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Understanding ENSO Physics—A Review

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Since the TOGA program, and in particular the maintenance of its observing system in the tropical Pacific, significant progress has been made in the understanding of ENSO. ENSO has been viewed as a self-sustained and naturally oscillatory mode or a stable mode triggered by stochastic forcing. Whatever the case, El Niño involves Bjerknes' positive ocean-atmosphere feedback that culminates with warm SST anomalies in the equatorial eastern and central Pacific. After an El Niño reaches its mature phase, negative feedbacks are required to terminate the growth of warm SST anomalies. Four major negative feedbacks have been proposed: wave reflection at the ocean western boundary, a discharge process due to Sverdrup transport, a western Pacific wind-forced Kelvin wave of opposite sign, and anomalous zonal advection. These negative feedbacks may work in varying combinations to terminate El Niño, and reverse it into La Niña.

The seasonal cycle can contribute to the irregularity and phase-locking of ENSO, and the intraseasonal variability can be a source of both ENSO's variability and irregularity. Tropical Pacific decadal-multidecadal variability and warming trends may modulate ENSO. Many mechanisms have been proposed to explain tropical Pacific decadal-multidecadal variability, and they are categorized by their tropical origins and tropical-extratropical connections. Mechanisms of tropical origins include stochastic forcing, interactions between the seasonal and interannual cycles, internal nonlinearity, asymmetry between El Niño and La Niña, and local ocean-atmosphere interaction, while those of tropical-extratropical connections involve oceanic bridges, wave propagation, and atmospheric bridges. Difficulties and uncertainties of studies on low-frequency variability and interpretation of warming trends, global warming, and ENSO are also discussed.

1. INTRODUCTION

At the end of the 19th century, the term El Niño was used to denote the annual occurrence of a warm ocean current that flowed southward along the west coast of Peru and Ecuador around Christmas. The Peruvian geographers noted that in some years the onset of warm conditions was stronger than usual and was accompanied by unusual oceanic and climatic phenomena. Starting with the arrival of foreign-based scientific expeditions off Peru in the early 20th century, the concept of referring to these unusual events as El Niño gradually spread through the world's scientific community. It wasn't

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until the 1950s/1960s that scientists realized that El Niño is far more than a coastal phenomenon, and that it is associated with basin-scale warming in the tropical Pacific Ocean. Sir Gilbert Walker in the 1920s and 1930s found that notable climate anomalies occur around the world every few years, associated with what he called the Southern Oscillation [*Walker*, 1923, 1924; *Walker and Bliss*, 1932]. The Southern Oscillation is characterized by an interannual seesaw in tropical sea level pressure (SLP) between the Western and Eastern Hemispheres, consisting of a weakening and strengthening of the easterly trade winds over the tropical Pacific. *Bjerknes* [1969] recognized that there is a close connection between El Niño and the Southern Oscillation (ENSO) and they are two different aspects of the same phenomenon.

Bjerknes hypothesized that a positive ocean-atmosphere feedback involving the Walker circulation is a cause of ENSO. An initial positive sea surface temperature (SST) anomaly in the equatorial eastern Pacific reduces the east-west SST gradient and hence the strength of the Walker circulation [*Gill*, 1980; *Lindzen and Nigam*, 1987], resulting in weaker trade winds around the equator. The weaker trade winds in turn drive the ocean circulation changes that further reinforce SST anomaly. This positive ocean-atmosphere feedback leads the equatorial Pacific to a warm state, i.e., the warm phase of ENSO (El Niño). At that time, Bjerknes did not know what causes a turnabout from a warm phase to a cold phase, which has been recently named La Niña [*Philander*, 1990].

After Bjerknes' seminal work, ENSO was not intensively studied until the 1980s. The intense warm episode of the 1982–83 El Niño, which was not recognized until it was well developed, galvanized the scientific community in an effort to understand and predict ENSO. The 1982–83 El Niño onset was not consistent with the prior "buildup" of sea level in the western Pacific by stronger than normal trade winds prior to 1982, presumed to be a necessary precursor of El Niño [*Wyrtki*, 1975]. Also, there was no warming off the west coast of South America in early 1982, considered to be part of the normal sequence of events characterizing the evolution of El Niño [e.g., *Rasmusson and Carpenter*, 1982]. Building on the earlier efforts of the Equatorial Pacific Ocean Climate Studies (EPOCS) program, this motivated a ten-year international Tropical Ocean-Global Atmosphere (TOGA) program (1985–1994) to study ENSO. TOGA built an ocean observing system in the tropical Pacific Ocean, conducted theoretical and diagnostic studies of the ENSO phenomenon, and developed a hierarchy of coupled ocean-atmosphere models to study and predict ENSO. A special volume of the *Journal of Geophysical Research* (volume 103, June 1998) provided a comprehensive review of observations, theory, modeling, and predictability of ENSO during the TOGA decade (also see ENSO reviews of *Enfield* [1989]; *Philander* [1990]; *McCreary and Anderson* [1991]; *Battisti and Sarachik* [1995]). The present paper reviews progress in ENSO understanding, with a major focus on development after the TOGA decade. However, for the sake of continuity it also briefly summarizes the progress made before and during the TOGA decade.

ENSO's low-frequency modulation and the relation of high-frequency influences to ENSO are recent research topics of ENSO. The 1997–98 El Niño was characterized by exceptionally strong high-frequency wind variability during the onset phase. Numerical models, which succeeded in predicting the onset of the 1997–98 El Niño, were unable to forecast its intensity [e.g., *Barnston et al*., 1999; *Landsea and Knaff*, 2000] until the March 1997 westerly wind burst (WWB) was incorporated. This may suggest the importance of the intraseasonal variability (the WWB and the Madden-Julian Oscillation), and has stimulated scientists to further investigate the roles of high-frequency forcing in ENSO.

ENSO is an irregular oscillation, both in frequency and amplitude. Its recurrence varies usually between two and seven years. Furthermore, its characteristics are modulated on decadal and multidecadal timescales [e.g., *Enfield and Cid-Serrano,* 1991; *Mokhov et al*., 2000]. In terms of amplitude, there are periods (decades or longer) during which ENSO is more energetic or there are more El Niños than La Niñas (e.g., since the mid-1970s), and vice versa. Such features can be viewed as a nearly regular ENSO oscillation superimposed on natural decadal and multidecadal oscillations and on a warming trend [*Lau and Weng*, 1999; *Cai and Whetton*, 2001a; *Philander and Fedorov*, 2003]. It can also be viewed as a chaotic/irregular ENSO, or stochastic fluctuations, or nonlinear modulation by a changing background state. Decadal-multidecadal variability of ENSO appears to influence the global atmospheric circulation [*Diaz et al*., 2001], and thus the climate over many parts of the world [*Power et al*., 1999; *Torrence and Webster*, 1999; *Janicot et al*., 2001; *Gershunov and Barnett*, 1998]. Such variability appears to alter the ocean productivity of the Pacific Ocean [*Chavez et al*., 2003] and ENSO predictability [e.g., *Balmaseda et al*., 1995; *Flugel and Chang*, 1998; *Kirtman and Schopf*, 1998]. Therefore, many studies, especially after the TOGA decade, have focused on ENSO's low-frequency modulation.

The present paper is organized as follows. Section 2 briefly describes observations of ENSO. Section 3 reviews our present understanding of ENSO mechanisms. Section 4 briefly summarizes the effects of high-frequency variability on ENSO. Section 5 reviews a newly and recently developed facet of ENSO, its low-frequency modulation. The paper ends in Section 6 with some discussions and ideas for the future.

2. OBSERVATIONS OF ENSO

2.1. The ENSO Observing System

The backbone of the ENSO observing system (Plate 1) is the TAO (Tropical Atmosphere Ocean) array of about 70 moored buoys [*Hayes et al*., 1991; *McPhaden,* 1995]. Most of them are equipped with a 500-m thermistor chain and meteorological sensors. At the equator five to seven moorings are equipped with ADCP (Acoustic Doppler Current Profiler) and current meters [*McPhaden*, 1995]. Developed during TOGA as a multinational program among France, Japan, South Korea, Taiwan, and United States, this array is now supported by the US with the dedicated R/V Ka'imimoana and by Japan with their TRITON program (hence the official name of TAO/TRITON since January 1, 2000). The ocean observing system is completed by a Voluntary Observing Ship (VOS) program, an island tide-gauge network, and a system of surface drifters. All the data are transmitted in near-real time to the Global Telecommunication System, for research and prediction purposes. A set of meteorological and oceanographic satellites complete all these measurements, with particular emphasis on the TOPEX/Poseidon altimeter that has proved especially useful in observing and analyzing tropical ocean variability [*Picaut and Busalacchi*, 2001]. Detailed information about the ENSO observing system, such as key variables, sampling requirements and uncertainties, can be found in *McPhaden et al*. [1998].

The obvious parameters for observing the ENSO coupled phenomenon are surface wind stress and SST (through a combination of satellite and in situ data) but the subsurface temperature has proved to be surprisingly useful in diagnosing ENSO. The basic 2–7 year period of ENSO is set by the thermal inertia of the ocean upper layer. Most of the heat content variability in low-latitude oceans is situated in this layer, and thus is directly reflected in sea-level height. Hence, measurements of the upper ocean thermal field and sea level are also fundamental to ENSO. Upper-layer temperature is mostly controlled by a specific low-latitude dynamics (i.e., equatorial waves), and current measurements are needed, especially near the equator with the vanishing of geostrophy (Coriolis force). Subsurface temperature data can provide estimates of equatorial upwelling and mixing above the thermocline, which are greatly needed. Although less important than in non-equatorial latitudes, surface heat fluxes are also required.

The TOGA observing system was devoted to the large-scale monitoring of the upper tropical oceans, with emphasis on the tropical Pacific. However, there has been considerable controversy regarding the physics that maintain and perturb the western Pacific warm pool, which is believed to be a center of action for ENSO. Hence, a multinational oceanography-meteorology experiment was conceived and carried out in 1992–93, with an intensive observation period (November 1992–February 1993) embedded into a yearlong period of enhanced monitoring. Twelve research vessels, seven research aircrafts, numerous ground-based stations, and additional moored and drifting buoys have collected a unique set of data. The plans for the TOGA Coupled Ocean-Atmosphere Response Experiment (COARE) are listed in *Webster and Lukas* [1992], and the results are summarized in *Godfrey et al*. [1998].

2.2. Lessons From the ENSO Observing System and Further Needs

The biggest achievement of TOGA was the installation for the first time of an ocean observing system. It improved the understanding and modeling of ENSO, and proved its prediction capability through the evidence of subsurface memory [*Latif et al*., 1998]. It provided since 1985 a set of high-quality data, which associated with older and less reliable data showed that El Niño behaves differently over the last decades (Plate 2). The warm SST anomalies associated with El Niño events between 1950 and 1976 first peaked along the South American coast in the boreal spring of the El Niño year and then propagated westward [*Rasmusson and Carpenter*, 1982]. The El Niño events between 1976 and 1996 seemed to start from the equatorial western and central Pacific, and the coastal warming occurred in the boreal spring subsequent to the El Niño year rather than in the boreal spring of the El Niño year [*Wang*, 1995a; *Wang and An*, 2002]. The 1997–98 El Niño started in both the central Pacific and the South American coast during the spring of 1997 and the 2002–03 El Niño started and remained in the equatorial central Pacific. Why El Niños started differently in the last five decades is not understood yet. Being the most likely factor, the role of high- and low-frequency variabilities will be discussed in the following sections.

The duration of TOGA-COARE was not sufficient to understand the link between the intraseasonal westerly winds, such as the Madden-Julian Oscillation (MJO) and Westerly Wind Bursts (WWBs), and El Niño [*Lengaigne et al*., this volume]. With the discovery of the salinity stratified barrier-layer in the western Pacific warm pool [*Lukas and Lindstrom*, 1991] and the possibility that it influences the development of El Niño [*Maes et al*., 2002], there is a strong need for more salinity measurements and in particular sea surface salinity (SSS). The end of TOGA was marked by the progressive replacement of bucket samples on VOS routes by thermosalinograph. Together with satellite missions such as SMOS and Aquarius, these in situ SSS measurements will undeniably improve the ENSO observing system [*Lagerloef and Delcroix*, 2001].

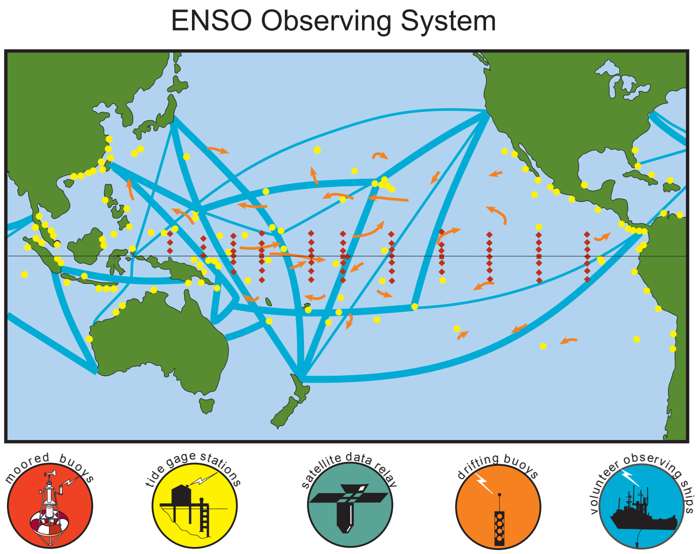

Plate 1. In situ components of the ENSO observing system. The four major elements are the TAO/TRITON array of moored buoys (red diamonds), an island tide-gauge network (yellow circles), surface drifters (arrows), and the volunteer ship program (blue lines). Various satellites are intensively used to complement the in situ network. This ensemble of instruments delivers in near-real time data on surface and subsurface temperature and salinity, wind speed and direction, sea level, and current velocity (Courtesy of Michael J. McPhaden, TAO Project Office).

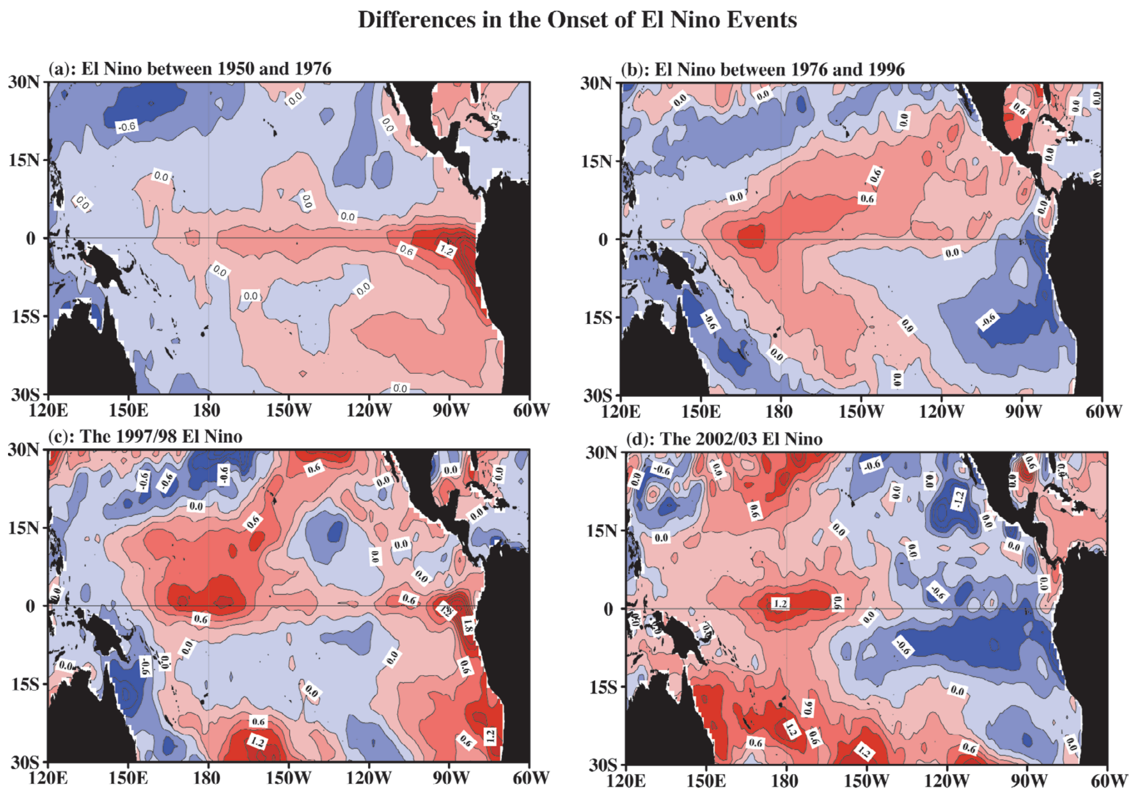

Plate 2. SST anomaly composites showing the differences in the onset of El Niño over the last five decades. (a) El Niño between 1950 and 1976, (b) El Niño between 1977 and 1996, (c) the 1997–98 El Niño, and (d) the 2002–03 El Niño. The composites are calculated by averaging the SST anomalies during March–May of the El Niño year. Since the 2002–03 El Niño started earlier, its composite used the SST anomalies of December 2001 to February 2002.

As compared to the western Pacific warm pool, the eastern tropical Pacific was somewhat neglected during TOGA, and the 5-year experiment EPIC (Eastern Pacific Investigation of Climate processes) was launched in 1999. This experiment was designed to improve the understanding of the intertropical convergence zone (ITCZ), its interaction with the cold water originating from the equatorial upwelling, and the physics of the stratus cloud deck that forms over the cold water off South America [*Cronin et al*., 2002].

As discussed in Section 5, understanding of the low-frequency variations of ENSO requires an expansion of the present ENSO observing system and its extension toward the western boundary and beyond the tropics. A main goal of the Pacific Basin Extended Climate Study (PBECS) is to provide sufficient additional in situ and satellite observations to constrain data-assimilating models well enough that the processes affecting decadal modulation of ENSO can be studied in detail [*Kessler et al*., 2001]. This will require a whole set of additional measurements, such as repeated high-resolution expendable and hydrographic sections, several process experiments, and the integration with the Argo program of profiling floats [*Roemmich et al.*, 2001] and the Global Ocean Data Experiment (GODAE) [*Smith et al*., 2001]. All these efforts are part of the CLIVAR (Climate Variability and Predictability) program.

It will be long before these observing systems and experiments produce sufficient high-quality observations to explain the decadal modulation of ENSO. This strengthens the need for historical and paleoclimate records of ENSO, with coral, tree-ring, tropical ice core, sediment or other proxies [e.g., *Ortlieb*, 2000; *Mann et al.*, 2000; *Markgraf and Diaz*, 2000; *Tudhope et al.,* 2001]. Associated with specific model studies, these records will help understand the evolution of ENSO in the past, present, and future.

3. ENSO MECHANISMS

The theoretical explanations of ENSO can be loosely grouped into two frameworks. First, El Niño is one phase of a self-sustained, unstable, and naturally oscillatory mode of the coupled ocean-atmosphere system. Second, El Niño is a stable (or damped) mode triggered by atmospheric random "noise" forcing. Whatever the case, ENSO involves the positive ocean-atmosphere feedback of *Bjerknes* [1969]. The early idea of *Wyrtki's* [1975] sea level "buildup" in the western Pacific warm pool treats El Niño as an isolated event. Wyrtki suggested that prior to El Niño, the easterly trade winds strengthened, and there was a "buildup" in sea level in the western Pacific warm pool. A "trigger" is a rapid collapse of the easterly trade winds. When this happens, the accumulated warm water in the western Pacific would surge eastward in the form of equatorial downwelling Kelvin waves to initiate an El Niño event. The recent studies have suggested atmospheric stochastic forcing as important "triggers" of El Niño. On the other hand, numerical models with tunable model parameters suggest that ENSO is a self-sustained mode of the coupled ocean-atmosphere system in some parameter regimes. Additionally, many studies have shown that the ocean-atmosphere coupling can produce slow modes that can explain both eastward and westward propagating events.

3.1. ENSO Oscillator Models

Bjerknes [1969] was the first to hypothesize that a positive ocean-atmosphere feedback causes El Niño. Although the starting point is arbitrary, an initial positive SST anomaly in the equatorial eastern Pacific reduces the east-west SST gradient and hence the strength of the Walker circulation, resulting in weaker trade winds along the equator. The weaker trade winds in turn drive the ocean circulation changes that further reinforce SST anomaly. This positive feedback leads the equatorial Pacific to a warm state. For the coupled system to oscillate, a negative feedback is needed to turn the warm state around. Since the 1980s, four major negative feedbacks have been proposed: (1) wave reflection at the western boundary, (2) a discharge process, (3) a western Pacific wind-forced Kelvin wave, and (4) anomalous zonal advection. These negative feedbacks correspond to the delayed oscillator [*Suarez and Schopf*, 1988; *Battisti and Hirst*, 1989], the recharge oscillator [*Jin*, 1997], the western Pacific oscillator [*Weisberg and Wang*, 1997a; *Wang et al*., 1999b], and the advective-reflective oscillator [*Picaut et al.*, 1997]. Since these models also operate for initial negative SST, they can produce ENSO-like oscillations.

With the different conceptual oscillator models capable of producing ENSO-like oscillations, more than one may operate in nature. Motivated by the existence of different oscillator models, *Wang* [2001a, b] formulated and derived a unified ENSO oscillator from the dynamics and thermodynamics of the coupled ocean-atmosphere system that is similar to the *Zebiak and Cane* [1987] coupled model. The unified oscillator includes the physics of all ENSO oscillator models (Figure 1). As suggested by the unified oscillator, ENSO may be a multi-mechanism phenomenon (see *Picaut et al*. [2002] for observational evidence) and the relative importance of different mechanisms may be time-dependent. Observations show that ENSO displays both eastern and western Pacific interannual anomaly patterns [e.g., *Rasmusson and Carpenter*, 1982; *Weisberg and Wang*, 1997b; *Mayer and Weisberg*, 1998; *Wang et al*., 1999b; *McPhaden*, 1999; *Wang and Weisberg*, 2000; *Vialard et al*., 2001]. Thus, the unified oscillator considers both eastern and western Pacific anomaly variations:

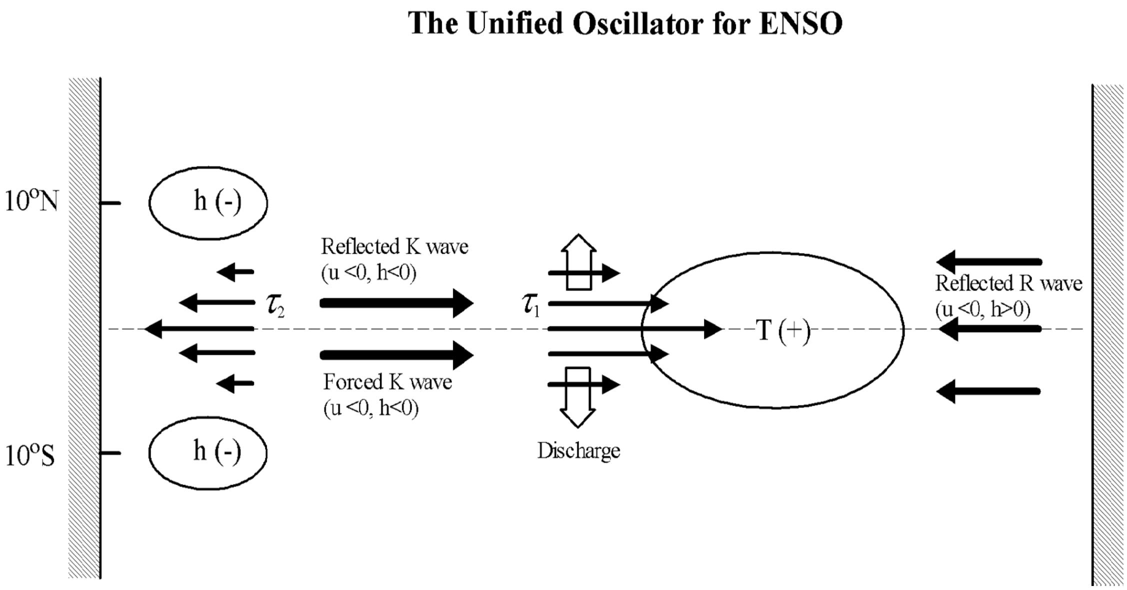

Figure 1. Schematic diagram of the unified oscillator for ENSO. Bjerknes positive ocean-atmosphere feedback leads the equatorial central/eastern Pacific to a warm state (El Niño). Four negative feedbacks, required to turn the warm state around, are (1) reflected Kelvin wave at the ocean western boundary, (2) discharge process due to Sverdrup transport, (3) western Pacific wind-forced Kelvin wave, and (4) reflected Rossby wave at the ocean eastern boundary. These negative feedbacks correspond to the delayed oscillator, the recharge oscillator, the western Pacific oscillator, and the advective-reflective oscillator. The unified oscillator suggests that all of the four negative feedbacks may work together in terminating El Niño warming. The four ENSO oscillators are special cases of the unified oscillator.

$$\frac{dT}{dt} = a\tau_1 - b_1\tau_1(t-\eta) + b_2\tau_2(t-\delta) - b_3\tau_1(t-\mu) - \varepsilon T^3, \quad (1a)$$

$$\frac{dh}{dt} = -c\tau_1(t-\lambda) - R_h h\ , \quad (1b)$$

$$\frac{d\tau_1}{dt} = dT - R_{\tau 1}\tau_1\ , \quad (1c)$$

$$\frac{d\tau_2}{dt} = eh - R_{\tau 2}\tau_2\ , \quad (1d)$$

where T is SST anomaly in the equatorial eastern Pacific, h is thermocline depth anomaly in the off-equatorial western Pacific, and τ_1 and τ_2 are zonal wind stress anomalies in the equatorial central Pacific and in the equatorial western Pacific, respectively. The parameters a, b_1, b_2, b_3, c, d, and e are constants. The parameters η, δ, μ, and λ represent the delay times. The parameters ε, R_h, $R_{\tau 1}$, and $R_{\tau 2}$ are damping coefficients.

The first term on the right-hand side (RHS) of equation (1a) represents the positive feedback in the coupled system. The second term represents the negative feedback due to wave reflection at the western boundary. The third term represents the negative feedback due to the wind-forced wave contribution in the equatorial western Pacific. The fourth term represents the effect of wave reflection at the eastern boundary. The last term is a cubic damping term that does not affect oscillatory behavior, but it limits anomaly growth [*Battisti and Hirst*, 1989; *Wang*, 2001a]. Equation (1b) states that the off-equatorial western Pacific thermocline anomaly is controlled by the wind stress in the equatorial central Pacific, with a damping rate of R_h. Equation (1c) shows that zonal wind stress anomaly in the equatorial central Pacific is related to the eastern Pacific SST anomaly, and equation (1d) states that the zonal wind stress anomaly in the equatorial western Pacific is related to the off-equatorial western Pacific thermocline anomaly. By further simplifications and assumptions, the unified oscillator can reduce to the different ENSO oscillators.

3.1.1. The delayed oscillator. The delayed oscillator (Figure 2) does not consider the coupled role of the western Pacific in ENSO and wave reflection at the eastern boundary. By setting $b_2 = 0$ and $b_3 = 0$ in equation (1a), the western Pacific variables τ_2 and h are decoupled from the coupled system. If we further drop the time derivative of equation (1c), the unified oscillator reduces to:

$$\frac{dT}{dt} = \frac{ad}{R_{\tau 1}}T - \frac{b_1 d}{R_{\tau 1}}T(t-\eta) - \varepsilon T^3\ . \quad (2)$$

Equation (2) is the delayed oscillator of *Suarez and Schopf* [1988] and *Battisti and Hirst* [1989]. The first term on RHS of equation (2) represents the positive feedback by ocean-atmosphere coupling in the equatorial eastern Pacific, i.e., the Bjerknes feedback. The second term is the delayed nega-

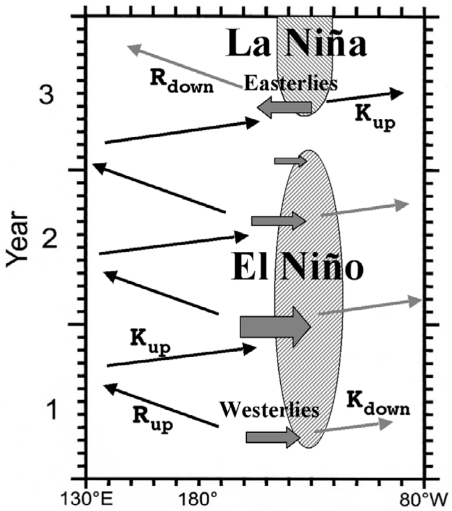

Figure 2. Schematic diagram of the delayed oscillator for ENSO. Positive SST anomalies in the equatorial eastern Pacific cause westerly wind anomalies that drive Kelvin waves eastward and act to increase the positive SST anomalies. The westerly wind anomalies also generate oceanic equatorial Rossby waves, which propagate westward and eventually reflect from the western boundary as equatorial Kelvin waves. Since the thermocline anomalies for the reflected Kelvin waves have an opposite sign to those of the directly forced Kelvin waves, they provide a negative feedback for the coupled system to oscillate.

tive feedback by free equatorial Rossby waves generated in the eastern Pacific coupling region that propagate to and reflect from the western boundary, returning as equatorial Kelvin waves to reverse the anomalies in the eastern Pacific coupling region.

The earliest idea of explaining the oscillatory nature of ENSO was proposed by *McCreary* [1983], based on the reflection of subtropical oceanic Rossby waves at the western boundary. *McCreary* [1983] and *McCreary and Anderson* [1984] explored shallow water ocean dynamics coupled to wind stress patterns that are changed by a discontinuous switch depending on thermocline depth. They showed how oceanic Rossby waves might be involved in generating the interannual oscillations associated with ENSO. In spite of the use of a discontinuous switch in their atmosphere and of reflection of subtropical Rossby waves, the idea of basin adjustment processes has been incorporated by later work. *Suarez and Schopf* [1988] introduced the delayed oscillator model of equation (2) as a candidate mechanism for ENSO. Based on the coupled model of *Zebiak and Cane* [1987], *Battisti and Hirst* [1989] formulated and derived a version of the *Suarez and Schopf* [1988] delayed oscillator model.

Graham and White [1988] presented sparse evidence of off-equatorial Rossby waves and their reflection at the western boundary and then empirically constructed a conceptual oscillator model for ENSO. As shown in *McCreary and Anderson* [1991], the conceptual equations of the *Graham and White* model can be reduced to a single equation that is similar to equation (2) (also see the comments of *Neelin et al*. [1998]).

The work of *McCreary* [1983], *McCreary and Anderson* [1984], and *Graham and White* [1988] emphasized the reflection of off-equatorial Rossby waves at the western boundary whose importance for ENSO has been debated. *Kessler* [1991] and *Battisti* [1989, 1991] argue that the equatorial Kelvin wave results primarily from the reflection of the gravest Rossby wave mode and that off-equator (poleward of ±8°) variations should not be a major factor in ENSO. In contrast, *Graham and White* [1991] contend that coupled model simulations of ENSO are greatly altered if effects poleward of ±8° are neglected. However, all of these studies recognized that wave reflection at the western boundary is important in terminating El Niño. *Li and Clarke* [1994] challenged the validation of the delayed oscillator by noting a low lag correlation between the western Pacific equatorial Kelvin wave amplitude and zonal wind forcing that is inconsistent with the delayed oscillator theory. *Mantua and Battisti* [1994] argued that wave reflection at the western boundary did account for the termination of El Niño and that the low lag correlation is due to irregularity of ENSO. In any case, the reflection efficiency of the western boundary is disrupted by the presence of the throughflow and numerous islands. Its estimation from simple models or observations [e.g., *Clarke*, 1991; *Zang et al*., 2002; *Boulanger et al*., 2003] is complicated by the coupled nature of the equatorial western Pacific and not enough observations.

3.1.2. The recharge oscillator. The recharge oscillator (Figure 3) considers variations of eastern Pacific SST and western Pacific thermocline anomalies. As argued by *Jin* [1997], equatorial wave dynamics are important in the adjustment of equatorial ocean, but wave propagations are not explicit in the recharge model. If the time derivatives in equations (1c) and (1d) are dropped and all delay parameters are set to zero (i.e., $\eta = 0$, $\delta = 0$, and $\lambda = 0$) and $b_3 = 0$, the unified oscillator reduces to:

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

$$\frac{dh}{dt} = -\frac{cd}{R_{\tau 1}} T - R_h h \,. \tag{3b}$$

The mathematical form of equation (3) is the same as the recharge oscillator of *Jin* [1997]. In the Jin's recharge oscillator model, h is the thermocline anomaly in the equatorial western Pacific.

Wyrtki [1975] first suggested a buildup in the western Pacific warm water as a necessary precondition to the development

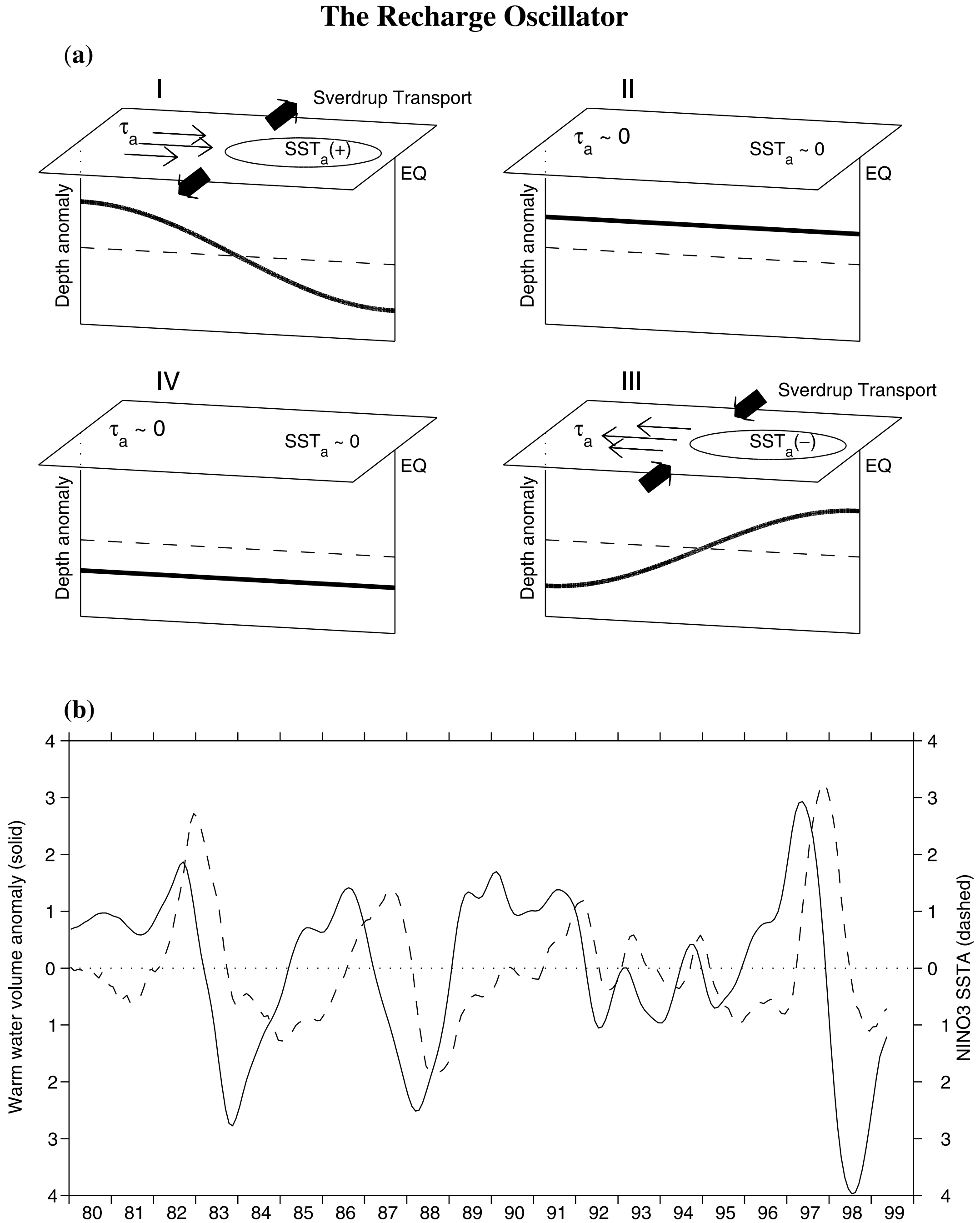

Figure 3. Schematic diagram of the recharge oscillator for ENSO. (a) The four phases of the recharge oscillation: (I) the warm phase, (II) the warm to cold transition phase, (III) the cold phase, and (IV) the cold to warm transition phase. During the warm phase of ENSO, the divergence of Sverdrup transport associated with equatorial central Pacific westerly wind anomalies and equatorial eastern Pacific warm SST anomalies results in the discharge of equatorial heat content. The discharge of equatorial heat content leads to a transition phase in which the entire equatorial Pacific thermocline depth is anomalously shallow due to the discharge of equatorial heat content. This anomalous shallow thermocline at the transition phase allows anomalous cold waters to be pumped into the surface layer by climatological upwelling and then leads to the cold phase. The converse occurs during the cold phase of ENSO. (b) Time series of the Niño3 SST anomalies (dashed; °C) and warm water volume anomalies (solid; 10^{14} m^3) over the entire equatorial tropical Pacific Ocean (5°S–5°N, 120°E–80°W) (courtesy of Christopher S. Meinen).

of El Niño. This concept was later modified by covering the entire tropical Pacific Ocean between 15°S and 15°N [*Wyrtki*, 1985]. Prior to El Niño, upper ocean heat content or warm water volume over the entire tropical Pacific tends to build up (or recharge) gradually, and during El Niño the accumulated warm water is flushed toward (or discharged to) higher latitudes. After the discharge, the eastern tropical Pacific becomes cold (La Niña) with the shallowing of the thermocline and then warm water slowly builds up again (recharge) before the occurrence of the next El Niño (see Figure 3). The recharge and discharge processes have been also examined by *Zebiak* [1989a], *Miller and Cheney* [1990], and *Springer et al*. [1990]. The concept of the recharge and discharge processes is further emphasized by *Jin* [1997]. Based on a coupled system that is similar to the coupled model of *Zebiak and Cane* [1987], *Jin* [1997] formulated and derived the recharge oscillator model.

Many studies have recently attempted to test the validity of the recharge oscillator model by using observational data [e.g., *Meinen and McPhaden*, 2000, 2001; *Hasegawa and Hanawa*, 2003a; *Holland and Mitchum*, 2003; *Sun*, 2003]. These observational studies basically demonstrate the recharge and discharge of the equatorial Pacific warm water during the evolution of ENSO. However, the more appropriate variable in equation (3) may be one that represents the warm water over the entire equatorial Pacific rather than the one only in the equatorial western Pacific. These studies show that the warm water in the entire equatorial Pacific band (for example, 5°S–5°N) highly correlates with the Niño3 SST anomalies, with the former leading the latter by about two seasons (Figure 3b). The correlation between the equatorial western Pacific warm water and the Niño3 SST anomalies is lower but still significant, with the western Pacific warm water leading by five seasons. *Mechoso et al*. [2003] tested the validity of the recharge oscillator model by fitting their coupled GCM output into this model. They suggested that the recharge oscillator could provide a plausible representation of their ENSO simulation. Misfits between the recharge oscillator and the coupled GCM oscillatory mode may be attributed to additional physics that are not included in the recharge oscillator.

There is a debate on the latitudinal bands of the recharge and discharge of warm water. *Wyrtki* [1985] defined the warm water in the tropical Pacific between 15°S and 15°N. *Miller and Cheney* [1990] and *Springer et al*. [1990] showed that, during El Niño, the warm water volume is decreased near the equatorial band (8°S–8°N and 5°S–5°N, respectively), whereas the volume of the tropical Pacific is not affected by ENSO due to water recirculation in the tropical North Pacific. Recently, *Holland and Mitchum* [2003] seem to reconcile this conflict by demonstrating that warm water is indeed lost from the tropical Pacific as a whole over the course of an El Niño event, as suggested by *Wyrtki* [1985]. This loss, however, is relatively small compared to the redistribution within the tropics. *Kug and Kang* [2003] showed that during El Niño meridional transport in the Northern Hemisphere is larger than that into the Southern Hemisphere, and that the asymmetric characteristics are mainly due to a southward shift of the maximum westerly wind anomalies during the mature phase of El Niño [*Harrison and Vecchi*, 1999].

Sun [2003, this volume] presented a "heat pump" hypothesis for ENSO. An increase in the warm pool SST increases the zonal SST contrast that strengthens the easterly trade wind and then helps the ocean to store more heat to the subsurface ocean. Because of the stronger wind and the resulting steeper tilt of the equatorial thermocline, the coupled system is potentially unstable and is poised to release its energy through a stronger warming (i.e., a larger amplitude of El Niño). The occurrence of El Niño pushes the accumulated heat poleward and prevents the further heat buildup in the western Pacific, thereby stabilizing the coupled system. This ENSO "heat pump" hypothesis is conceptually similar to the physics of the recharge oscillator.

3.1.3. The western Pacific oscillator. The western Pacific oscillator (Figure 4) emphasizes the coupled role of the western Pacific anomaly patterns in ENSO. This oscillator model does not necessarily require wave reflections at the western and eastern boundaries. Neglecting the feedbacks due to wave reflections at the western and eastern boundaries in the unified oscillator by setting $b_1 = 0$ and $b_3 = 0$, equations (1a)–(1d) reduce to:

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

$$\frac{dh}{dt} = -c\tau_1(t-\lambda) - R_h h\ , \tag{4b}$$

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

$$\frac{d\tau_2}{dt} = eh - R_{\tau 2}\tau_2\ . \tag{4d}$$

Equation (4) is the western Pacific oscillator of *Weisberg and Wang* [1997a].

Arguing from the vantage point of a *Gill* [1980] atmosphere, condensational heating due to convection in the equatorial central Pacific [*Deser and Wallace*, 1990; *Zebiak*, 1990] induces a pair of off-equatorial cyclones with westerly wind anomalies on the equator. These equatorial westerly wind anomalies act to deepen the thermocline and increase SST in the equatorial eastern Pacific, thereby providing a positive feedback for anomaly growth [represented by the first term of RHS of equation (4a)]. On the other hand, the off-equatorial cyclones raise the thermocline there via Ekman pumping.

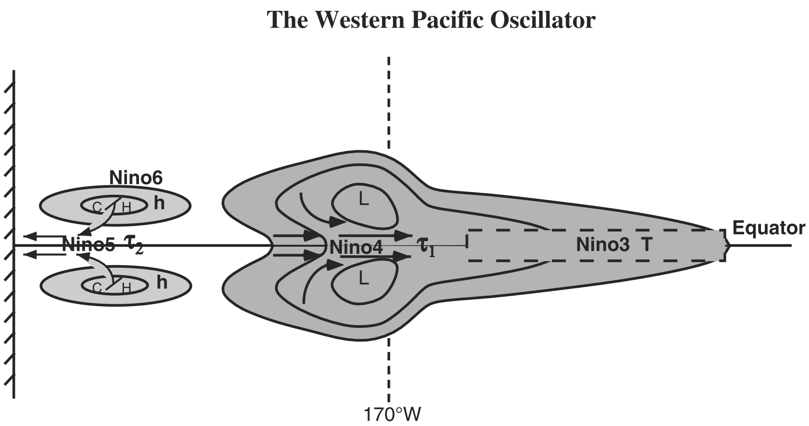

Figure 4. Schematic diagram of the western Pacific oscillator for ENSO. Condensational heating in the central Pacific induces a pair of off-equatorial cyclones with westerly wind anomalies in the Niño4 region. The Niño4 westerly wind anomalies act to deepen the thermocline and increase SST in the Niño3 region. On the other hand, the off-equatorial cyclones raise the thermocline there via Ekman pumping. Thus, a shallow off-equatorial thermocline anomaly expands over the western Pacific leading to a decrease in SST and an increase in SLP in the Niño6 region. During the mature phase of El Niño, the Niño6 anomalous anticyclone initiates equatorial easterly wind anomalies in the Niño5 region. The Niño5 easterly wind anomalies cause upwelling and cooling that proceed eastward as a forced Kelvin wave response providing a negative feedback for the coupled system to oscillate.

Thus, a shallow off-equatorial thermocline anomaly expands over the western Pacific [represented by equation (4b)], leading to a decrease in SST and an increase in sea level pressure in the off-equatorial western Pacific. This results in off-equatorial anomalous anticyclones during the mature phase of El Niño, which initiate equatorial easterly wind anomalies in the western Pacific [*Wang et al*., 1999b; *Wang*, 2000]. These equatorial easterly wind anomalies cause upwelling and cooling that proceed eastward as a forced ocean response providing a negative feedback [represented by the second term on RHS of equation (4a)]. Equations (4c) and (4d) relate the zonal wind stress anomalies in the equatorial central Pacific to the equatorial eastern Pacific SST anomalies, and the zonal wind stress anomalies in the equatorial western Pacific to the off-equatorial western Pacific thermocline anomalies, respectively. The western Pacific oscillator is also consistent with the onset of El Niño. During the onset and development phases of an El Niño, twin anomalous cyclones in the off-equatorial western Pacific initiate equatorial westerly wind anomalies [e.g., *Wang and Weisberg*, 2000] that produce downwelling Kelvin waves to warm the equatorial central and eastern Pacific.

Earlier studies have shown that the equatorial easterly wind anomalies in the western Pacific can force upwelling Kelvin waves that raise the thermocline in the east [e.g., *Tang and Weisberg*, 1984; *Philander*, 1985]. Recently, *McPhaden and Yu* [1999], *Delcroix et al*. [2000], *Boulanger and Menkes* [2001], *Vialard et al*. [2001], *Picaut et al*. [2002], *Boulanger et al*. [2003], and *Hasegawa and Hanawa* [2003a] have shown that the western Pacific oscillator operates in nature. The western Pacific wind-forced Kelvin waves play an important role in terminating ENSO. For example, *Boulanger and Menkes* [2001] and *Boulanger et al*. [2003] demonstrated that, for the 1997–98 El Niño, about two-thirds of the Kelvin wave amplitude is actually forced by easterly wind in the western Pacific and the other one-third is due to wave reflection at the western boundary. In nature, the equatorial easterly wind anomalies in the western Pacific are observed to become larger and larger (both amplitude and fetch) and move eastward after the mature phase of El Niño. The impact of the easterly wind-forced upwelling Kelvin waves is thus gradually strengthened by the increasing fetch and eastward migration of the easterly wind anomalies [e.g., *Picaut et al*., 2002].

3.1.4. The advective-reflective oscillator. Picaut et al. [1996] found an oceanic convergence zone at the eastern edge of the warm pool, which is advected in phase with the Southern Oscillation Index over thousands of kilometers, eastward during El Niño, westward during La Niña. Based on this finding, the study of *Picaut and Delcroix* [1995] regarding zonal advection and wave reflection, and the fact that westerly (easterly) winds penetrate into the central (western) equatorial Pacific during El Niño (La Niña), *Picaut et al*. [1997] proposed a conceptual advective-reflective oscillator for ENSO (Figure 5). In this concept, they emphasize a positive feedback of zonal currents that advect the western Pacific warm pool toward the east during El Niño. Three negative feedbacks tending to push the warm pool back to its original position and then into the western Pacific are: anomalous zonal current associated with wave reflection at the western boundary, anomalous zonal current associated with wave reflection at the eastern

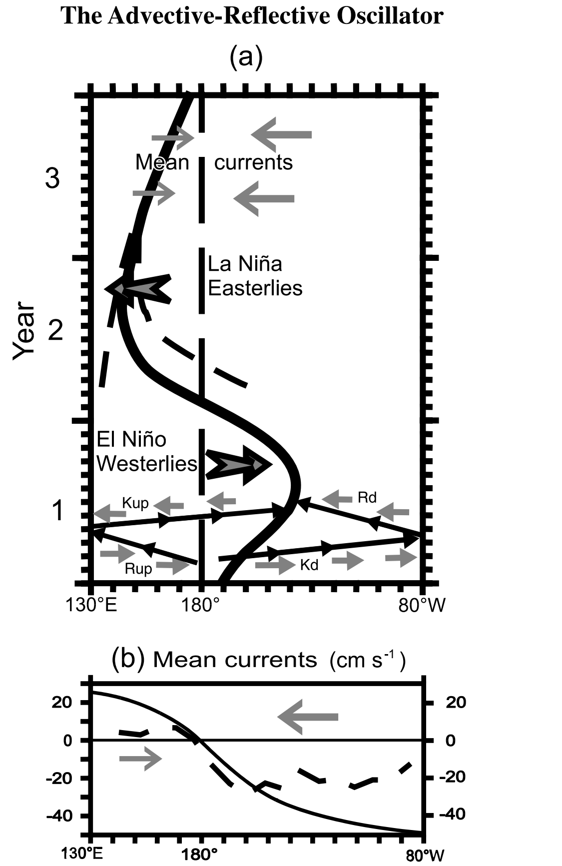

Figure 5. Schematic diagram of the advective-reflective oscillator for ENSO. This oscillator emphasizes a positive feedback of zonal currents that advect the western Pacific warm pool toward the east during El Niño. Three negative feedbacks tending to push the warm pool back to the western Pacific are: anomalous zonal current associated with wave reflection at the western boundary; anomalous zonal current associated with wave reflection at the eastern boundary; and mean zonal current converging at the eastern edge of the warm pool.

boundary, and mean zonal current converging at the eastern edge of the warm pool. During the warm phase of ENSO, equatorial westerly wind anomalies in the central Pacific produce equatorial upwelling Rossby and downwelling Kelvin waves that propagate westward and eastward, respectively. The westward propagating upwelling Rossby waves reflect into upwelling Kelvin waves after they reach the western boundary, whereas the eastward propagating downwelling Kelvin waves reflect into downwelling Rossby waves at the eastern boundary. Since both upwelling Kelvin and downwelling Rossby waves have westward zonal currents in the equatorial band, they tend to push the warm pool back to its original position and then into the western Pacific. These negative feedbacks along with the negative feedback of the mean zonal current make the coupled system to oscillate. Recent observational and modeling support of the advective-reflective oscillator can be found in *Delcroix et al*. [2000], *Clarke et al.,* [2000], *An and Jin* [2001], *Picaut et al.* [2001, 2002], and *Dewitte et al*. [2003].

The advective-reflective oscillator of *Picaut et al.* [1997] can also be represented by a set of simple and heuristic equations. By setting $b_2 = 0$ in equation (1a), the unified oscillator model is reduced to:

$$\frac{dT}{dt} = a\tau_1 - b_1\tau_1(t-\eta) - b_3\tau_1(t-\mu) - \varepsilon T^3, \tag{5a}$$

$$\frac{d\tau_1}{dt} = dT - R_{\tau 1}\tau_1. \tag{5b}$$

In derivation and formulation of the unified oscillator model [*Wang*, 2001a], it is shown that two advection terms of $u\partial\overline{T}/\partial x$ and $\overline{u}\partial T/\partial x$ are included in the first term of $a\tau_1$ in equation (5a) (also see *Battisti and Hirst* [1989]). Thus, the effects of zonal current are included in the term of $a\tau_1$. The effect of anomalous zonal current associated with wave reflection at the western boundary can be explained by the term of $-b_1\tau_1(\tau-\eta)$ in equation (5a) (also see *Clarke et al*. [2000]). The negative feedback of wave reflection at the eastern boundary is represented by the third term of RHS of equation (5a).

3.2. Slow (SST) Modes

Interaction between the tropical Pacific Ocean and atmosphere can produce coupled slow modes. The simple coupled system (with constant mean states) displays a slow westward propagating unstable mode [*Gill*, 1985; *Hirst*, 1986] and a slow eastward propagating unstable mode [*Philander et al*., 1984; *Yamagata*, 1985; *Hirst*, 1986]. These two modes are further investigated numerically by *Hirst* [1988] and analytically by *Wang and Weisberg* [1996], showing that they can propagate and continuously regenerate on interannual timescales. The delayed oscillator is not relevant to these unstable modes. For example, *Wang and Weisberg* [1994] showed that the evolution of the eastward propagating mode is nearly identical for the closed and open ocean western boundary conditions (the open western boundary does not allow waves to be reflected).

Neelin [1991] introduced a slow SST mode theory by emphasizing physical processes in the oceanic surface layer (not related to wave dynamics). Whether the coupled system favors the SST modes or the ocean-dynamics modes (associated to the delayed oscillator) is determined by the ocean adjustment process. For the ENSO timescale, there are two key adjustments: one associated with the dynamical adjustment of the equatorial ocean, and the other associated with the thermodynamical changes in SST due to air-sea coupling. When the dynamical adjustment of the ocean is fast compared with the changes in SST, the behavior of the coupled ocean-atmosphere system depends critically on the time evolution of the SST, but is less influenced by the ocean-wave dynamics. On the other hand, if the dynamical adjustment of the ocean is slow, the coupled ocean-atmosphere system is dominated by the equatorial wave dynamics that provide the "memory" for an interannual oscillation. *Jin and Neelin* [1993] and *Neelin and Jin* [1993] provided a unified view between the slow SST mode and the ocean-dynamics modes, by arguing that in most of the parameter space the coupled modes will have a mixed nature, i.e., 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.

A number of physical processes compete in terms of the direction of modes' propagation (eastward and westward). Propagation is first discussed by considering a region with a positive SST anomaly in the equatorial eastern/central Pacific. The wind responses to the west of the region are anomalous westerly, whereas they are easterly to the east [*Gill*, 1980]. This wind distribution drives anomalous zonal currents that advect warm water to the west of the region and cold water to the east. At the same time, the anomalous westerly (easterly) winds to the west (east) induce local anomalous downwelling (upwelling). As a result of both zonal advection and downwelling, the region of positive SST anomaly expands on its western side and the SST anomaly then propagates westward. On the other hand, the anomalous westerly winds induce a deepening of the thermocline in the east, which warms SST in the east through mean upwelling. Additionally, the nonlinearity of the anomalous vertical temperature gradient by the anomalous upwelling can also warm SST in the east [*Jin et al*., 2003]. Thus, both can make the SST anomaly propagate eastward. Second, propagation of interannual anomalies is considered in the western Pacific Ocean from both dynamical and thermodynamical air-sea coupling [*Wang*, 1995b; *Philander and Fedorov*, 2003]. From the dynamical point of view, a modest disturbance in the form of a brief burst of westerly winds (or the Madden-Julian Oscillation) in the western Pacific will generate currents that transport some of the warm water eastward, thus decreasing the zonal temperature gradient. The resultant weakening of the trade winds will cause more warm water to flow eastward, causing even weaker winds. From the thermodynamical point of view, during the boreal winter and spring, the climatological zonal wind in the equatorial western Pacific varies from a weak westerly at 130°E–150°E to an easterly near the date line, with a direction reversal around 150°E. Superposition of an equatorial westerly anomaly in the above mean zonal wind in the western Pacific will have different effects on SST. In the region of a weak mean easterly between 160E°–170°E, a westerly anomaly implies a reduction in the total wind speed, resulting in an increase in SST due to reduced evaporation. However, in the region of a weak mean westerly at west of 150°E, a westerly wind anomaly increases the total wind speed, inducing the cooling of SST through enhanced evaporation. Therefore, an eastward SST gradient is produced, which in turn reinforces the equatorial westerly wind anomalies [*Lindzen and Nigam*, 1987]. The feedback between the eastward SST gradient and westerly anomalies promotes the eastward propagation of the equatorial westerly anomalies observed during an El Niño event.

3.3. A Stable Mode Triggered by Stochastic Forcing

In the ENSO views of Sections 3.1 and 3.2, the coupled tropical Pacific Ocean-atmosphere system is dynamically unstable. However, as model parameters are changed, the oscillatory and slow modes can become stable (e.g., see model parameter studies of *Battisti and Hirst* [1989]; *Hirst* [1988]; *Neelin and Jin* [1993]; *Wang and Weisberg* [1996]; *Jin* [1997]; *Wang* [2001a]). In this case, a stochastic trigger (forcing term) must be added to an oscillator model to excite an irregular oscillation [e.g., *Graham and White*, 1988; *Jin*, 1997]. ENSO as a stable mode triggered by stochastic forcing (or noise) has been suggested by many authors [e.g., *McWilliams and Gent*, 1978; *Lau*, 1985; *Penland and Sardeshmukh*, 1995; *Blanke et al*., 1997; *Kleeman and Moore*, 1997; *Eckert and Latif*, 1997; *Moore and Kleeman*, 1999a, b; *Thompson and Battisti*, 2001; *Dijkstra and Burgers*, 2002; *Larkin and Harrison, 2002; Kessler, 2002; Zavala-Garay et al*., 2003]. This hypothesis proposes that disturbances, unrelated to internal ENSO dynamics, are the source of stochastic forcing that drives ENSO. It should be pointed out that stochastic forcing might also have a low-frequency spectral tail (as a result of cumulative effect of strong or extended series of random events) that can directly drive ENSO [*Moore and Kleeman*, 1999a]. An attractive feature of this hypothesis is that it offers a natural explanation in terms of noise for the irregular behavior of ENSO variability. Since this view of ENSO requires the presence of atmospheric "noise", it easily explains why each El Niño is distinct and El Niño is so difficult to predict [e.g., *Landsea and Knaff*, 2000; *Fedorov et al*., 2003].

No matter whether El Niño is a self-sustained mode or a stable mode triggered by stochastic forcing, El Niño matures with warm SST anomalies in the equatorial central and eastern Pacific. After an El Niño reaches its mature phase, negative feedbacks are required to terminate the growth of the mature El Niño anomalies in the central and eastern Pacific. In other words, the negative feedbacks associated with the delayed oscillator, the recharge 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 triggered by stochastic forcing.

A stable mode can be either oscillatory or non-oscillatory (highly damped). For a non-oscillatory mode, there is no necessary connection between one El Niño event and the next, i.e., El Niño is sporadic, not cyclic [e.g., *Larkin and Harrison*, 2002; *Kessler*, 2002; *Philander and Fedorov*, 2003], and a random disturbance is needed to initiate each new event. For an oscillatory mode, each El Niño is related to the ensuing ENSO phases. *Mantua and Battistti* [1994] discussed three simple ENSO scenarios: (1) periodic ENSO cycle, (2) non-periodic ENSO cycle, and (3) non-periodic, non-cyclic ENSO event. In the latter two cases, the warm SST anomalies in the eastern and central Pacific are initiated by something other than a reflected Kelvin wave issued by the preceding cold event. However, the reflected upwelling Kelvin waves can be responsible for shutting down the growing instability in the equatorial central and eastern Pacific. A sequence of independent warm events can still be consistent with delayed oscillator physics since the termination of individual El Niño can occur as a result of wave reflection at the western boundary.

4. EFFECTS OF HIGH-FREQUENCY VARIABILITY ON ENSO

Variability with frequency higher than the ENSO frequency includes the seasonal cycle and the intraseasonal variability (ISV). Both seasonal cycle and ISV play roles in ENSO.

4.1. Seasonal Cycle

The seasonal cycle can contribute to the irregularity of ENSO and the ENSO phase-locking [e.g., *Jin et al.*, 1994; *Tziperman et al*., 1995; *Chang et al*., 1995]. Using numerical models, these studies showed that interannual variability is periodic without seasonal cycle forcing, but as model parameters (related to the seasonal cycle and the ocean-atmosphere coupling) are increased the interannual model solution undergoes a transition from periodic to irregular (or chaotic) through a sequence of rational fractions of the seasonal cycle: ENSO remains phase-locked to the seasonal cycle. In models, the seasonal cycle is influenced through mean background states of atmospheric wind divergence, oceanic upwelling, and so on. *Mantua and Battisti* [1995] hypothesized that interaction between ENSO and the "mobile" mode (a near-annual and westward propagating mode) is the cause for irregular variability in the *Zebiak and Cane* [1987] model simulations. The transition to chaos of a model system can occur in any of three universally recognized scenarios: the period doubling route [*Chang et al,* 1995], the quasi-periodicity route [*Tziperman et al*., 1995], and the intermittency route [*Wang et al.*, 1999a].

4.2. Intraseasonal Variability (ISV)

The prominent ISV in the western and central Pacific includes westerly wind bursts (WWB) and the Madden-Julian Oscillation (MJO). Although both the WWB and MJO show westerly winds over the western Pacific, they differ temporally and spatially. Based on the region of maximum zonal wind anomalies, *Harrison and Vecchi* [1997] and *Vecchi and Harrison* [2000] identified eight different types of WWB event. On average, WWB has zonal width between 30° and 40° longitude, meridional width between 10° and 15° latitude, and duration between 7 and 10 days. The MJO, a wave-like atmospheric phenomenon, has a time-scale of between 30–90 days and has a much larger structure than the WWB [*Madden and Julian,* 1994; *Slingo et al*., 1999]. The MJO propagates eastward and the WWB does not necessarily. The WWB tends to develop during active phases of the MJO (also tends to form from paired tropical cyclones and cold surges from mid-latitude), but the exact relationship between the WWB and MJO is not clear. They both have an influence on oceanic variability. However, the quantitative differences between the effects on the ocean by the WWB and MJO have not yet been determined. Therefore, we herein collectively review their roles in the ocean and ENSO, while *Lengaigne et al.* [this volume] provide a detailed review of the WWBs and their influence on the tropical Pacific Ocean-atmosphere system.

The ISV, associated with WWB and MJO, has a local effect on the western Pacific and a remote effect on the eastern Pacific. The local effect includes a change in mixed layer depth, surface jets, and an oceanic cooling in the western Pacific that can be explained by varying both shortwave radiation and latent heat flux. Convective activity associated with ISV increases atmospheric cloudiness that reduces shortwave radiation and then cools the western Pacific Ocean [e.g., *Weller and Anderson*, 1996]. During the boreal winter and spring, the climatological zonal wind in the equatorial western Pacific west of 150°E is a weak westerly. Thus, a westerly wind anomaly (associated with ISV) increases the total wind speed, inducing the cooling of SST through enhanced evaporation.

The remote effect of ISV on the eastern Pacific is via downwelling Kelvin waves generated by westerly wind anomalies in the western Pacific [e.g., *Kessler et al*., 1995; *Hendon et al*., 1998; *Zhang*, 2001; *Zhang and Gottschalck*, 2002; *Kutsuwada* *and McPhaden,* 2002; *Cravatte et al*., 2003]. The generated downwelling Kelvin waves propagate eastward along the thermocline to the central and eastern Pacific. The resulting rises in sea level along the South American coast are observed to occur approximately 6–7 weeks following the WWB events that generate them. Along the equator the Kelvin waves are also accompanied by anomalous surface currents that induce an eastward displacement of the eastern edge of the western Pacific warm pool [*Matsuura and Iizuka,* 2000; *Picaut et al*., 2002*; Lengaigne et al.,* 2002]. The combined effects of zonal advection and thermocline depression increase the SST in the central and eastern Pacific and thus decrease the zonal SST gradient and weaken the trade winds [*Lindzen and Nigam*, 1987]. The weakening of the trade winds will cause more warm water to flow eastward, causing even weaker trade winds. This positive feedback can result in the onset of an El Niño event, as hypothesized by *Bjerknes* [1969]. As an example, both observations and numerical models have shown that the westerly wind anomalies in the western Pacific during the boreal winter and spring of 1996–97 played an important role in the onset of the 1997–98 El Niño [e.g., *McPhaden*, 1999; *McPhaden and Yu*, 1999; *Wang and Weisberg*, 2000; *van Oldenborgh*, 2000; *Boulanger et al*., 2001; *Bergman et al*., 2001; *Picaut et al*., 2002]. *Cravatte et al*. [2003] recently noticed an oscillation in the surface winds over the warm pool with an approximate 120-day period. This oscillation, of unknown origin, generates equatorial Kelvin waves as strong as those excited by the MJO. The two sets of downwelling Kelvin waves induced by the MJO and the 120-day wind oscillation seem to be stronger during the onset of El Niño and may interfere in its development. Note that easterly winds associated with the MJO generate upwelling Kelvin waves that can participate in the demise of El Niño [*Takayabu et al.,* 1999].

As discussed in Section 3.3, the ISV has been treated as noise or disturbances that can drive or sustain ENSO. Considering the ENSO oscillator models discussed in Section 3.1, noise terms can be incorporated through equations that control the variations of atmospheric winds [e.g., *Graham and White*, 1988; *Jin*, 1997]. If the ISV is acting on a self-sustained oscillatory system, then it is a source of the irregularity of ENSO. On the other hand, if the ISV is acting on a stable system, then it is the source of both its variability and irregularity. In addition to the stability of the coupled system, the temporal and spatial structures of noise may also determine the impact of noise on ENSO [e.g., *Bergman et al*., 2001; *Fedorov*, 2002]. For example, strong MJO activity was also evident during the boreal winter of 1989–90 and the early stage of development was similar to that of 1996–97. However, the development of El Niño was aborted in May 1990. The MJO was relatively quiescent during the boreal winter of 1981–82. A strong El Niño developed during 1982, but not as rapidly as it did during 1997. All of these suggest a complex relationship between MJO and ENSO.

A more theoretical approach to El Niño as a stable mode driven by stochastic forcing has been provided in the framework of generalized stability theory [*Farrell and Ioannou*, 1996a, b]. It is argued that the coupled tropical Pacific Ocean-atmosphere system is non-normal (i.e., its low-frequency eigenvectors are non-orthogonal) [e.g., *Moore and Kleeman*, 1996; *Moore and Kleeman*, 1999a, b]. In a non-normal ENSO system, a perturbation can still experience transient growth (grow to a finite amplitude and then decay) even if the system is asymptotically stable. This is because the low-frequency eigenvectors can interfere constructively with one another since they are non-orthogonal. The spatial structure of stochastic forcing is important if it is to increase variability on seasonal-to-interannual timescales. The stochastic optimals are the spatial patterns that stochastic forcing must have in order to produce large response in a coupled model, and the optimal perturbations represent the fastest growing perturbations that can exist in the coupled system [*Moore et al*., 2003]. Using an intermediate coupled ocean-atmosphere model, *Moore and Kleeman* [1999a, b] showed that when the coupled model is subjected to stochastic noise forcing that projects on the stochastic optimals, perturbations with initial structures that are similar to the optimal perturbations are excited and subsequently grow rapidly. The stochastic component of the NCEP/NCAR reanalysis has been shown to possess such optimal structures [*Zavala-Garay et al*., 2003]. The idea is also supported by *Penland and Sardeshmukh* [1995] and *Penland* [1996] who analyzed the optimal perturbations of observed SST in the tropical Pacific and Indian Oceans.

Some studies have argued that ISV does not play a critical role in ENSO [e.g., *Zebiak*, 1989b; *Slingo et al*., 1999; *Syu and Neelin*, 2000; *Kessler and Kleeman*, 2000]. *Zebiak* [1989b] showed that, in his intermediate model, the atmospheric ISV does not seem to affect ENSO. *Syu and Neelin* [2000] demonstrated that a noisier signal with shorter timescales does not appear to have an obvious relation to the ENSO cycle in their model. *Kessler and Kleeman* [2000] concluded that the MJO can interact constructively with the onset of El Niño to amplify a developing warm event, however, the MJO on its own does not appear to be the cause of El Niño. *Slingo et al*. [1999] could not find an interannual relationship or linkage between the MJO and El Niño, based on the MJO index defined from global winds and convection. When the MJO index was based on local signals of the MJO in the Pacific, *Zhang and Gottschack* [2002] found a relation between Kelvin wave ISV forcing and SST anomalies in the eastern equatorial Pacific during El Niño, at least for the 1980–99 period. Obviously, the MJO-ENSO relationship depends on indices that measure variations of the MJO.

5. LOW-FREQUENCY VARIABILITY OF ENSO

In this section, the observational evidence of decadal-multidecadal variability and warming trends in both the tropical and mid-latitude Pacific are first discussed. The mechanisms proposed for tropical Pacific decadal-multidecadal variability are summarized, and the interpretation of tropical Pacific warming trends, global warming, and ENSO are reviewed. The difficulties and uncertainties of the studies of low-frequency variability are finally discussed. *Seager et al*., in this volume, discuss predictability of Pacific decadal variability.

5.1. Observational Evidence of Decadal-Multidecadal Variability and Warming Trends in the Tropical and Mid-latitude Pacific

Decadal and multidecadal variability in the North Pacific has been analyzed for more than a decade [e.g., *Nitta and Yamada*, 1989; *Trenberth*, 1990; *Minobe*, 2000]. Although this low-frequency variability is relatively well documented, it is still unclear if one major mode or several co-equal decadal modes of variability affect this region. For example, there is some evidence of four decadal ocean-atmosphere statistical modes that occupy a thick layer of the North Pacific Ocean [*Luo and Yamagata*, 2002]. The most studied signal has been called the PDO for Pacific (inter) Decadal Oscillation [*Mantua et al*., 1997] or NPO for North Pacific decadal-multidecadal Oscillation [*Gershunov and Barnett*, 1998]. Both correspond to the leading EOF of SST North of 20°N. The PDO appears as a recurring pattern of ocean-atmosphere variability centered over the mid-latitudes of the North Pacific. Cold PDO regimes prevailed in 1880–1924, and in 1947–1976, while warm regimes prevailed in 1925–1946 and from 1977 to the mid-1990s. Despite sparse data coverage, there is evidence of decadal variability in the mid-latitudes and subtropics of the Southern Pacific [*Garreaud and Battisti*, 1999; *Linsley et al*., 2000; *Chang et al*., 2001; *Mantua and Hare*, 2002]. In particular, the position of the South Pacific convergence zone is subject to an interdecadal oscillation, in addition to an ENSO oscillation. Both oscillations have similar amplitudes, but they appear independent [*Folland et al*., 2002]. The interdecadal variability in the South Pacific can be regarded as the quasi-symmetric manifestation of the PDO.

The tropics and in particular the tropical Pacific are marked by several decadal-multidecadal coupled modes [*Goswami and Thomas*, 2000; *White et al*., 2003]. Examination of the PDO over the entire Pacific basin reveals that its spatial signature in SST, SLP, and wind stress is somewhat similar to the "horse shoe" signature of ENSO [*Mantua et al*., 1997; *Zhang et al*., 1997; *Garreaud and Battisti*, 1999; *Mestas-Nuñez and Enfield*, 2001; *Salinger et al*., 2001]. It is marked by an equator-straddling SST anomaly in the eastern tropical Pacific less confined than those of ENSO, and by a relatively greater SST anomaly of opposite sign in the North Pacific (Plate 3). However, the PDO may also be distinct from the ENSO-like Pacific-wide decadal oscillation, as they appear dominated by 50 years and 20–30 years oscillations, respectively [*Minobe*, 2000; *Liu et al*., 2002]. On the other hand, *Tourre et al*. [2001] found two distinct decadal (9–12 years) and interdecadal (12–25 years) signals in the Pacific basin. Excepting this last study, the spectral peaks found on decadal and interdecadal timescales are not significantly different from red noise, and it is proper to refer to variability rather than oscillation for these timescales. In any case, the links between these tropical and/or Pacific-wide decadal-multidecadal variability and ENSO may be crucial, either through the modulation of the basic ENSO oscillation in the tropical Pacific or through their teleconnections [*Gershunov and Barnett*, 1998; *Alexander et al.*, this volume].

Decadal-multidecadal variability is difficult to comprehend from temporally limited data, and several authors have thus focused on the recent 1976 global climate shift [*Guilderson and Schrag*, 1998; *Zhang et al*., 1998; *Karspec and Cane*, 2002; *Giese et al*., 2002]. Its signature in the tropical Pacific is particularly important with a rapid increase of SST over the span of a year. This warming is associated with an increase in the amplitude and period of ENSO, and an eastward displacement along the equator of the maximum anomalies of SST gradient, westerly wind, and thermocline slope [*Wang and An*, 2002]. The origin of this warming and climate shift is still unclear. *Zhang et al*. [1998] suggest that subducted warm-water issued from the North Pacific perturbed the tropical thermocline (a hypothesis refuted by *Guilderson and Schrag* [1998]), while *Giese et al*. [2002] consider also a subsurface bridge but originating from the subtropical South Pacific. Note that other SST shifts in the last century may have occurred around 1924–25, 1941–42 and 1957–58 [*Chao et al*., 2000], most probably as phase transitions of several decadal-multidecadal oscillations [*Minobe*, 2000]. The dominance of the 1976 shift may be related to the acceleration of the 20th century warming trend observed in the tropical Pacific. *Knutson and Manabe* [1998] noted that this warming trend in a broad triangular region of the eastern tropical and subtropical Pacific increases from 0.41°C (100 yr)$^{-1}$ since 1900 to 2.9°C (100 years)$^{-1}$ since 1971. Like the Pacific-wide decadal mode, this warming trend has an El Niño-like structure.

5.2. Mechanisms of Tropical Pacific Decadal-Multidecadal Variability

As discussed above, both the tropical and mid-latitude Pacific show decadal-multidecadal variability. *Latif* [1998],

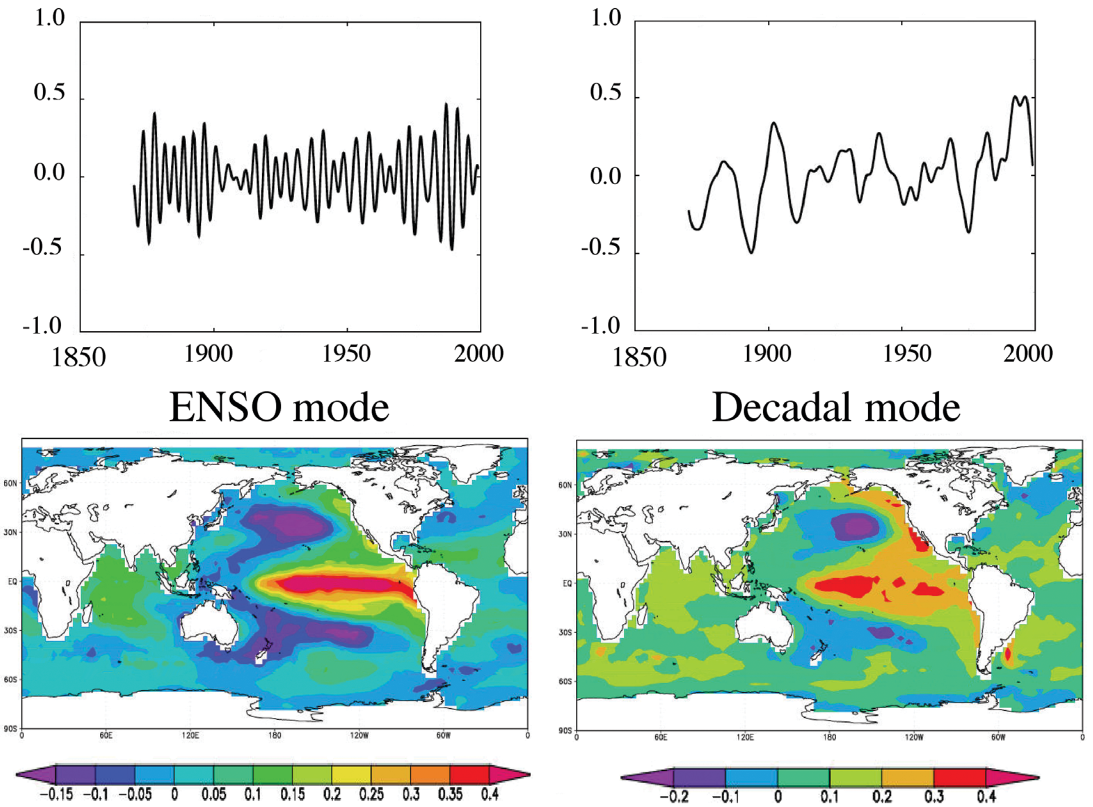

Plate 3. The two leading modes of Niño4 SST anomalies extracted through singular spectrum analysis. The upper panels represent the reconstruction of the Niño4 SST anomaly index using the two leading modes (20% of variance for ENSO mode and 25% for decadal mode); units are in [K]. The lower panels represent the regression patterns of SST anomalies using the upper two time-series as indices; units are [K]/standard deviation (courtesy of Mojib Latif).

Miller and Schneider [2000], *Minobe* [2000], and *Mantua and Hare* [2002] review mechanisms of North Pacific decadal-multidecadal variability. This subsection reviews the mechanisms of tropical Pacific decadal-multidecadal variability that can be divided into two categories: (1) tropical origins and (2) tropical-extratropical connections.

5.2.1. Tropical origins. Tropical Pacific decadal-multidecadal variability can be generated in the tropics only, without involving extratropical processes. Stochastic atmospheric forcing can lead to decadal-multidecadal variability in the tropical Pacific [e.g., *Kirtman and Schopf*, 1998; *Latif et al*., 1998; *Burgers*, 1999; *Thompson and Battisti*, 2001]. Using a simple model, *Wang et al*. [1999a] showed that tropical Pacific decadal-multidecadal variability might result from the nonlinear interactions between the seasonal and interannual cycles. *Timmermann et al.* [2003] hypothesized that ENSO variations can grow until they reach the maximum intensity of El Niño, then a quick reset takes place and small ENSO variations grow again. Due to the asymmetry of El Niño and La Niña, the decadal amplitude modulation of ENSO is translated into decadal background changes. Such nonlinear behaviors were found in coupled model simulations by *Timmermann* [2003] and *Rodgers et al.* [2004].

Linear dynamics and local ocean-atmosphere interaction can be at the origin of decadal variability in the tropical Pacific Ocean. Tropical local wind may force the decadal variability in the tropical Pacific Ocean [*Schneider et al*., 1999a; *Karspeck and Cane*, 2002]. The decadal changes in the background wind, before and after the 1976 climate shift, qualitatively reproduce the observed changes in ENSO properties noted above [*Wang and An*, 2002]. The origin of the changes in the winds is unclear, with a mid-latitude SST influence suggested by *Pierce et al*. [2000] and tropical ocean-atmosphere coupling proposed by *Liu et al*. [2002]. Yet, these last authors suggest that the decadal variability in the tropical Pacific can be enhanced by extratropical oceanic teleconnection. Using a coupled GCM, *Schneider* [2000] suggests an interesting decadal mode effective within the tropical Pacific, in which advection of salinity compensated temperature along isopycnals (termed spiciness anomalies) sets the decadal timescale.

Several authors have considered the inclusion of higher vertical and horizontal modes in the oceanic part of the ENSO delayed action oscillator to tentatively explain the decadal tropical variability, through wider ocean-atmosphere coupling, longer time in Rossby wave propagation and a reflected slow equatorial coupled wave. Using coupled models, *Knutson and Manabe* [1998], *Yukimoto et al*. [2000], and *Jin et al*. [2001] note westward phase propagations of decadal upper ocean temperature or thermocline depth around 9–12°N, 20°N and 15–25°N, respectively. Similar decadal propagating signals appear in observations [*White et al*., 2003] and in a model forced over the 1958–97 period [*Capotondi and Alexander*, 2001]. Similarly, the ENSO recharge oscillator has been amended to include off equatorial Rossby waves and eastern subtropical wind variability [*Jin*, 2001; *Wang et al.,* 2003]. This amendment is supported by the observation and simulation of decadal variability in upper-heat content in the tropical Pacific [*Hasegawa and Hanawa*, 2003b; *Alory and Delcroix*, 2002].

5.2.2. Tropical-extratropical connections. Gu and Philander [1997] consider an oceanic bridge that subducts and advects, in about 10 years, mid-latitude surface waters of anomalous temperature all the way to the Equatorial Undercurrent (EUC) via shallow subtropical cells (STCs) [*Schott et al*., this volume]. The anomalous waters are subsequently brought to the surface by equatorial upwelling and finally moved poleward by Ekman divergence [*Johnson*, 2001]. The circuit can be closed through this poleward surface oceanic bridge. It can also be closed through the upwelling-induced changes in eastern equatorial SST that influence the tropical and extratropical winds, which in turn affect the initial mid-latitude surface water anomalies. There is some evidence that anomalous surface water masses may subduct from the Northern and Southern Hemispheres toward the equator, using observations [*Deser et al*., 1996; *Zhang et al*., 1998; *Johnson and McPhaden*, 1999] and using models [*McCreary and Lu*, 1994; *Liu*, 1994; *Rothstein et al*., 1998; *Harper*, 2000; *Solomon et al*., 2003]. However, the detailed data analysis of *Schneider et al*. [1999a] does not find any significant decadal link between the North Pacific and the equator through anomalous subduction. Moreover, the temperature advected by the EUC is subject to strong seasonal and interannual variations that probably blur any remaining decadal signal [*Izumo et al*., 2002]. Finally, model studies [*Schneider et al*., 1999b; *Hazeleger et al*., 2001] indicate that decadal variability in the tropics is largely independent of the arrival of water anomalies subducted from the mid-latitudes. While *Gu and Philander* [1997] suggested anomalous temperature transported by STCs as a mechanism of tropical decadal variability, *Kleeman et al*. [1999] proposed that changes in STC strength vary the amount of cold water transported into the equatorial thermocline. This mechanism is supported by the observation of a slowdown of STCs from the PDO related regime shift of mid-1970s to the late 1990s together with a decrease in equatorial upwelling [*McPhaden and Zhang*, 2002]. The results of an OGCM forced by observed winds are also consistent with the mechanism of STC strength [*Nonaka et al*., 2002]. However, the two-year lag, found by these last authors between STC-induced and wind-forced equatorial SSTs, suggests that STC strength is not dominant in generating tropical decadal oscillations, acting more to amplify than to initiate them.

A number of the previous studies have focused on adiabatic oceanic processes. For example, in *Gu and Philander* [1997] the oceanic bridge is assured adiabatically through the subduction and advection of temperature anomalies. On the other hand, the surface forcing in the subduction region of the central North Pacific seems predominately diabatic [*Schneider et al*., 1999a]. The theoretical study of *Boccaletti et al.* [2004] indicates that basin-wide diabatic processes may control the tropical thermocline on decadal timescales, without involving explicit connection between the tropics and mid-latitudes. The relative importance of diabatic and adiabatic processes may depend on timescales. In a Pacific study using expendable bathythermograph data and an oceanic model over 1970–88, *Auad et al.* [1998] found that the 10-year variability is strongly influenced by adiabatic heat changes, while 20-year variability is more influenced by diabatic processes, such as direct thermal forcing by the atmosphere.

Another mechanism involves a wave signal exchanged between mid-latitude and the tropics. *Jacobs et al*. [1994] suggested that the 1982–83 El Niño could have decadal effects on the northwestern Pacific circulation, through mid-latitude Rossby waves reflected from equatorial Kelvin waves impinging on the American coasts. *Lysne et al*. [1997] found a weak decadal signal in their search for an oceanic bridge driven by wave dynamics: anomalous temperature is propagated by mid-latitude Rossby waves to the western boundary, then equatorward by coastal Kelvin waves, and finally modifies equatorial SST via equatorial Kelvin waves.

Since the PDO is one of the most important oscillations of decadal-multidecadal timescales on earth, it is an obvious candidate for forcing the tropical decadal variability through an atmospheric bridge. Moreover, on decadal timescale the largest SST anomalies and ocean heat content occur at the mid-latitude not in the tropics [*Giese and Carton*, 1999]. The decadal change in the northern atmosphere is wide enough to alter the wind stress over the equatorial Pacific, hence the mean state of the equatorial thermocline and upwelling, and ultimately ENSO activity [*Barnett et al*., 1999; *Pierce et al*., 2000; *Wang and An*, 2002]. The atmospheric bridge between the decadal variability of the North Pacific and ENSO may well be the imprint of a common internal variability in the atmosphere [*Pierce*, 2002]. In fact, *Wang and Weisberg* [1998] found that the out-of-phase SST decadal signal in the mid-latitudes and the tropics (Plate 3) is the result of a tropical-extratropical oscillation that involves feedbacks from the atmospheric Hadley and Walker circulations. As noted above, the PDO has a notable signature in the Southern Hemisphere, and the studies of *Luo and Yamagata* [2001] and *Luo et al.* [2003] illustrate the role of the South Pacific in sustaining tropical decadal variability through anomalous cyclonic or anticyclonic atmospheric circulation. The recent study of *Yu and Boer* [2004] on heat content anomalies forced in the western South Pacific by surface heat forcing and in the western North Pacific by ocean dynamics, underlines the combined role of the two hemispheres in generating Pacific decadal variability.

5.3. Interpretation of Tropical Pacific Warming Trends, Global Warming, and ENSO

Despite the observational difficulties in separating decadal variability from long-term trends, there is an undeniable acceleration of the warming trend over the last 50 years in the eastern tropical and subtropical Pacific [e.g., *Knutson and Manabe*, 1998] but its origins remain uncertain. Using SST and several atmospheric parameters, *Curtis and Hastenrath* [1999] find long-term trends in the tropical Pacific compatible with the radiative but not with the wind forcing. They also note the resemblance of these trends to El Niño patterns. *Liu and Huang* [2000] attribute the SST warming trend to the weakening trade wind, which reduces the vertical and zonal advection of cold water in the eastern and western equatorial Pacific, respectively. *Cane et al*. [1997] argue that the eastern equatorial Pacific has instead cooled since 1900, under increasing trade winds (difficulties in building historical surface fields and in particular correcting wind products are briefly discussed in the next sub-section). In fact, *Lau and Weng* [1999] found a secondary cooling trend centered near the Niño3 region, superimposed on a general warming trend.

Using global coupled ocean-atmosphere models, *Knutson and Manabe* [1998] did not find that the observed warming trends are quantitatively consistent in magnitude and duration with internal climate variability alone. They conclude that part of the warming trend may be attributed to sustained thermal forcing, such as greenhouse warming. In the same way, statistics on the Southern Oscillation behavior by *Trenberth and Hoar* [1996, 1997] open the possibility, challenged by *Harrison and Larkin* [1997] and *Rajagopalan et al*. [1997], that greenhouse gas is involved in the tendency for more frequent El Niño since the late 1970s. *Meehl and Washington* [1986] was one of the first model studies that looked at the changes within the tropics under an increase of atmospheric CO_2. Most coupled models in the late 1990s [e.g., *Meehl and Washington*, 1996; *Knutson and Manabe*, 1998; *Timmermann et al*., 1999] suggested that the eastern equatorial Pacific would warm more rapidly than the west. The SST gradient along the equator slackened together with the easterlies, and this results in an El Niño-like pattern of changes. Another school of studies suggests that the CO_2 warming response should be La Niña-like, with an increase of the equatorial SST gradient [*Cane et al*., 1997; *Seager and Murtugudde*, 1997] and maximum warming in mid-latitudes. There are a number of arguments for an El Niño-like pattern in response to global warming. Under this scenario the cloud-shielding thermostat over the warm pool [*Ramanathan and Collins*, 1991; *Meehl and Washington*, 1996] or the evaporative surface cooling [*Knutson and Manabe*, 1995] will make the warming less efficient in the west than the east, and the SST equatorial gradient will decrease. The warm pool can also expand toward the east, thus increasing the overlying atmospheric convection and westerly winds [*Yu and Boer*, 2002]. In the absence of ocean dynamics, the atmospheric response to global warming over the equatorial Pacific is a decrease of easterlies [*Vavrus and Liu*, 2002]. On the contrary, the La Niña-like pattern is due to equatorial upwelling that reduces the surface warming in the east, leading to an increased SST gradient along the equator and thus stronger easterlies. A coupled model forced by historical (1880–1990) and future greenhouse gas concentrations results in a warming trend, which initially is larger in the extra-tropics and has a La Niña-like pattern in the tropics [*Cai and Whetton*, 2000]. The pattern becomes El Niño-like after the 1960s and remains in this state during the 21st century. In this simulation, the shift (analogous to the observed 1976 shift) is explained by the delayed arrival of warm waters from the initial extra-tropical warming by STCs in the equatorial thermocline [*Cai and Whetton*, 2001b]. As noted by *Liu* [1998], any global warming study of the equatorial SST gradient must consider both the relatively rapid air-sea interaction process and the long-term thermocline process.

The plausible tendency for an El Niño-like pattern under global warming does not mean that the tropical Pacific will stay in a permanent El Niño. Superimposed on the new mean warm state, ENSO oscillations may continue but with probable changes in its behavior (possibly due to the change of the mean state). Using a low-resolution coupled model, *Knutson and Manabe* [1997] found a slight decrease in ENSO amplitude, no significant change in ENSO frequency and more pronounced multidecadal modulation of ENSO, in response to doubling or quadrupling of CO_2. With a finer model resolution, *Timmermann et al*. [1999] found more frequent El Niños and stronger La Niñas. *Collins* [2000a] had to quadruple the concentration of greenhouse gas in order to see ENSO changes. More frequent El Niños and La Niñas occur with 20% larger amplitude for both. Increases in meridional temperature gradients on either side of the equator and in the vertical gradient of temperature in the thermocline are respectively responsible for the increases of ENSO frequency and amplitude. The recent coupled model study of *Hu et al*. [2001] also results in an El Niño-like mean pattern but with stronger La Niñas and weaker El Niños. These incoherent results, which underline the complexity of coupled model behavior under greenhouse warming, are discussed briefly in the following.

5.4. Difficulties and Uncertainties of Studies of Low-Frequency Variability

A number of studies of low-frequency modulation of ENSO rely on the analyses of historical surface data (mostly SST, SLP, surface winds) that have been interpolated in time and space in a drastic way. It is recognized that global data coverage is adequate after 1950, if not after 1980 with the satellite era. Several research groups have built global products on a monthly basis and on a latitude-longitude grid that varies from 5° down to 1°. Most have extended the 1950 limit back to 100 years where volunteer observing ships were very rare, particularly in their journey through the equatorial and southern Pacific. For example, *Kaplan et al.* [1998] pointed out the discrepancies that arise from using several interpolated fields in the search of warming or cooling trends in the eastern tropical Pacific since 1900 (from – 0.3°C/100 years to + 0.3°C/100 years). A disputable assumption for building these products is the stability of their structural relations over the last one and a half century, i.e., stationarity. While these research groups are very aware of the errors associated with these fields, other (internet) users may forget to consider such errors in their analyses. Other articles on the low-frequency modulation of ENSO have been written solely on the Southern Oscillation Index that is extended to 1866. However, it appears difficult to imagine that decadal or multidecadal oscillations have not displaced the centers of action of the Southern Oscillation from Tahiti and/or Darwin.

Similarly, several articles discussing decadal variations of ENSO during the last millennium rely on either a single set of paleoclimate proxies over an extended period of time [e.g., *Linsley et al*., 2000] or several sets of proxies in the same location over separate periods of time [e.g., *Cobb et al*., 2003]. Paleoclimate proxies are typically diverse, involving tropical corals, mid-latitude tree rings and tropical or polar/subpolar ice cores. The labor, time and cost involved in data collection and processing render multi-proxy analyses of ENSO decadal variability uncommon as yet [e.g., *Evans et al*., 2001]. Multi-proxy reconstruction of ENSO time series suffers from technical and stability problems but also from assumptions about climate influence. For example, ice cores from tropical ice caps in Peru were originally thought to reflect ENSO variability, but recently they have been shown to be more affected by the large-scale atmospheric variability over Amazonian and the western tropical Atlantic than over the eastern tropical Pacific [*Thompson et al*., 2000]. In any case, the potential of paleoclimate indicators is tremendous for understanding the decadal variability and long-term trend of ENSO and thus for separating the natural contributions from the anthropogenic contribution [e.g., *Cole*, 2001].

Statistical tools are sometimes not adequate in extracting and explaining decadal and multidecadal ENSO signals, because of the shortness and uncertainty of the data and products or the difficulty in separating signals that have similar patterns (e.g., ENSO and ENSO-like), or the skewed nature of the signals. As noted by *Liu et al*. [2002], these tools are important for the diagnosis of decadal variability, but they may not be able to identify the true physical modes of variability (see also the series of papers pursuant to the article "A cautionary note on the interpretation of EOFs" by *Dommenget and Latif* [2002]).

Other difficulties in the search for decadal mechanisms arise from the use of observations. Oceanic bridges between the mid-latitudes and the equator are so far impossible to prove given the reduced number of hydrographical or CTDs observations in these regions over the last four decades. Delayed-type decadal oscillators cannot be truly established due to the complexity of extracting propagating signals from sparse data near the Swiss-cheese western boundary. As a consequence, many of the previous studies have used simplified or sophisticated models. Simplified models, such as those used by McCreary and collaborators (see *Schott et al.* [this volume]), have the great advantage of pinpointing mechanisms. Sophisticated ocean models suffer less from insufficient physics, but they are still unable to reproduce subsurface equatorial countercurrents and thus complete STC patterns. In any case, the use of variable forcing instead of seasonal forcing in models that simulate STCs may result in an open circuit rather than a closed circuit [*Fukumori et al*., 2004], hence making STC-induced decadal variability difficult or impossible to prove. Wind-forced models with or without data assimilation are also subject to spurious decadal variability and long-term trends due to unrealistic flux corrections, changes in the density and quality of data or measurement methods.

Coupled models have their own flaws. Many of them suffer from climate drifts, which are often corrected in dubious ways, and hardly reproduce realistic ENSO and annual cycles [*AchuaRao and Sperber*, 2002; *Davey et al*., 2002; *Latif et al.,* 2001]. Most of the simulated ENSOs are too close to a biennial cycle, or have weak amplitudes, or cannot reproduce the well-known horseshoe pattern of SST anomalies over the tropical Pacific (Plate 3). Hence, their ability to reproduce realistic decadal variability or warming trends in the tropical Pacific must be questioned. As a result, the projections of coupled models into the 21st century with and without CO_2 are as yet hard to believe. As an example, using version 2 of the Hadley Centre coupled model, *Collins* [2000a] found that the amplitude and frequency of ENSO increase with a quadrupling of CO_2. In version 3, *Collins* [2000b] attributed the lack of significant modification of ENSO behavior to subtle non-linear changes in the physical parametrization schemes, rather than the main differences between the two versions of the model (horizontal resolution and flux adjustments). A promising way to infer ENSO response to global warming is through international multi-model intercomparison projects such as CMIP [*AchutaRao and Sperber*, 2002], which use a quantitative probabilistic approach that takes into account model errors.

6. DISCUSSIONS AND THE FUTURE

As detailed above, significant advances have been made in understanding ENSO, especially during and after the TOGA decade. However, there are many issues or topics that that are still under debate and/or need to be further investigated. This section discusses some issues and questions that should probably be resolved and addressed in the near future.

In addition to the different origins and development of El Niño events as discussed in Section 2, El Niño events show some property changes during the last five decades. The changes include an increase of ENSO period and amplitude, changes in the propagation of interannual anomalies, and a zonal shift of maximum equatorial westerly wind anomalies [e.g., *Wang and An*, 2002; *Wittenberg*, 2004]. In spite of these differences, the equatorial westerly wind anomalies in the western Pacific preceding the warm SST anomalies in the eastern Pacific by 3–4 months is a common characteristic for all El Niño events. Some studies have attempted to understand some changes of El Niño characteristics or properties [e.g., *Fedorov and Philander*, 2001; *Wang and An*, 2002]. However, our understanding of El Niño property changes is poor. In particular, why El Niño events originate differently in the last five decades is not known yet.

The issue of ENSO as a self-sustained oscillation mode or a stable mode triggered by random forcing is not settled. *Philander and Fedorov* [2003] showed that the stability and period of ENSO depend on the mean states of the tropical Pacific Ocean-atmosphere system. It is possible that ENSO is a self-sustained 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 triggered by stochastic forcing than if ENSO is a self-sustained mode, because then its irregularity depends on random disturbances.

Since 1988, four concepts have been proposed for the oscillatory and self-sustained nature of ENSO. They also represent the negative feedbacks of a growing ENSO stable mode triggered by stochastic forcing, and are unified in a single concept. More data and model diagnoses are needed to test these concepts or to discover others. Several authors have extended three concepts of ENSO to explain the decadal variability in the tropical Pacific. Off-equatorial Rossby waves are at the root of the modified delayed action oscillator of *White et al.* [2003] and the recharge oscillator of *Jin* [2001] for decadal variability. *Yu and Boer* [2004] note the resemblance of the ENSO western Pacific oscillator of *Weisberg and Wang* [1997a] with their findings on decadal variability and heat content anomalies in the western North and South Pacific. A unified concept of decadal oscillator in the tropical Pacific has yet to be determined.

Since the seasonal cycle can greatly affect interannual variations, a critical feature for coupled ocean-atmosphere models is the ability to simulate both seasonal and interannual variations. The El Niño simulation intercomparison project (ENSIP) showed that almost all twenty-four coupled ocean-atmosphere models still have problems in simulating the SST climatology [*Latif et al*., 2001]. No model is able to simulate realistically all aspects of the interannual SST variability. Therefore, coupled model development and improvement are still major issues.

Interactions among intraseasonal variability (e.g., WWBs and MJO), interannual ENSO variability, and decadal-multi-decadal variability have recently received considerable attention. The unexpected intensity of the 1997–98 El Niño underlines the role of intraseasonal variability in the triggering and growth of El Niño. On the other hand, the two major El Niño events over the last two decades of the 20th century may be the signature of decadal variability or global warming. Research on these topics is still in an early stage, and it will likely need additions to the ENSO observing system. Intraseasonal studies will demand finer resolution (in time and space) and improved sampling aloft (in the atmosphere) over the tropical Indo-Pacific region, while decadal ENSO variability will demand wider (in latitude) and deeper (in the ocean) extensions together with better and more comprehensive paleoclimate proxies.

The relationship between ENSO and global warming is largely unknown. We are not even sure if greenhouse warming will result in an El Niño-like or La Niña-like pattern in the tropical Pacific. To understand the relationship between anthropogenic and natural climate variability, global coupled ocean-atmosphere models must be greatly improved and simulate both ENSO and the response to greenhouse warming.

Although it is an important topic of the CLIVAR-GOALS (Global Ocean Atmosphere Land System) program, ENSO is no longer the focus of major international process studies. Keeping in mind that one of the major successes of TOGA was the synergy among observationalists, theoreticians and modelers, every effort should be made to keep this synergy alive.

Despite great progress since the beginning of TOGA, understanding the mechanisms of ENSO is far from completed. In view of the rate of advancement in modeling and assimilation over the same time, it is expected that realistic simulation of ENSO will be feasible by the end of CLIVAR. This goal is linked for obvious socio-economic reasons to the improvement of ENSO prediction. However, one should never forget that understanding a fascinating puzzle of nature, like ENSO, must remain a key driver of future research.

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