling Rossby wave-train is again generated by its reflection. After 275 days (Figure 6h) one can still identify the tail of the original (forced) Rossby wave, its reflected Kelvin wave and the Rossby wave generated by the reflection of the initial forced Kelvin wave. The reflection off a western boundary provides the negative feedback mechanism for the de-
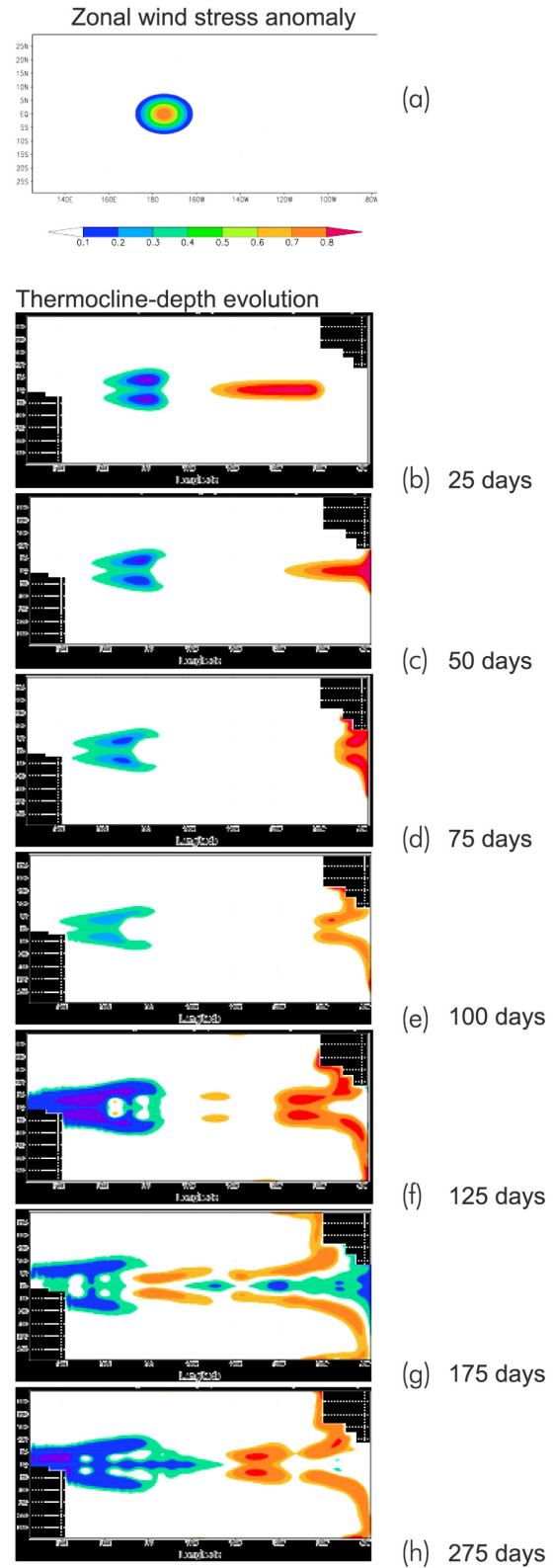

Fig. 6. Ocean adjustment sequence to wind perturbations. (a) Zonal wind-stress anomaly. (b) to (h) Evolution in time of the ocean thermocline in a reduced gravity model in response to the wind anomaly of panel (a), (see text for details). Figure obtained from the International Research Institute of Climate Change http://iri.columbia.edu/climate/ENSO/theory/.

mise of ENSO events, in one of the first and very successful toy models of ENSO: The delayed action oscillator.

This equatorial-ocean adjustment time-scale is smaller than the period of the ENSO cycle, but it is the ocean-atmosphere coupling that gives rise to the 2-7 year period time-scale for ENSO. The coupling also provides a positive feedback that can lead to instability of the coupled system (e.g. via Bjerknes ideas). The memory of the system is nevertheless governed by upper-ocean adjustment processes which provide the means (negative feedback) to bring the system back to normal or to its opposite phase. Two limits have been explored to understand the air-sea interaction process:

(a) FAST WAVE LIMIT: Time-derivatives in the dynamics represented in equations (1a,b) are set to zero, based on assumptions that the time-scale of adjustment by waves is shorter than the ENSO time-scale. The unstable modes that result from this limit are controlled by the SST adjustment time to surface processes, wind and thermocline anomalies. The modes that result from this limit are not very realistic, but explain the behavior of some numerical models, and is a useful limit for understanding the ENSO mode.

(b) FAST SST LIMIT. Here, the time derivative in equation 1d is set to zero, so the SST equation becomes diagnostic. This means that the adjustment of SST is much faster than the dynamic adjustment. The unstable and damped coupled modes that result from this limit are more realistic, and are the basis of the very successful delayed oscillator model discussed below.

Recent studies (Neelin and Jin, 1993; Neelin *et al*., 1994; Jin, 1997) suggest that the ENSO mode is really a mixed mode instability that contains features of the two limits discussed above. A thorough study of the parameter space (e.g. Neelin *et al*., 1994) indicates that the transition from fast SST to fast wave limits is continuous, and sensible parameter values (e.g. air-sea coupling coefficients) indicate that ENSO is located within both limits in parameter space, i.e., it is a mixed mode that contains features of the fast SST and wave limits.

4.6 Toy models

Several very simple and heuristic models have been put forward to explain the ENSO mechanism. In these, positive and negative feedbacks in the coupled system are represented by ad-hoc coefficients and simple parameteriza-tions, although they all can be deduced by taking appropriate limits or neglecting some of the physical mechanisms present in the intermediate models described by equations 1 to 6. We discuss the classic delayed action oscillator and some of its

variants, as well as the recent recharge-discharge oscillator proposed by Jin (1997).

4.6.1 The Delayed Action Oscillator

This simple model was first proposed by Schopff and Suarez (1988), and later used by many authors to explain the results of many intermediate models (Battisti and Hirst, 1989; Anderson and McCreary, 1985; see Neelin *et al*., 1998 for a review). A typical form of the delayed oscillator equation is

$$\frac{dT}{dt} = AT - B(T - \delta) - \varepsilon T^3,$$

where T are eastern equatorial SST anomalies (although a similar equation can be obtained using other variables such as thermocline-depth anomalies). The first term on the right represents the positive feedback mechanisms from air-sea coupling that generate an instability. The third term represents negative feedback mechanisms that limit perturbation growth from the nonlinearity in the system. The middle term is a delayed negative feedback provided by ocean adjustment processes. The delay time-scale δ represents the time taken by forced upwelling Rossby waves to reach the western boundary, reflect as a Kelvin wave and reach the eastern Pacific. This simple model can be derived from the intermediate model if one neglects the impact of the eastern boundary reflection. Figure 7 depicts the basic mechanism of the delayed oscillator (taken from Tzipermann, 2001):

A weakening of the Trades (1) generates a downwelling Kelvin wave (2) that warms SST's, which amplify the wind anomaly and an instability of the air-ocean system begins. At the same time, off-equatorial upwelling Rossby waves (3), arise from changes in the wind stress curl there; they travel to the western Pacific and reflect as upwelling Kelvin waves that shut down the instability on arrival to the eastern boundary. The same mechanism explains the start and end of cold events. The delayed oscillator mechanism explains the results of some models and observed ENSO events, but the parameter range in which realistic oscillations occur is relatively limited: there seem to be other important players in the story. To produce irregular oscillations with this model one can either add stochastic forcing or include forcing related to the seasonal cycle, which produces a chaotic behavior from nonlinear resonance between ENSO timescales and the seasonal cycle.

4.6.2 Western Pacific Oscillator (Weisberg and Wang, 1997)

During the early1990's a discussion arose on whether reflection of signals at the discontinuous western boundary was warranted (Battisti, 1989; Graham and White, 1991; White and Tai, 1992) and responsible of the negative-feedback mechanism for ENSO. Delayed oscillator models in-
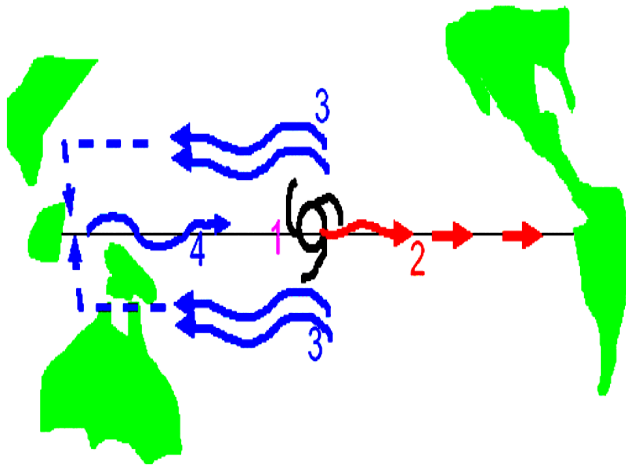
Fig. 7. The delayed-action oscillator model of ENSO (taken from Tziperman, 2001), see text for details.

feedback to produce an oscillatory ENSO mode. This mechanism is depicted in Figure 8 (taken from Wang, 2001). Westerly wind anomalies in the central equatorial Pacific (Niño4 region) generate downwelling Kelvin waves and upwelling off-equatorial Rossby waves. The upwelling signal of the latter, in due course, generates negative SST anomalies off the western equatorial Pacific (Niño6 region) which give rise to positive sea level pressure anomalies, which in turn, generate easterly (westward) wind anomalies along the equator. These anomalies produce upwelling Kelvin waves that provide the negative feedback to shut down the eastern equatorial instability started by the westerly wind anomalies in the Niño4 region. This means that reflection of signals off the western boundary are not essential to produce an oscillation. A delayed oscillator model can be derived from this assumptions, so this model contains essentially the same physics, i.e. the fast SST limit of the instability.

troduced partial reflection coefficients to account for this. But Weisberg and Wang (1997) argued that by looking carefully to observations one could notice that, although smaller, anomalies in the western Pacific could provide the negative

4.6.3 Advective-Reflective Oscillator (Picaut *et al.*, 1997)

The model proposed by Picaut *et al.* (1997), emphasizes the role of advection produced by reflection of signals

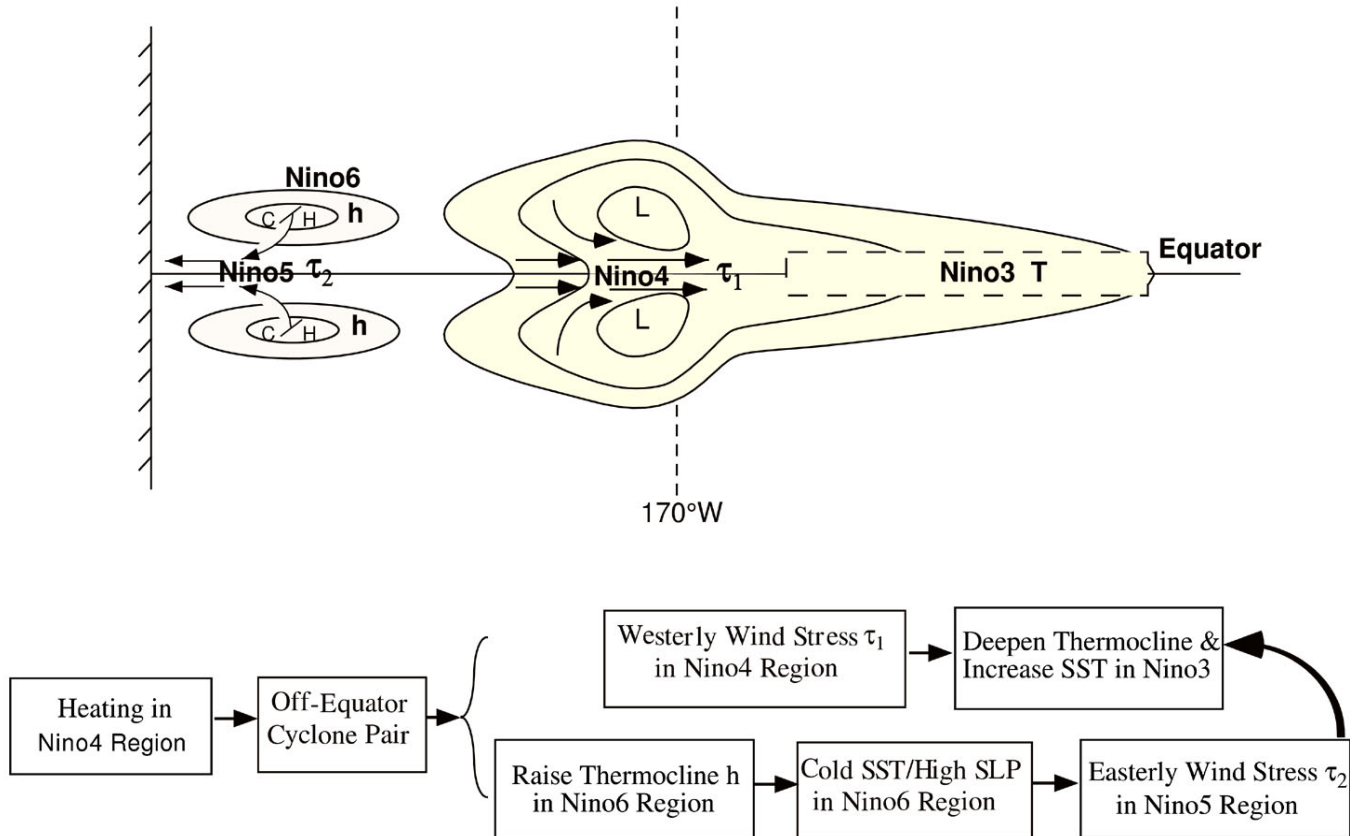## The Western Pacific Oscillator

Fig. 8. The western Pacific Oscillator mechanism (see text for details). Taken from Wang (2002).

at both western and eastern boundaries as the negative feedback mechanism to produce an ENSO mode. It is argued that upwelling Kelvin waves (that result from the reflection of off-equatorial Rossby waves generated by westerly anomalies in the central Pacific) and downwelling Rossby waves generated by reflection of Kelvin waves at the eastern boundary, both have westward anomalous currents, which together with the mean zonal currents, return the warm SST's to the west. It is unclear how to express these assumptions in a simple model, but results from a numerical model have been interpreted using these ideas. Figure 9 illustrates the mechanism.

4.6.4 Recharge-Discharge Oscillator

The Recharge-Discharge oscillator mechanism proposed by Jin (1997) is basically a simple model that represents the mixed nature of the ENSO mode, proposed by Jin and Neelin (1993). The physics of the mode are interpreted in terms of ideas proposed by Wyrtki (1986) and early interpretations of results from the model of Cane and Zebiak (1985), who argued the ENSO oscillation was related to a
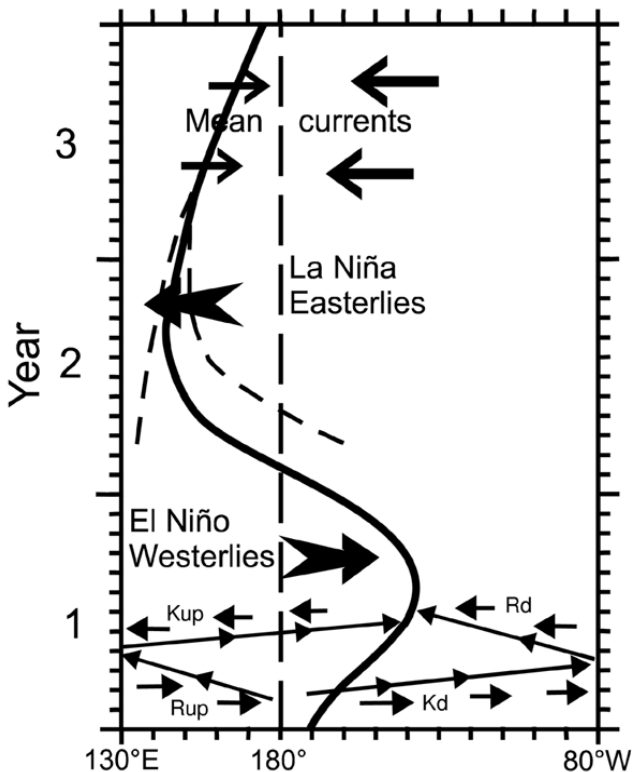
The Advective-Reflective Oscillator

Fig. 9. The advective-reflective oscillator mechanism of Picaut *et al*. (1997).

build-up of equatorial "heat content" (recharge phase) and its discharge phase (El Niño events) in which heat was exported to off-equatorial regions. The equations of the model are the following:

$$\frac{dT_E}{dt} = CT_E + Dh_w - åT_E^3$$

$$\frac{dh_w}{dt} = -ET_E - R_h h_w,$$

where T_E represents eastern equatorial SST and h_w the western equatorial thermocline depth. The model does not need the explicit inclusion of a time-delay parameter, so the details of the subsurface oceanic adjustment don't need to be considered explicitly. However Galanti and Tziperman (2000) show that one can reformulate the problem including wave propagation and explicit time delays. The model incorporates two time-scales, one associated with subsurface adjustment processes and another related to SST adjustment, which represent the mixed-mode nature of the ENSO mode. The delayed oscillator model can be obtained from these equations if one neglects the SST adjustment time-scale (i.e the SST adjustment occurs instantaneously). Figure 10 illustrates the basic mechanism of the recharge-discharge oscillator: During the warm phase, westerly wind anomalies that produce warm SST in the east Pacific and off-equatorial Rossby waves produce a subsurface poleward Sverdrup transport that exports heat to off-equatorial regions. At the end of this warm phase, the equatorial thermocline is on average shallower than normal across the equatorial Pacific. This anomalous thermocline produces cold SST's in the east, which initiate the cold phase, and wind anomalies that produce a meridional Sverdrup transport that replenishes the equatorial heat content, and in the end, gives rise to an equatorial thermocline which is deeper than normal. This generates warm anomalies in the east, which produce westerly wind anomalies, and the cycle repeats.

Wang (2001) has developed a unified oscillator model that includes all of the previous models as limits. His model consists of four equations and is an extension of the western Pacific oscillator discussed above.

5. ENSO'S IRREGULARITY

All of the simple models discussed above produce unstable, damped modes and regular oscillations whose frequency varies with the coupling strength. The irregularity of ENSO is one of its basic characteristics, and there is much debate regarding what is the cause of this irregular behavior. Tziperman *et al*. (1994) suggested that the irregularity is due to nonlinear interaction of ENSO phenomena with the seasonal cycle. Using a seasonally-forced delayed action oscil-

The Recharge-Discharge Oscillator

Warm Phase

a

τ SST(+)

h

0

Warm to Cold Transition Phase

b

$\tau \sim 0$ SST~0

h

0

Cold to Warm Transition Phase

d

$\tau \sim 0$ SST-0

h

0

Cold Phase
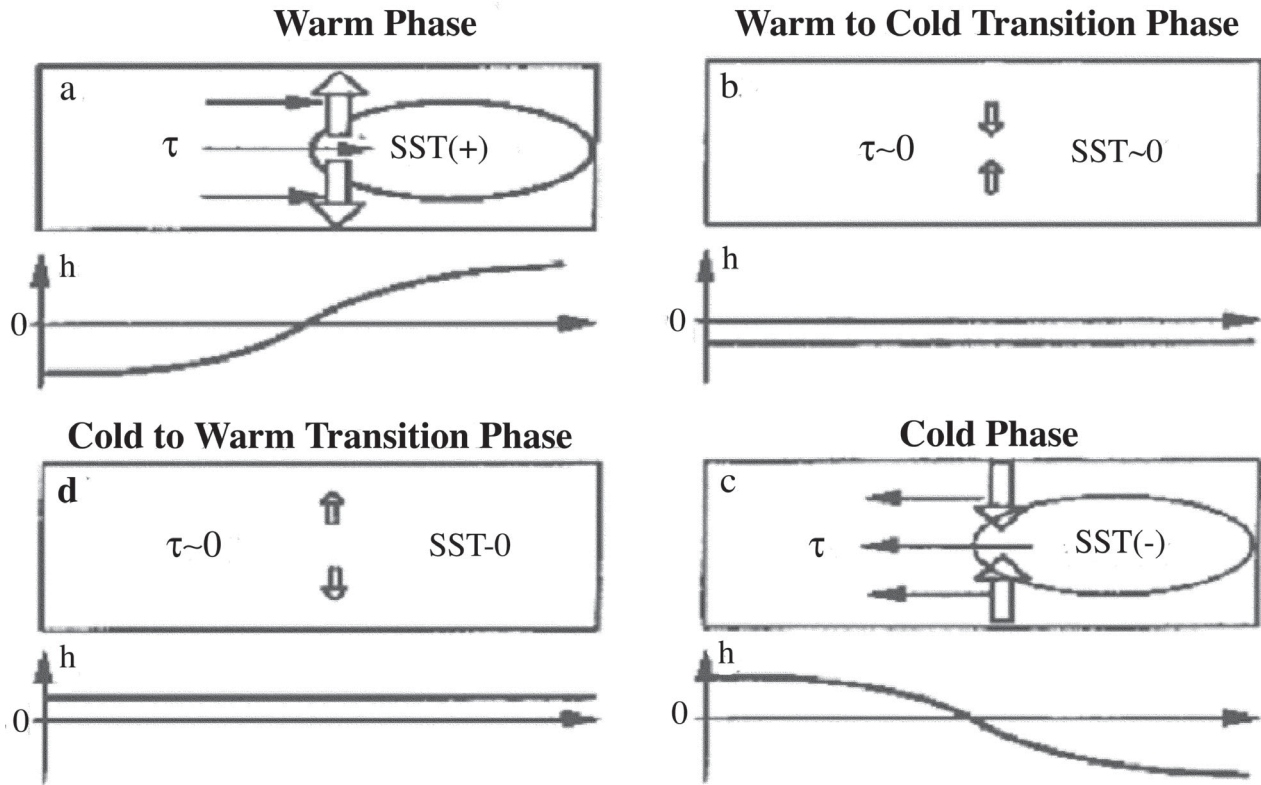
c

τ SST(-)

h

0

Fig. 10. The Recharge-Discharge mechanism for ENSO (Jin, 1997).

lator model, they argued that the irregularity was caused by nonlinear resonance and chaotic behavior of the system, i.e., deterministic chaos intrinsic to the system is what causes the irregular cycle. Others (e.g. Moore and Kleeman, 1999) show that inclusion of stochastic forcing into the models, which intends to represent "weather" or processes external to the system, also can cause irregularity. In a somewhat more drastic statement, Penland and Magorian (1993) and Penland and Sardeshmukh (1995), argued that ENSO is a damped mode of the coupled system, which is excited by stochastic forcing. Using ideas related to optimal perturbation growth and non-normality of the linearized system they showed how particular structures at initial time could lead to ENSO type anomalies later. An example is shown in Figure 11.

Whether ENSO is a damped or unstable oscillation excited by stochastic forcing, or a chaotic system whose irregularity comes from within, has implications for its predictability, so the issue is not merely academic.

Using data from the TAO array, Kessler (2002) suggests that the phase relationship between eastern equatorial SST and zonal mean thermocline depth is consistent with the recharge-discharge model of Jin (1997) but that ENSO is not really a self-sustained oscillation: whilst warm events seem to contain the seeds for their termination and transition to cold events, the latter apparently do not. The cycle seems to break down, and warm events appear to be generated by external forcing. This suggests that ENSO is closer to a damped or stable mode excited by stochastic forcing. That warm events seem to start and peak during boreal winter months could be explained by considering that this is a period of high sensitivity and/or increased external forcing activity. Galanti and Tziperman (2000) give a different explanation based on wave dynamics and seasonal changes in the coupling, i.e., in terms of processes internal to the system.

5. SUMMARY

Tremendous advances have taken place in the last 30 years in our understanding and prediction of ENSO and its warm and cold phases. Still, some of the basic questions of the phenomena remain obscure. The fact that neither sophisticated, intermediate, simple or statistical models of ENSO were able to predict the rapid transition from El Niño condi-
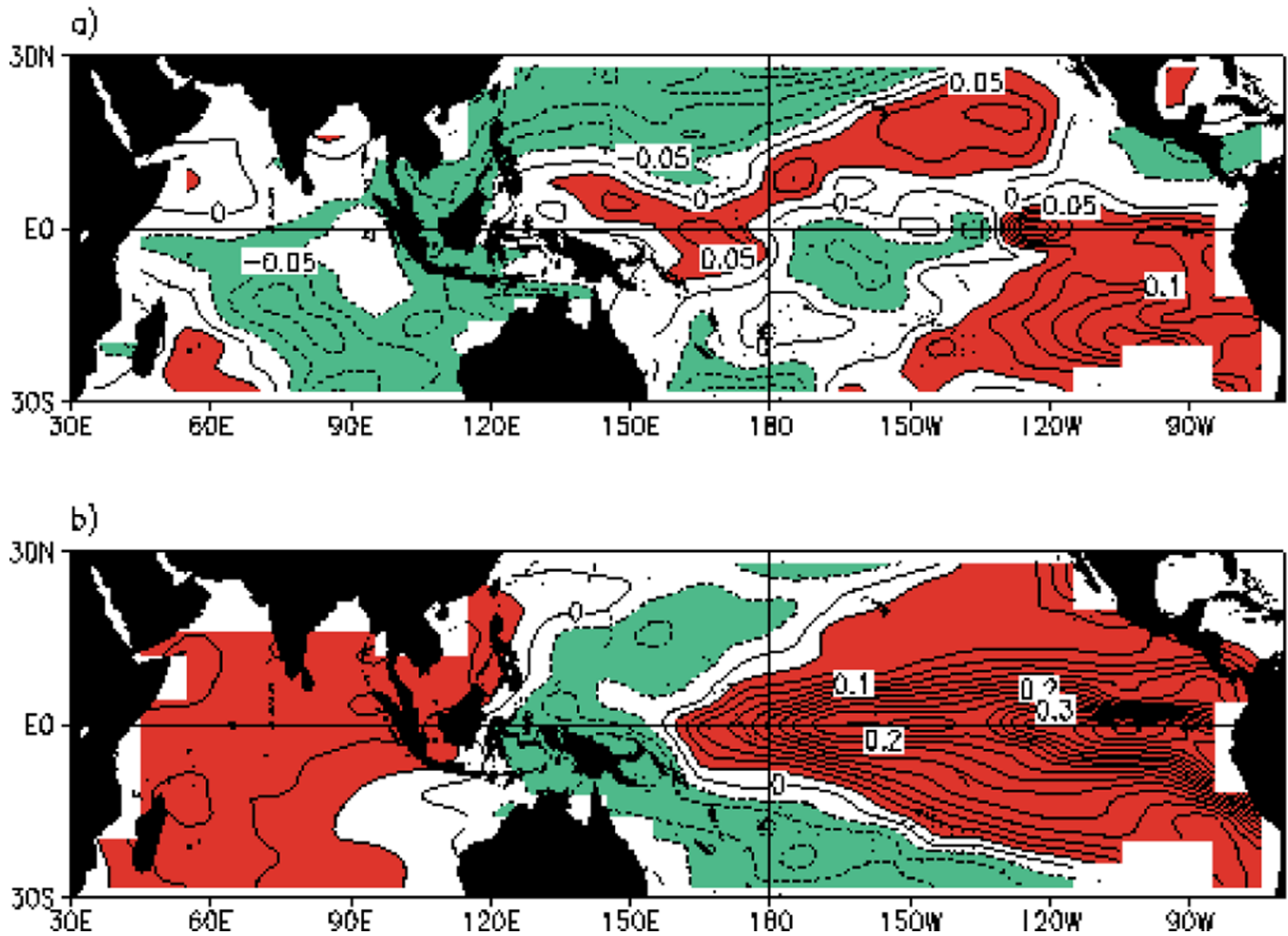
Fig. 11. Stochastic forcing of a linear model (Penland and Magorian, 1993) can give rise to ENSO conditions. Top panel shows optimal initial conditions of the linear inverse model. Bottom panel, El Niño conditions obtained after integration of the model from the optimal initial state.

tions to La Niña conditions in 2003 should serve as an example that there is still much work to be done to really understand the air-sea interaction processes that give rise to ENSO.

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