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ENSO variability and the eastern tropical Pacific: A review

Chunzai Wang [a,*], Paul C. Fiedler [b]

[a] NOAA Atlantic Oceanographic and Meteorological Laboratory, Physical Oceanography Division, 4301 Rickenbacker Causeway, Miami, FL 33149, USA

[b] NOAA Southwest Fisheries Science Center, La Jolla, CA, USA

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Abstract

El Niño-Southern Oscillation (ENSO) encompasses variability in both the eastern and western tropical Pacific. During the warm phase of ENSO, the eastern tropical Pacific is characterized by equatorial positive sea surface temperature (SST) and negative sea level pressure (SLP) anomalies, while the western tropical Pacific is marked by off-equatorial negative SST and positive SLP anomalies. Corresponding to this distribution are equatorial westerly wind anomalies in the central Pacific and equatorial easterly wind anomalies in the far western Pacific. Occurrence of ENSO has been explained as either a self-sustained, naturally oscillatory mode of the coupled ocean–atmosphere system or a stable mode triggered by stochastic forcing. Whatever the case, ENSO involves the positive ocean–atmosphere feedback hypothesized by Bjerknes. After an El Niño reaches its mature phase, negative feedbacks are required to terminate growth of the mature El Niño anomalies in the central and eastern Pacific. Four requisite negative feedbacks have been proposed: reflected Kelvin waves at the ocean western boundary, a discharge process due to Sverdrup transport, western Pacific wind-forced Kelvin waves, and anomalous zonal advections. These negative feedbacks may work together for terminating El Niño, with their relative importance being time-dependent.

ENSO variability is most pronounced along the equator and the coast of Ecuador and Peru. However, the eastern tropical Pacific also includes a warm pool north of the equator where important variability occurs. Seasonally, ocean advection seems to play an important role for SST variations of the eastern Pacific warm pool. Interannual variability in the eastern Pacific warm pool may be largely due to a direct oceanic connection with the ENSO variability at the equator. Variations in temperature, stratification, insolation, and productivity associated with ENSO have implications for phytoplankton productivity and for fish, birds, and other organisms in the region. Long-term changes in ENSO variability may be occurring and are briefly discussed. This paper is part of a comprehensive review of the oceanography of the eastern tropical Pacific.

1. Introduction

Our first knowledge of El Niño came from Peruvian geographers, who at the end of the 19th century were interested in the unusual climate aberrations that occurred along the Peru coast in the odd year (Eguiguren,

* Corresponding author.
E-mail address: Chunzai.Wang@noaa.gov (C. Wang).

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1894). They took note of what a knowledgeable ship captain said about the fishermen in northern Peru, who typically saw a switch from cold to tropical ocean conditions around Christmas of every year and attributed this to a southward warm ''El Niño current''. This name was a reference to the annual celebration of the birth of the Christ child, who is much more prominent than Santa Claus/Saint Nicholas in Latin American traditions of the Christmas season. The 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 gradually spread through the world's scientific community that El Niño referred to the unusual events (Murphy, 1926; Lobell, 1942). The annual occurrence was forgotten.

It was separately noted by Sir Gilbert Walker in the 1920s and 1930s that notable climate anomalies occur around the world every few years, associated with what he called the Southern Oscillation (SO) (Walker, 1923, 1924, 1928; Walker and Bliss, 1932). The SO is a large interannual fluctuation in tropical sea level pressure (SLP) between the Western and Eastern Hemispheres (SO index is defined as SLP anomaly difference between Tahiti and Darwin). It was not until the 1960s that scientists came to realize that the warming off Peru is only part of an ocean-wide perturbation that extends westward along the equator out to the date line. Berlage (1957, 1966) recognized the linkage between the SO and episodic warmings of sea surface temperature (SST) along the coast of Southern Ecuador and Northern Peru, known locally as El Niño. El Niño became associated with unusually strong warmings that occur every two to seven years in concert with basin-scale tropical Pacific Ocean anomalies.

About the same time, the noted meteorologist Jacob Bjerknes proposed that El Niño was just the oceanic expression of a large-scale interaction between the ocean and the atmosphere. Using observed data in the context of earlier studies dating back to those of Walker (1924), Bjerknes (1966, 1969) provided evidence that the long-term persistence of climate anomalies associated with the Walker's SO (Walker and Bliss, 1932) is closely associated with slowly evolving SST anomalies in the equatorial eastern and central Pacific. Bjerknes recognized the importance of ocean–atmosphere interaction over the eastern tropical Pacific. He hypothesized that a positive ocean–atmosphere feedback involving the Walker circulation is responsible for the SST warming observed in the equatorial eastern and central Pacific. In his seminal paper, he stated (Bjerknes, 1969, p. 170):

> ''A decrease of the equatorial easterlies weakens the equatorial upwelling, thereby the eastern equatorial Pacific becomes warmer and supplies heat also to the atmosphere above it. This lessens the east–west temperature contrast within the Walker Circulation and makes that circulation slow down.''

This positive ocean–atmosphere feedback or coupled ocean–atmosphere instability leads the equatorial Pacific to a never-ending warm state. During that time, Bjerknes did not know what causes a turnabout from a warm phase to a cold phase (Bjerknes, 1969, p. 170):

> ''There is thus ample reason for a never-ending succession of alternating trends by air–sea interaction in the equatorial belt, but just how the turnabout between trends takes place is not quite clear.''

The positive ocean–atmosphere feedback of Bjerknes (1969) has influenced later studies. The essence of Bjerknes' hypothesis still stands as the basis of present day work. Oceanographers and meteorologists began to combine their efforts to expand and refine the Bjerknes' hypothesis by systematically studying the El Niño and the Southern Oscillation together in what we now call ''El Niño-Southern Oscillation'', or ENSO.

After Bjerknes published his hypothesis, ENSO was not intensively studied until the 1980s. The intense warm episode of the 1982–1983 El Niño, which was not recognized until it was well developed, galvanized the tropical climate research community to understand ENSO and ultimately predict ENSO. The 1982–1983 El Niño was not consistent with the ''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 characterized the evolution of El Niño (Rasmusson and Carpenter, 1982). This motivated the 10-year international TOGA (Tropical Ocean-Global Atmosphere) program (1985–94) to study and predict ENSO. One outcome was to build the ENSO observing system that includes the TAO/TRITON array of moored buoys (Hayes et al., 1991; McPhaden et al., 1998), an island tide-gauge network, surface drifters, the volunteer ship program, and various satellite observations. TOGA also supported analytical and diagnostic studies of the ENSO phenomenon (Wallace et al., 1998), and the development of a sequence of coupled ocean–atmosphere models to study and predict ENSO (Philander, 1990; Neelin et al., 1998; Wang and Picaut, 2004). Since TOGA, our understanding of ENSO has been greatly advanced by focusing on interaction between the tropical Pacific Ocean and atmosphere. This paper will provide a brief review of ENSO observations and of our present understanding of ENSO, with a focus on patterns and processes in the eastern tropical Pacific Ocean.

ENSO variability in the eastern tropical Pacific is centered along the equator, but is closely related to variability of the tropical Western Hemisphere warm pool (WHWP), which has been defined as the region covered by water warmer than 28.5 °C (Wang and Enfield, 2001, 2003). The WHWP is comprised of the eastern north Pacific west of Central America; the Intra-Americas Sea (IAS), i.e., the Gulf of Mexico and the Caribbean; and the western tropical North Atlantic. The WHWP is the second-largest tropical warm pool on Earth. Unlike the western Pacific warm pool in the Eastern Hemisphere, which straddles the equator, the WHWP is entirely north of the equator. The WHWP has a large seasonal cycle and the interannual fluctuations of its area are comparable to the annual variation, although it does not undergo large anomalous zonal excursions such as occur in the western Pacific. The WHWP is a critical component of the boreal summer climate of the Caribbean and surrounding land areas. From an oceanographic point of view, the WHWP can be separated into two parts by the Central American landmass: the eastern north Pacific warm pool and the Atlantic warm pool. To the atmosphere, the WHWP is a monolithic heat source that annually migrates and changes in size (Wang, 2002b), with little regard for the narrow landmass of Central America. We nevertheless recognize that WHWP development may involve oceanographic processes that are fundamentally different between the two oceans (Wang and Enfield, 2003). This paper will focus on review of seasonal and interannual variations of the eastern Pacific component of the WHWP, because of the eastern Pacific warm pool being part of the eastern tropical Pacific that is the subject of this review volume.

ENSO variability and the eastern Pacific warm pool are related to eastern tropical Pacific interdecadal variability reviewed by Mestas-Nuñez and Miller (2006), the ocean circulation of the eastern tropical Pacific by Kessler (2006), atmospheric forcing of the eastern tropical Pacific by Amador et al. (2006), and hydrography of the eastern tropical Pacific by Fiedler and Talley (2006). ENSO variability is associated with biological and ecological variability in the eastern tropical Pacific. We herein also briefly review biological and ecological effects of ENSO. The paper is organized as follows. Section 2 briefly describes major observed features of ENSO. Section 3 reviews our present understanding of ENSO. Section 4 presents seasonal and interannual variations of the eastern Pacific warm pool. Section 5 briefly reviews ENSO biological and ecological variability. Section 6 discusses changes in ENSO variability. Finally, Section 7 provides a summary.

2. Observations of ENSO

ENSO variability has been documented in the written record over hundreds of years (e.g., Quinn et al., 1987; Enfield, 1989). It is evident in paleoclimatic records, with slight changes in amplitude or frequency, over thousands of years (Diaz and Markgraf, 1992, 2000). For example, Rodbell et al. (1999) showed that the frequency of ENSO variability increased progressively over the period from about 7000–5000 years ago, and archaeological evidence suggests that El Niño events were either absent or very different from today for several millennia prior to that time (Sandweiss et al., 2001). This paper will not review ENSO variability based on paleoclimatic records. Instead, we will focus on ENSO variability from modern observational data.

Numerous observational studies had been published by the early 1980s describing the structure and evolution of ENSO (e.g., Trenberth, 1976; Weare et al., 1976; Quinn et al., 1978; Van Loon and Madden, 1981; Pazan and Meyers, 1982; Wooster and Guillen, 1974; Ramage and Hori, 1981; Weare, 1982; Wyrtki, 1975; Rasmusson and Carpenter, 1982). During the TOGA decade (1985–1994), more observational papers provided a much improved description and understanding of ENSO. McPhaden et al. (1998) and Wallace et al. (1998) have provided a comprehensive review of ENSO variability and structure from an observational point of view for the TOGA decade. In this section, we only briefly describe major observed features of ENSO variability.

The SST monthly anomalies in the Niño3 region (5° S–5° N, 150° W–90° W) are often used to index ENSO variability. Niño3 includes the eastern equatorial Pacific, west of the Galapagos, but no part of the eastern

North Pacific warm pool. From 1950–2003, there were eight significant El Niño warm events (1957–1958, 1965–1966, 1972–1973, 1982–1983, 1986–1987, 1991–1992, 1997–1998; and 2002–2003) and eight recognized La Niña cold events (1955–1956, 1964–1965, 1970–1971, 1973–1974, 1975–1976, 1984–1985, 1988–1989, and 1999–2000) (see Fig. 1a and b). The maximum (minimum) Niño3 SST anomalies for each warm (cold) event occur during the calendar months from November to January, except for the 1986–1987 El Niño event which has double peaks with the major one in the boreal summer. This indicates a robust tendency for the mature phase of El Niño to occur toward the end of the calendar year (e.g., Rasmusson and Carpenter, 1982; Fig. 1a); the peak phase of La Niña also occurs in the boreal winter (see Fig. 1b). That is, ENSO is phase-locked to the seasonal cycle. Because of this ENSO phase-locking to the seasonal cycle, oceanographers and meteorologists usually calculate and derive ENSO composites for better understanding of the evolving nature of ENSO.

The horizontal patterns of the tropical Pacific sea surface temperature (SST), sea level pressure (SLP), surface wind, and outgoing longwave radiation (OLR) anomalies during the mature phase of El Niño are shown in Fig. 2 (Wang et al., 1999), by compositing COADS data (Comprehensive Ocean–Atmosphere Data Set;

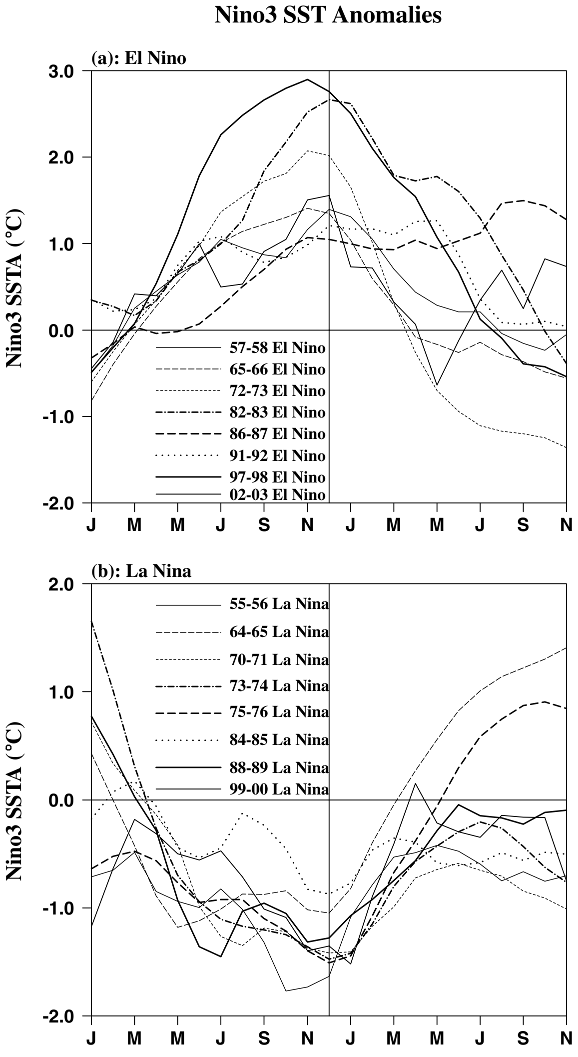

Fig. 1. SST anomalies in the Niño3 region (5° S–5° N, 150° W–90° W) from January of the ENSO development year to November the following year for (a) eight El Niño warm events and (b) eight La Niña cold events during 1950–2003. The data are from the NCEP monthly mean SST.

Woodruff et al., 1987) and NCEP OLR data. During the El Niño warm phase of ENSO, warm SST and low SLP anomalies are found in the eastern equatorial Pacific, and low OLR anomalies are in the central equatorial Pacific. Associated with the distributions of SST, SLP, and OLR anomalies, zonal wind anomalies are westerly in the central equatorial Pacific, indicating weakened easterly trade winds. During the mature and decay phases, the region of maximum westerly wind anomalies in the central/eastern Pacific is shifted to the south of the equator which may facilitate the El Niño decay (Harrison and Vecchi, 1999) since it causes a relaxation of westerly anomalies on the equator. ENSO is a basin-scale phenomenon, so it also shows western Pacific patterns in addition to eastern Pacific patterns. During the mature phase of El Niño, when the warmest SST anomalies are in the eastern equatorial Pacific, the coldest SST anomalies are located to the north and south of the equator in the western Pacific, instead of on the equator. Since atmospheric convection over the western Pacific warm pool shifts into the central equatorial Pacific during the warm phase of ENSO, the region of the lowest OLR anomalies is located to the west of the warmest SST anomalies. Similar to the zonal offset of SST and OLR anomalies in the equatorial eastern and central Pacific, in the western Pacific the off-equatorial region of highest OLR anomalies is positioned west of the off-equatorial region of coldest SST anomalies. The off-equatorial western Pacific cold SST anomalies are also accompanied by off-equatorial western Pacific high SLP anomalies. As a result of off-equatorial high SLP anomalies, easterly wind anomalies appear in the far equatorial western Pacific, as shown in Fig. 2c. Thus, during the mature phase of El Niño, the equatorial eastern Pacific shows warm SST and low SLP anomalies, and the equatorial central Pacific shows low OLR anomalies, while the off-equatorial western Pacific shows cold SST and high SLP anomalies, and the off-equatorial far western Pacific shows high OLR anomalies. Associated with these SST, SLP, and OLR anomaly patterns are equatorial westerly wind anomalies in the central Pacific and equatorial easterly wind anomalies in the far western Pacific. The nearly out-of-phase behavior between the eastern and western tropical Pacific is also observed during the mature phase of La Niña, but with anomalies of opposite sign

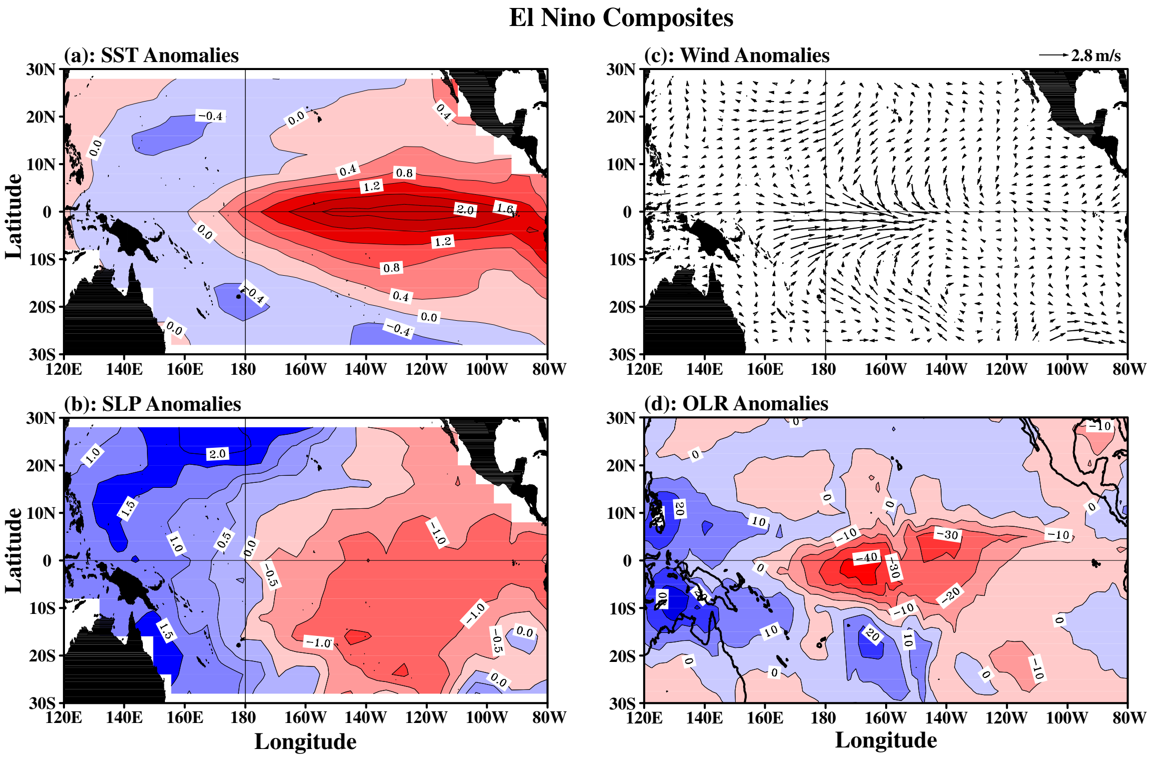

Fig. 2. El Niño composites of (a) sea surface temperature (SST) anomalies (°C), (b) sea level pressure (SLP) anomalies (mb), (c) surface wind anomalies (m s^{-1}), and (d) out-going longwave radiation (OLR) anomalies (W m^{-2}).

(Fig. 3). Figs. 1–3 also show asymmetry between El Niño and La Niña, with anomalies of El Niño larger than those of La Niña.

The evolution of El Niño and La Niña can be seen in the SST, zonal wind, and 20 °C isotherm depth (a proxy for thermocline depth) anomalies of the TAO/TRITON array data along the equator from January 1986 to August 2003 (Fig. 4). The TAO/TRITON moored data show warm events occurring in 1986–1987, 1991–1992, 1997–1998, and 2002–2003, and cold events in 1988–1989, 1995–1996, and 1999–2000. There is a close relationship between zonal wind anomalies in the western Pacific and thermocline depth anomalies in the central and eastern Pacific (e.g., Kessler et al., 1995). Fig. 4 shows that zonal wind fluctuations in the equatorial western Pacific correspond to 20 °C isotherm depth signals propagating eastward across the basin at Kelvin wave-like speeds. Remote forcing (from the western Pacific) is clearly important for thermocline and SST anomalies in the central and eastern Pacific (e.g., Kessler and McPhaden, 1995).

The importance of the western Pacific can also be seen by comparing the Niño3 SST anomalies with the indices of the 850-mb zonal wind anomalies in the equatorial western Pacific (5° S–5° N, 120° E–170° E) and in the equatorial eastern Pacific (5° S–5° N, 150° W–100° W) from the NCEP-NCAR reanalysis field, as shown in Fig. 5 (Wang, 2002a). The maximum correlation of 0.56 occurs when the western Pacific zonal wind anomalies lead the Niño3 SST anomalies by four months, whereas the maximum correlation between the eastern Pacific zonal wind anomalies and the Niño3 SST anomalies is 0.67 at zero month lag. The correlation relations suggest that the eastern equatorial Pacific is a location of strong ocean–atmosphere interaction (Battisti and Hirst, 1989; Cane et al., 1990) and that the western Pacific is an important region for initiating and terminating El Niño (e.g., McCreary, 1976; Busalacchi and O'Brien, 1981; Philander, 1981; Tang and Weisberg, 1984; Philander, 1985; Wang et al., 1999; McPhaden and Yu, 1999; Wang and Weisberg, 2000; Boulanger and Menkes, 2001; Wang, 2001a; Picaut et al., 2002; Boulanger et al., 2003). In addition, the salinity-stratified barrier layer in the western Pacific is hypothesized to influence the development of El Niño (Maes et al., 2002).

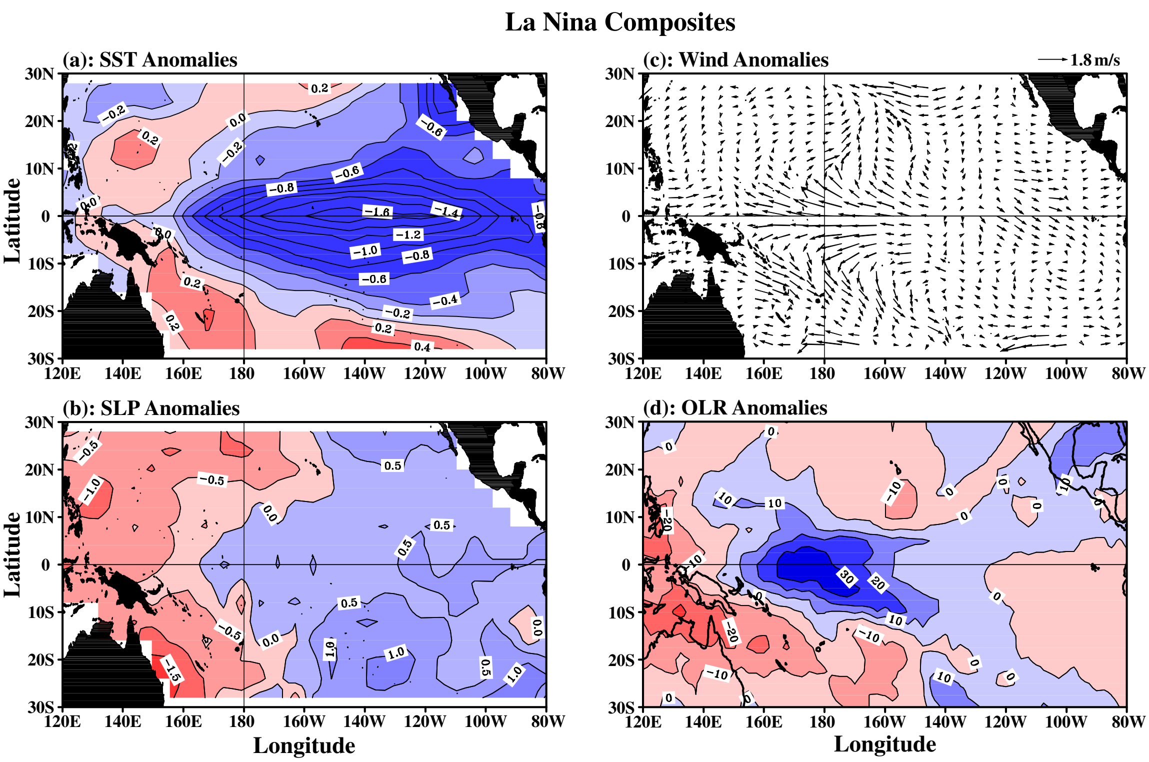

Fig. 3. La Niña composites of (a) sea surface temperature (SST) anomalies (°C), (b) sea level pressure (SLP) anomalies (mb), (c) surface wind anomalies (m s^{-1}), and (d) out-going longwave radiation (OLR) anomalies (W m^{-2}).

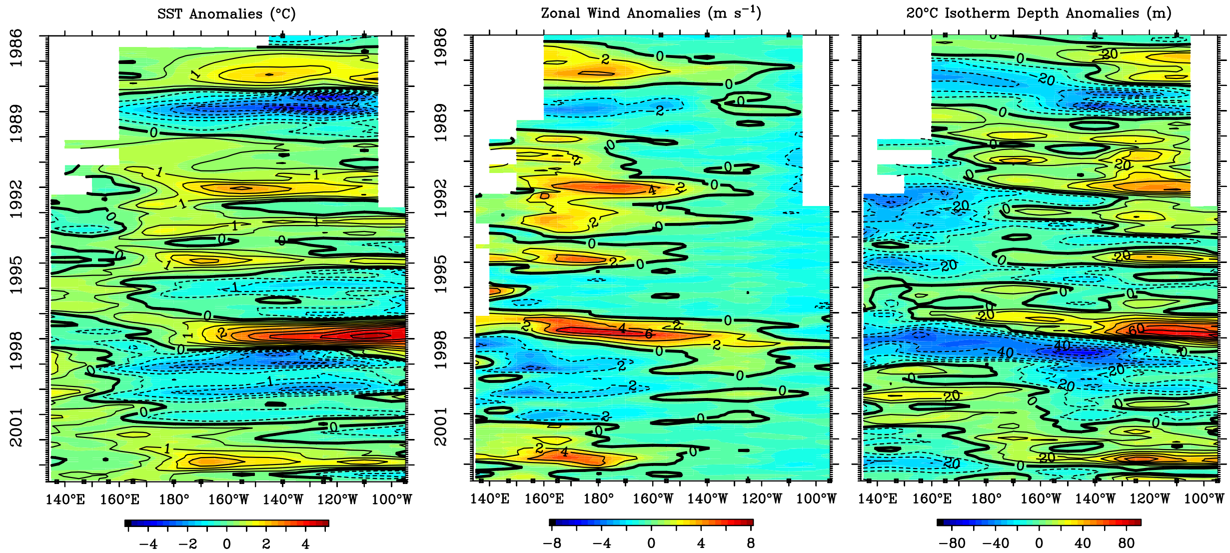

Fig. 4. Time-longitude sections of monthly SST, zonal wind, and 20 °C isotherm depth anomalies between 2°S to 2° N from January 1986 to August 2003. The data are provided by TAO/TRITON project office.

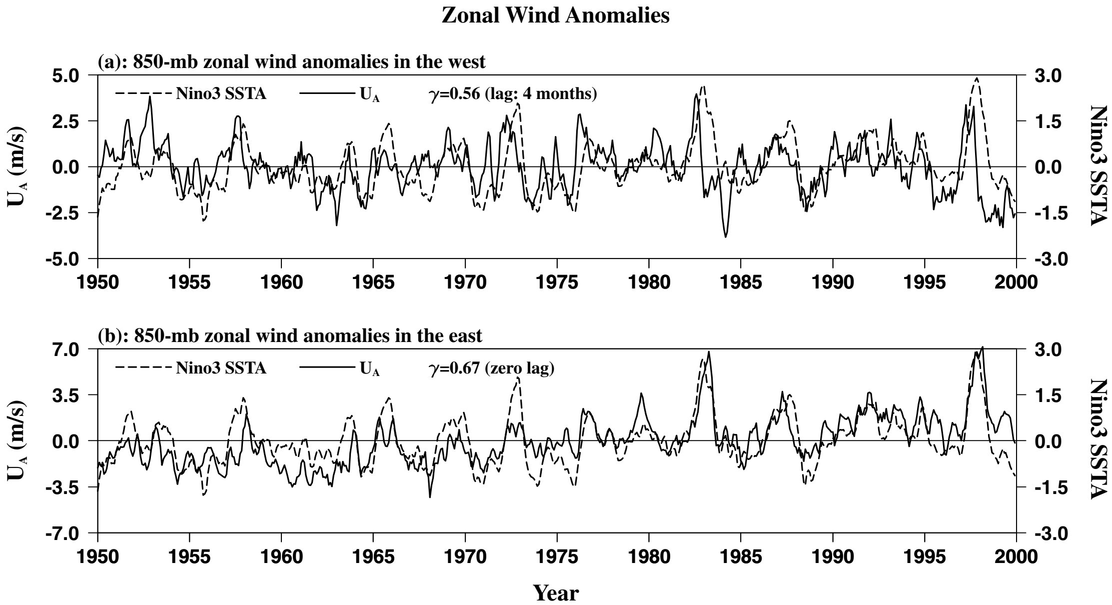

Fig. 5. Comparison of the Niño3 SST anomalies with (a) 850-mb zonal wind anomalies in the equatorial western Pacific (5° S–5° N, 120 °E–170 °E), (b) 850-mb zonal wind anomalies in the equatorial eastern Pacific (5° S-5° N, 150° W–100° W). SST data are from the NCEP SST and winds are from the NCEP-NCAR reanalysis. All of the time series are three-month running means. The γ represents correlation coefficient.

By studying the El Niño events between 1949 and 1976, Rasmusson and Carpenter (1982) showed that the SST anomalies along the South American coast reached peak warming in the boreal spring of the El Niño year. However, El Niño events have evolved differently after 1977 (e.g., Wang, 1995; Trenberth and Stepaniak, 2001). The coastal warmings for the El Niño events between 1977 and 1996 occurred in the boreal spring subsequent to the El Niño year rather than in the boreal spring of the El Niño year (Wang, 1995). The 1997–98 El Niño developed in both the central Pacific and the South American coast during the spring of

1997 (Wang and Weisberg, 2000). Fig. 6 shows different origins and development for the El Niños between 1950–76, the El Niños between 1977 and 1996, the 1997–98 El Niño, and the 2002–2003 El Niño (Wang and Picaut, 2004). The coastal warming over South America appeared in March–May for El Niños before 1976, but not for the El Niños between 1977 and 1996. For the 1997–1998 El Niño, the initial warming occurred in both the equatorial central Pacific and along the South American coast. The 2002–2003 El Niño started and remained in the equatorial central Pacific. Why El Niños originated in varied ways in the last five decades is not understood yet, although it may be related to high- or low-frequency variability.

ENSO is an interannual phenomenon that superimposes on the seasonal variability. A rough partitioning of global monthly SST variance is illustrated in Fig. 7 (also see Delcroix, 1993; Fiedler, 2002a). Seasonal (1–12 months) variability is high in the eastern boundary currents, the equatorial cold tongue, and at high latitudes. Seasonal variability is low in the equatorial Pacific west of the equatorial cold tongue, in the Gulf of Panama, and in a band north of the cold tongue corresponding to the thermal equator (Fiedler, 2002b; Fiedler and Talley, 2006). Interannual (ENSO) timescale (1–10 years) variability is high along the equator, and along coastal Ecuador, Peru, and Baja California. Interannual variability is low in the subtropical gyres and in the eastern Pacific warm pool. Interannual variance exceeds seasonal variance between 10° S and 10° N west of 110° W, and also east of the Galapagos. Interannual variability is low in the Atlantic side of the Western Hemisphere warm pool (WHWP), both in absolute magnitude and in relation to seasonal variability.

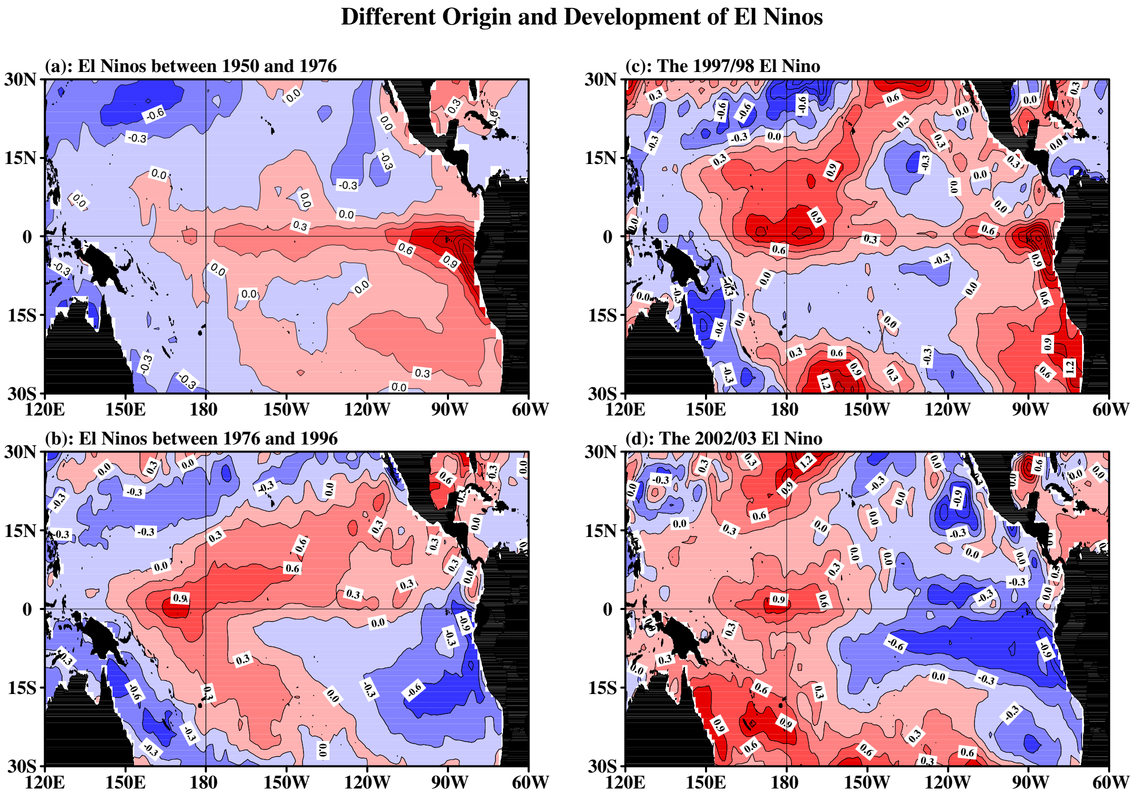

Fig. 6. SST anomaly composites showing different origin and development of El Niños in the last five decades. (a) El Niños between 1950 and 1976, (b) El Niños between 1977 and 1996, (c) the 1997–1998 El Niño, and (d) the 2002–2003 El Niño. The composites are calculated by averaging the SST anomalies during the development phase of El Niño (March–May of the El Niño year). Since the 2002–2003 El Niño starts earlier, its composite is used the SST anomalies of December 2001 to February 2002.

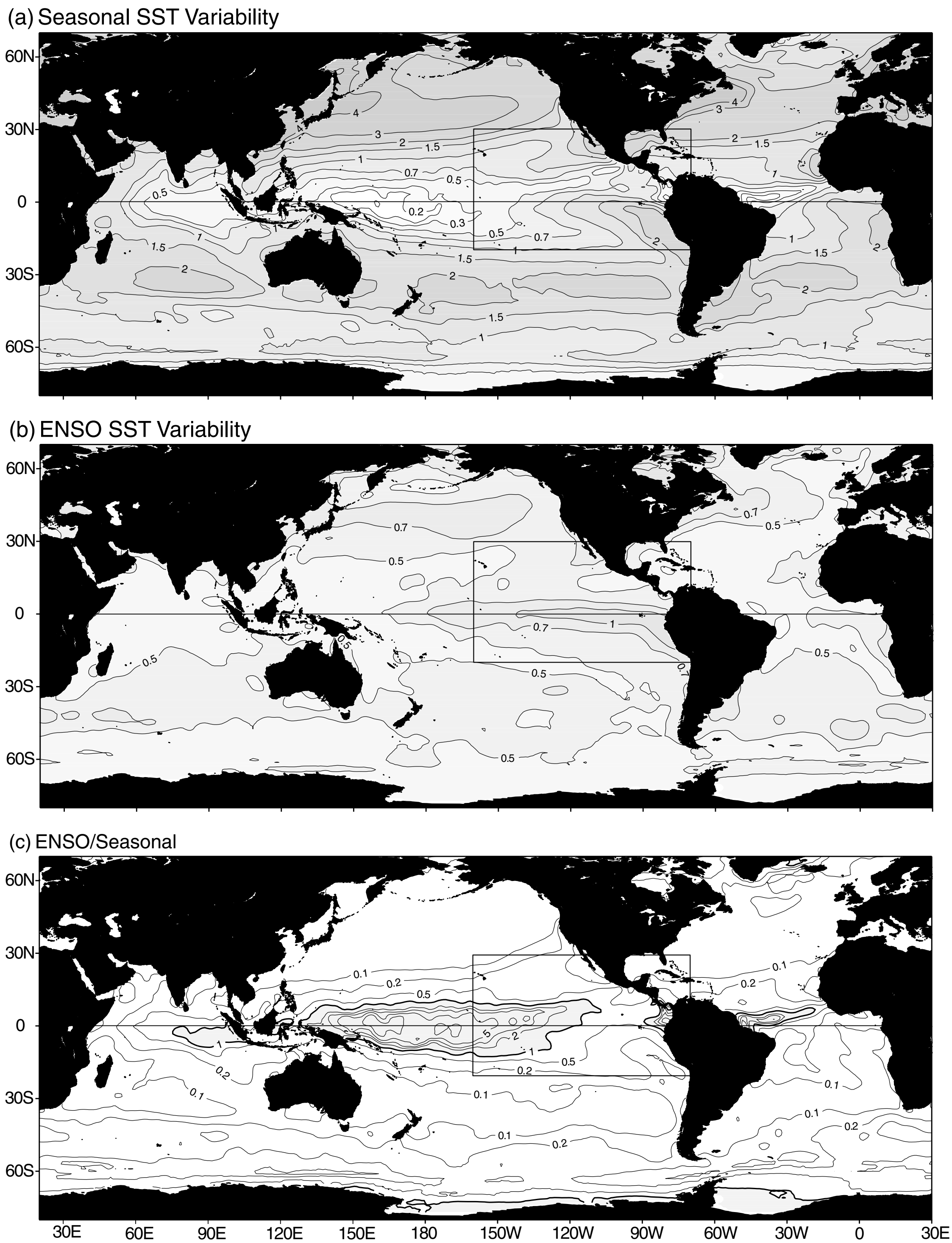

Fig. 7. Monthly global SST variability (1950–2002) from GISST data set (UK Met Office – Global sea Ice Coverage and Sea Surface Temperature v2.3b, Parker et al., 1995). (a) Intra-annual or seasonal variability, estimated as sqrt{Var(SST) – Var(SSTA)}, (b) ENSO variability, estimated as sqrt{Var(SSTA)} or the standard deviation of monthly SST anomalies, (c) Ratio of interannual (ENSO) to seasonal variance. Variability on scales greater than 10 years was removed by subtracting a 10-year running mean from the monthly time series (SST).

3. Mechanisms of ENSO

The eastern tropical Pacific is a region that can both involve local ocean–atmosphere interaction and be remotely affected by processes in the western Pacific, because the absence of a Coriolis effect causes the equatorial ocean to act as a waveguide (Gill, 1982). Bjerknes (1969) first hypothesized that interaction between the atmosphere and the equatorial eastern Pacific Ocean causes El Niño. In Bjerknes' view 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 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 never-ending warm state. A negative feedback is needed to turn the coupled ocean–atmosphere system around. However, during that time, it was unknown what causes a turnabout from a warm phase to a cold phase, which is commonly referred to as La Niña (Philander, 1985, 1990).

Wyrtki (1975) noticed that the local southeast winds off Peru are not abnormally weak during times of El Niño (they can even be stronger, which are upwelling-favorable), but the thermocline deepens and SST warms along the South American coast. This led him to search for other ways to explain El Niño. Using the equatorial wind and sea level data, Wyrtki (1975) proposed a "buildup" mechanism for El Niño. He suggested that prior to El Niño, the trade winds along the equator strengthened, and there was a "buildup" in heat content and a consequent rise in sea level in the western Pacific warm pool. A trigger for El Niño is a rapid collapse of the easterly trade wind, which remained unexplained since Wyrtki (1975) was not addressing the coupled problem. When the easterly trade wind rapidly decreases, the accumulated warm water in the western Pacific would collapse and surge eastward in the form of Kelvin waves to initiate a warm El Niño event. He concluded that El Niño warming along the South American coast is not the result of local change in surface wind forcing, but rather reflects a remote response to a rapid decrease in the easterly trade wind. Wyrtki's mechanism addressed the question of the onset of the coastal El Niño warming over South America and emphasized that an El Niño is forced by an abrupt change in zonal wind stress, differing from Bjerknes' (1969) hypothesis of a slowly evolving warming due to ocean–atmosphere interaction. Wyrtki's idea was supported by simple model studies of Hurlburt et al. (1976), McCreary (1976), and Busalacchi and O'Brien (1981). However, Wyrtki's "buildup" mechanism is not consistent with the development of the 1982–1983 El Niño that showed no rise in sea level in the western Pacific and no intensification of the easterly trade winds before 1982, and no initial warming off the west coast of South America in 1982 (e.g., Fig. 6).

Failure to recognize the 1982–1983 El Niño (the strongest in over a hundred years) until it was well developed motivated the scientific community to intensively study ENSO. Since the 1980s, our understanding of ENSO has been greatly advanced (e.g., see ENSO reviews of Philander, 1990; Neelin et al., 1998; Wang and Picaut, 2004). The theoretical explanations of El Niño can be summarized as two views. First, El Niño is one phase of a self-sustained, naturally oscillatory mode of the coupled ocean–atmosphere system. Second, El Niño is a stable (damped) mode triggered by stochastic forcing (e.g., random "noise"). Whatever the case, El Niño begins 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 growth of the mature El Niño anomalies in the central and eastern Pacific. That is, negative feedbacks are needed to turn the coupled system from a warm phase to a cold phase. In search of necessary negative feedbacks for the coupled system, four conceptual ENSO oscillator models have been proposed: the delayed oscillator (Suarez and Schopf, 1988; Graham and White, 1988; Battisti and Hirst, 1989; Cane et al., 1990), the recharge oscillator (Jin, 1997a,b), the western Pacific oscillator (Weisberg and Wang, 1997b; Wang et al., 1999), and the advective-reflective oscillator (Picaut et al., 1997). These oscillator models respectively emphasized the negative feedbacks of reflected Kelvin waves at the ocean western boundary, a discharge process due to Sverdrup transport, western Pacific wind-forced Kelvin waves, and anomalous zonal advection. These negative feedbacks may work together for terminating El Niño warming, as suggested by the unified oscillator (Wang, 2001a).

3.1. The delayed oscillator

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

Graham and White (1988) presented observational evidence of off-equatorial Rossby waves and their reflection at the western boundary and then empirically constructed a conceptual oscillator model for ENSO. As shown in McCreary and Anderson (1991), the conceptual equations of the Graham and White model can be reduced to a single equation that has similar form to the delayed oscillator (also see the comments of Neelin et al. (1998)). The schematic diagram of the delayed oscillator is shown in Fig. 8. The positive ocean–atmosphere feedback occurs in the equatorial eastern Pacific (e.g., in the Niño3 region), leading the Niño3 SST anomaly to a warm state. The delayed negative feedback is by free Rossby waves generated in the eastern Pacific coupling region that propagate to and reflect from the western boundary, returning as Kelvin waves to reverse the Niño3 SST anomalies in the eastern Pacific coupling region. The delayed oscillator assumes that the western Pacific is an inactive region for air–sea interaction and that ocean eastern boundary wave reflection is unimportant, emphasizing the importance of wave reflection at the western ocean boundary. The role of the eastern tropical Pacific in this model is initiation of an El Niño by air–sea coupling.

3.2. The recharge oscillator

Wyrtki (1975) first suggested a buildup in the western Pacific of warm water as a necessary precondition to the development of El Niño. This concept was later modified to involve 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 charge) gradually, and during El Niño warm water is flushed toward (or discharged to) higher latitudes. After the discharge, the tropical Pacific becomes cold (La Niña) and then warm water slowly builds up again (recharge) before occurrence of next El Niño. The recharge and discharge processes have also been examined by Cane et al. (1986), Zebiak (1989), Miller and Cheney (1990), and Springer et al. (1990). Holland and Mitchum (2003) showed that interannual volume redistribution within the tropical Pacific is much greater than volume loss for the tropical Pacific as a whole.

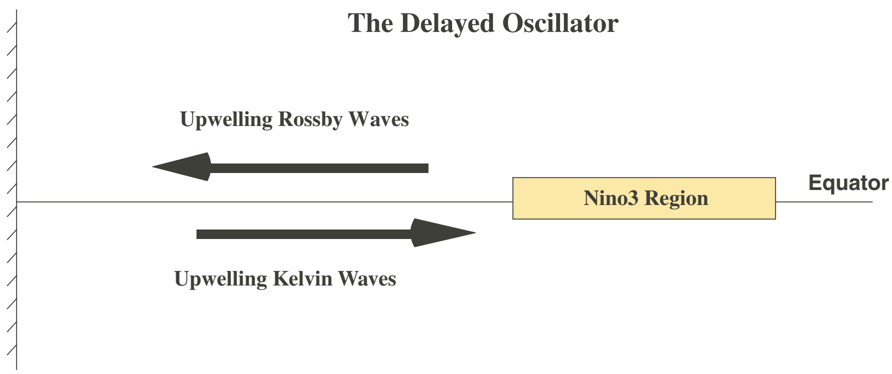

Fig. 8. Schematic diagram of the delayed oscillator for ENSO. The delayed oscillator model has a positive feedback and a negative feedback, assuming that the western Pacific is inactive for air–sea interaction and wave reflection in the ocean eastern boundary is unimportant. The positive feedback is represented by local ocean–atmosphere coupling in the equatorial eastern Pacific (for example, in the Niño3 region). The delayed negative feedback is represented by free Rossby waves generated in the eastern Pacific coupling region that propagate to and reflect from the western boundary, returning as Kelvin waves to reverse the anomalies in the eastern Pacific coupling region. Thus, the coupled ocean–atmosphere system oscillates on interannual time scales.

The concept of the recharge and discharge processes is further emphasized by Jin (1997a,b). Based on the coupled model of Zebiak and Cane (1987), Jin (1997a,b) formulated and derived the recharge oscillator model. 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 (Fig. 9). 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, leading to the cold phase. The converse occurs during the cold phase of ENSO. It is the recharge–discharge process that makes the coupled ocean–atmosphere system oscillate on interannual time scales. The role of the eastern tropical Pacific in this model is primarily as one extreme of the tropical Pacific redistribution of heat and mass.

3.3. The western Pacific oscillator

Observations show that ENSO displays both eastern and western Pacific interannual anomaly patterns (e.g., Rasmusson and Carpenter, 1982; Mayer and Weisberg, 1998; Wang et al., 1999; Wang and Weisberg, 2000; Figs. 2 and 3). During the warm phase of ENSO, warm SST anomalies in the equatorial eastern Pacific are accompanied by cold SST and shallow thermocline depth anomalies in the off-equatorial western Pacific. Also, while the zonal wind anomalies over the equatorial central Pacific are westerly, those over the equatorial western Pacific are easterly. Consistent with these observations, Weisberg and Wang (1997b) developed a conceptual western Pacific oscillator model for ENSO. This model emphasizes the role of the western Pacific in ENSO that has been overlooked in the delayed oscillator. In particular, off-equatorial SST anomalies (and off-

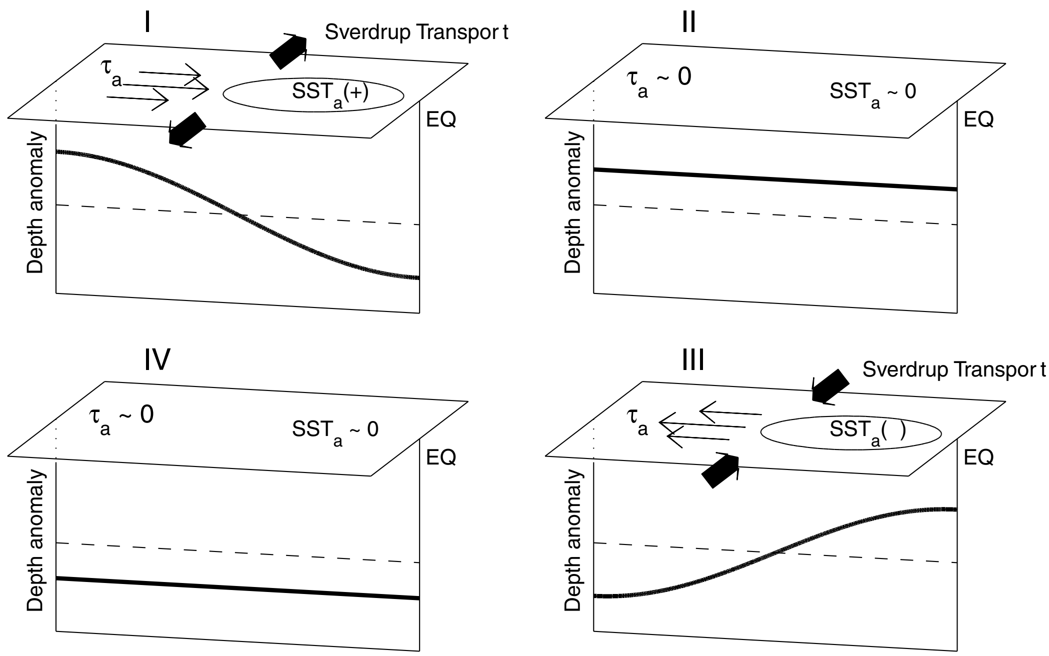

Fig. 9. Schematic diagram of the recharge oscillator for ENSO. The four phases of the recharge oscillation are: (I) the warm phase, (II) the warm to cold transition phase, (III) the cold phase, and (IV) the cold to warm transition phase. During the warm phase of ENSO, the divergence of Sverdrup transport 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 in the transition phase allows anomalously 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.

equatorial anomalous anticyclones) in the western Pacific induce equatorial western Pacific wind anomalies that affect the evolution of ENSO.

Arguing from the vantage point of a Gill (1980) atmosphere, condensation heating due to convection in the equatorial central Pacific (Deser and Wallace, 1990; Zebiak, 1990; Weisberg and Wang, 1997a) induces a pair of off-equatorial cyclones with westerly wind anomalies on the equator (Fig. 10). These equatorial westerly wind anomalies act to deepen the thermocline and increase SST in the equatorial eastern Pacific, thereby providing a positive feedback for anomaly growth. On the other hand, the off-equatorial cyclones raise the thermocline there via Ekman pumping. Thus, a shallow off-equatorial thermocline anomaly expands over the western Pacific leading to a decrease in SST and an increase in SLP in the off-equatorial western Pacific (e.g., Wang et al., 1999; Wang, 2000). During the mature phase of El Niño, the off-equatorial anomalous anticyclones initiate equatorial easterly wind anomalies in the western Pacific. These equatorial easterly wind anomalies cause upwelling and cooling that proceed eastward as a forced ocean response providing a negative feedback, allowing the coupled ocean–atmosphere system to oscillate. The role of the western Pacific wind-forced Kelvin waves in terminating ENSO variability in the eastern Pacific has been demonstrated by McPhaden and Yu (1999), Delcroix et al. (2000), Boulanger and Menkes (2001), Vialard et al. (2001), Picaut et al. (2002), and Boulanger et al. (2003). As in other oscillator models, the eastern tropical Pacific plays a role in this model of ENSO by supporting a positive ocean–atmosphere feedback.

3.4. The advective-reflective oscillator

Picaut et al. (1996) found that zonal displacement of the oceanic convergence zone at the eastern edge of the western Pacific warm pool is in phase with the Southern Oscillation Index. Based on this finding and the study of Picaut and Delcroix (1995) regarding wave reflection, Picaut et al. (1997) proposed a conceptual advective-reflective oscillator model for ENSO (Fig. 11). In this conceptual model, 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 of the western Pacific are: anomalous zonal current associated with wave reflection at the western boundary, anomalous zonal current associated with

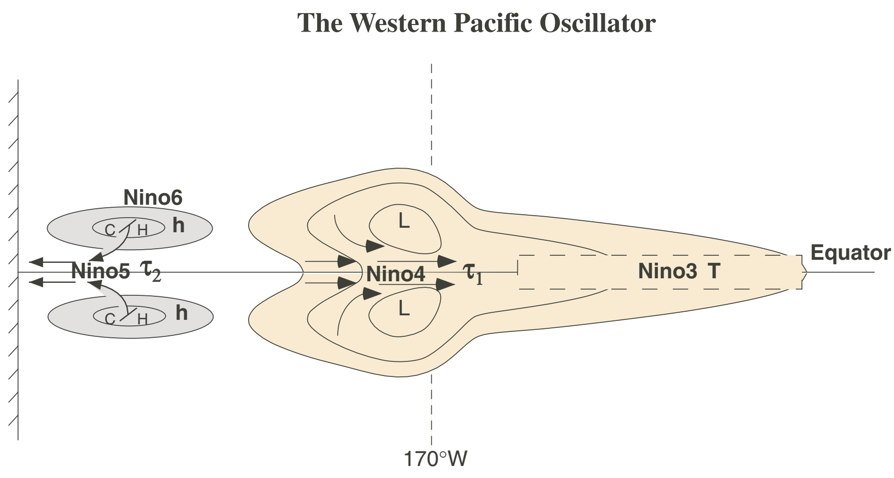

Fig. 10. Schematic diagram of the western Pacific oscillator for ENSO. This oscillator emphasizes the role of the western Pacific in ENSO which has been overlooked in the delayed oscillator. Arguing from the vantage point of a Gill atmosphere, condensation heating in the central Pacific induces a pair 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, thereby providing a positive feedback for anomaly growth. On the other hand, the off-equatorial cyclones raise the thermocline there via Ekman pumping. Thus, a shallow off-equatorial thermocline anomaly expands over the western Pacific leading to a decrease in SST and an increase in SLP in the Niño6 region. During the mature phase of El Niño, the Niño6 high SLP initiates equatorial easterly wind anomalies in the Niño5 region. The Niño5 easterly wind anomalies cause upwelling and cooling that proceed eastward as a forced ocean response providing a negative feedback for the coupled ocean–atmosphere system to oscillate on interannual time scales.

The Advective-Reflective Oscillator

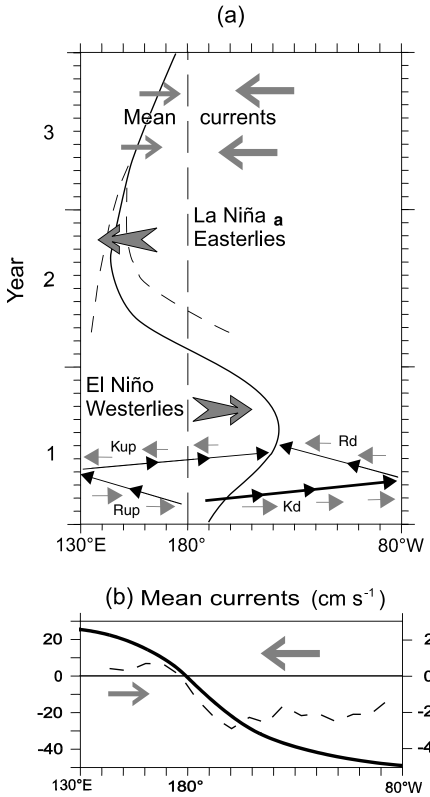

Fig. 11. Schematic diagram of the advective-reflective oscillator for ENSO. In (a), the thick line represents the eastern edge of the western Pacific warm pool. In (b), the dashed line shows observational mean zonal current between 2° N and 2° S, and the thick line is an idealized zonal current. This 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. Equatorial westerly wind anomalies in the central Pacific produce upwelling Rossby (downwelling Kelvin) waves that propagate westward (eastward). The upwelling Rossby (downwelling Kelvin) waves reflect to upwelling Kelvin (downwelling Rossby) waves after they reach the western (eastern) boundary. Since both the upwelling Kelvin and downwelling Rossby waves have westward zonal currents, they tend to push the warm pool back to its original position of the western Pacific. These negative feedbacks along with the negative feedback of the mean zonal current cause the coupled ocean–atmosphere system to oscillate.

wave reflection at the eastern boundary, and mean zonal current converging at the eastern edge of the warm pool. During the warm phase of ENSO, equatorial westerly wind anomalies in the central Pacific produce upwelling Rossby and downwelling Kelvin waves that propagate westward and eastward, respectively. The westward propagating upwelling Rossby waves reflect as upwelling Kelvin waves after they reach the western boundary, whereas the eastward propagating downwelling Kelvin waves reflect as downwelling Rossby waves at the eastern boundary. Since both the upwelling Kelvin and downwelling Rossby waves have westward zonal currents, they tend to push the warm pool back to its original position in the western Pacific. These negative feedbacks along with the negative feedback of the mean zonal current, force the coupled ocean–atmosphere system to oscillate. The role of the eastern tropical Pacific in this model is reflection of downwelling Rossby waves as one component of this negative feedback.

3.5. The unified oscillator

With several different conceptual oscillator models capable of producing ENSO-like oscillations, more than one may operate in nature. Motivated by existence of the above oscillator models, Wang (2001a,b) formulated and derived a unified ENSO oscillator model from the dynamics and thermodynamics of the coupled ocean–atmosphere system that is similar to the Zebiak and Cane (1987) coupled model. Since ENSO is observed to show both eastern and western Pacific anomaly patterns, this oscillator model is formulated and constructed to consider SST anomalies in the equatorial eastern Pacific, zonal wind stress anomalies in the equatorial central Pacific, thermocline depth anomalies in the off-equatorial western Pacific, and zonal wind stress anomalies in the equatorial western Pacific. This model can oscillate on interannual time scales. The unified oscillator includes the physics of all oscillator models discussed above (Fig. 12). All of the above ENSO oscillator models are special cases of the unified oscillator. As suggested by the unified oscillator, ENSO is a multi-mechanism phenomenon (see Picaut et al. (2002) for observations of different ENSO mechanisms) and the relative importance of different mechanisms is time-dependent.

3.6. A stable mode triggered by stochastic forcing

Another view of ENSO is that El Niños are a series of discrete warm events punctuating periods of neutral or cold conditions (La Niñas). That is, ENSO can be characterized as a stable (or damped) mode triggered by stochastic atmospheric forcing (e.g., Lau, 1985; Penland and Sardeshmuhk, 1995; Neelin et al., 1998; Moore and Kleeman, 1999; Thompson and Battisti, 2001; Dijkstra and Burgers, 2002; Philander and Fedorov, 2003; Kessler, 2003; Zavala-Garay et al., 2003). This hypothesis proposes that disturbances external to the coupled system are the source of random forcing that drives ENSO. An attractive feature of this hypothesis is that it offers a natural explanation in terms of noise to 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; Philander and Fedorov, 2003). The external atmospheric forcing may include the Madden-Julian oscillation and westerly wind bursts (see the review of Lengaigne et al. (2004)), and may even involve explosive volcanism (Adams et al., 2003) although this remains a controversial hypothesis.

No matter whether El Niño is a self-sustained cycle or a stable mode triggered by stochastic forcing, El Niño begins with warm SST anomalies in the equatorial central and eastern Pacific. After an El Niño reaches

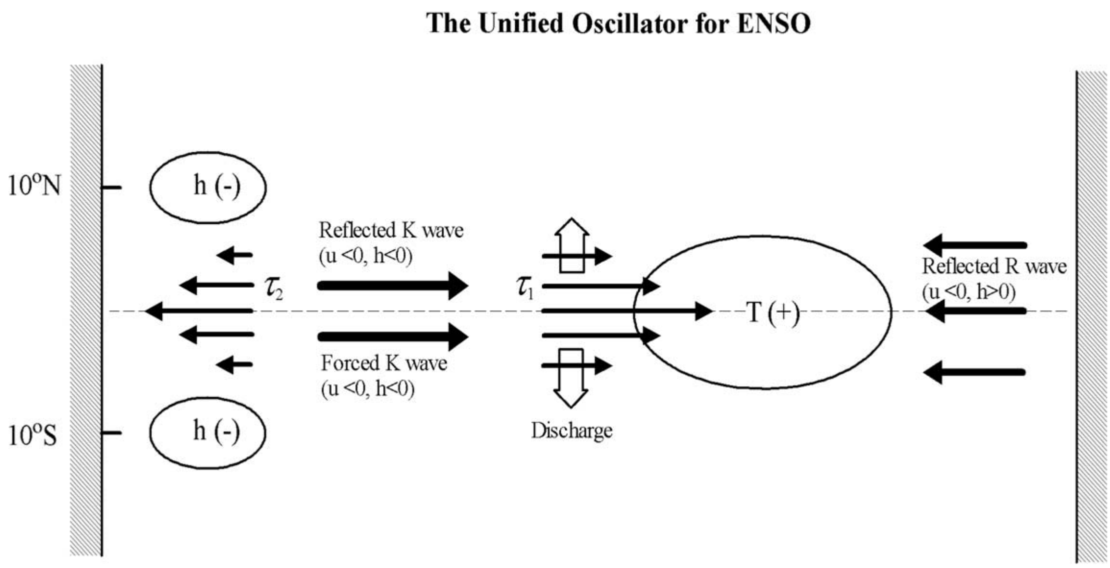

Fig. 12. Schematic diagram of the unified oscillator for ENSO. Bjerknes' positive ocean–atmosphere feedback leads the equatorial central/eastern Pacific to a warm state (El Niño). Four negative feedbacks, required to turn the warm state around, are (1) reflection of Kelvin waves at the ocean western boundary, (2) discharge due to Sverdrup transport, (3) western Pacific wind-forced Kelvin waves, and (4) reflection of Rossby waves 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.

its mature phase, negative feedbacks are required to terminate growth of the mature El Niño anomalies in the central and eastern Pacific. In other words, the negative feedbacks of 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. As discussed by Mantua and Battisti (1994), a sequence of independent warm events can still be consistent with delayed oscillator physics, since the termination of an individual El Niño event still requires negative feedback that can be provided by wave reflection at the western boundary.

4. The eastern Pacific warm pool

As stated in Section 1, the Western Hemisphere warm pool (WHWP) is defined as the region covered by water warmer than 28.5 °C on both the Pacific and Atlantic sides of Central America (Wang and Enfield, 2001, 2003). These are temperatures that have a significant impact on organized tropical convection (e.g., Graham and Barnett, 1987). The choice of 28.5 °C is based not only on limiting the WHWP to a closed region, but also on the fact that the depth of the 28.5 °C isotherm is closest to the average mixed layer depth in the WHWP (Wang and Enfield, 2003). The WHWP is separated into two parts by the Central American landmass: the eastern north Pacific warm pool and the Intra-Americas Sea (IAS) warm pool (see Fig. 13). We review here seasonal and interannual variations of the eastern Pacific warm pool. Kessler (2002) and Kessler (2006) provided the ocean circulation patterns in the eastern tropical Pacific. Xie et al. (2005) used high-resolution satellite observations to examine ocean–atmosphere interaction over the eastern Pacific warm pool. Amador et al. (2006) reviewed the atmospheric forcing of the eastern tropical Pacific.

4.1. Seasonal cycle

The warm pool starts to develop in the eastern north Pacific during the boreal early spring (Fig. 13). By May, the warm pool is well developed in the eastern Pacific, but begins to shrink shoreward and spread along the coast where tropical storms frequently develop in the early summer. This becomes the core region of warm waters, which, in conjunction with strengthening of the intertropical convergence zone (ITCZ), triggers the onset of the Mexican and North American summer monsoons. June is a transition period when the warm pool on the Atlantic side of Central America starts to develop, while the warm pool in the eastern Pacific decays. By July, water warmer than 28.5 °C is well developed in the Gulf of Mexico, and the WHWP covers the Gulf of Mexico and to less extent the eastern Pacific. In August, the warm water in the Gulf of Mexico reaches its maximum with a large area covered by water warmer than 29.5 °C. By September, the warm pool has expanded south into the Caribbean and eastward into the western tropical North Atlantic, while the water in the Gulf of Mexico has cooled. The WHWP decays quickly after October.

The seasonal cycle of the eastern Pacific warm pool is clearly shown in Fig. 14a, which is calculated as the mean NCEP SST (Smith et al., 1996) over the area enclosed by the 28.5 °C isotherm. There is no water warmer than 28.5 °C in the eastern Pacific during January and February, and the seasonal cycle shows double peaks. The SST in the eastern Pacific warm pool displays two maximum values in May and August. This pattern seems to relate to the midsummer drought discussed by Magaña et al. (1999) (also see Amador et al., 2006). Magaña et al. (1999) showed double peaks in the annual cycle of precipitation over the southern part of Mexico and Central America with maxima during June and September, separated by the so-called the midsummer drought during July/August. Fig. 14a shows that the eastern Pacific warm pool SST leads double peaks of precipitation over the southern part of Mexico and Central America by one month, and that the area of the eastern Pacific warm pool is approximately in phase with the precipitation.

Using the NCEP-NCAR reanalysis, the NCEP SST, and the Levitus mixed layer depth data, Fig. 14b shows surface net heat flux and heat storage tendency ($\rho C_P h \partial T/\partial t$) over the eastern Pacific warm pool. The Levitus data show that the mixed layer depth across the warm pool is on average about 25 m (Wang and Enfield, 2003), so we choose $h = 25$ m for our calculation. The penetration of shortwave radiation follows the recent work of Murtugudde et al. (2002): $Q_P = 0.47 Q_S \exp(-h/17)$. Although the heat storage tendency is approximately in phase with the net heat flux, the net heat flux is larger than the heat storage tendency

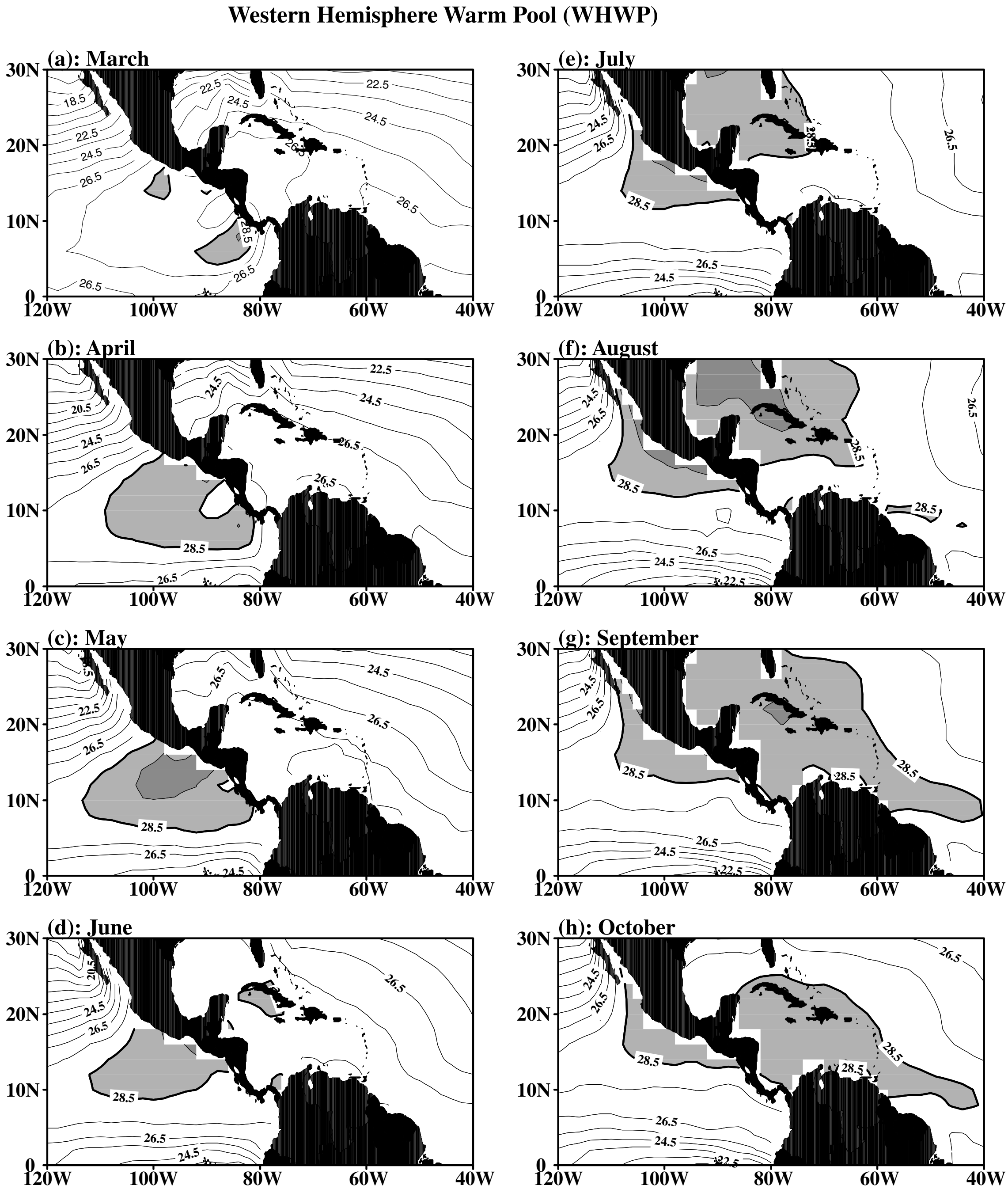

Fig. 13. Seasonal distributions of SST for the tropical Western Hemisphere warm pool (WHWP): (a) March, (b) April, (c) May, (d) June, (e) July, (f) August, (g) September, and (h) October. The shading and dark contour represent water warmer than 28.5 °C. The data are from the NCEP SST.

whether or not heat flux includes the penetration of shortwave radiation. This suggests that processes other than surface flux are needed to cool SST in the eastern Pacific warm pool, such as ocean advection. Notice that the eastern Pacific warm pool is different from the IAS warm pool, where surface heat fluxes are primarily responsible for seasonal warming (Wang and Enfield, 2003).

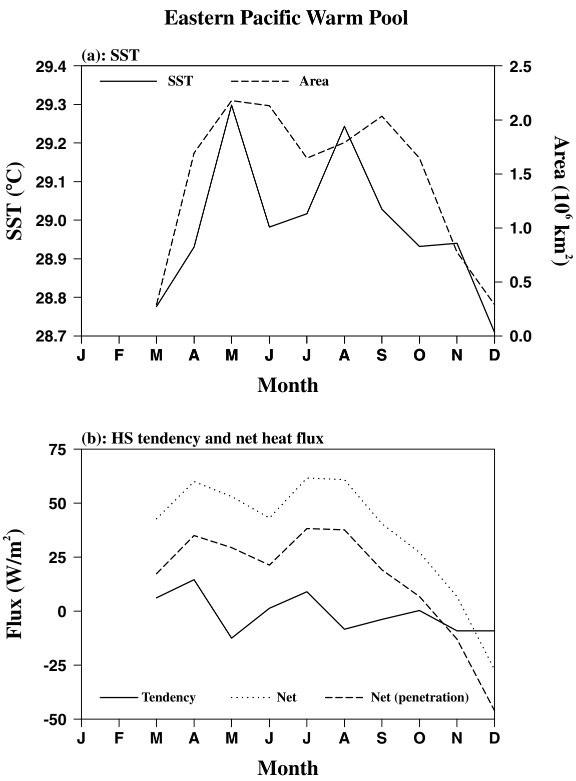

Fig. 14. (a) Seasonal variations of SST over the eastern Pacific warm pool (larger than 28.5 °C) and the area enclosed by water warmer than 28.5 °C; (b) heat storage (HS) tendency ($\rho C_P h \partial T/\partial t$) of the eastern Pacific warm pool, surface net heat flux (shortwave radiation minus latent heat flux, longwave radiation, and sensible heat flux), and surface net heat flux after considering penetration of shortwave radiation. The data are from the NCEP SST, the NCEP-NCAR reanalysis field, and the Levitus data.

The eastern Pacific and IAS warm pools are a heat source of summer Hadley circulation and play a key role in the transition from the South American to the North American Monsoon. Fig. 15 shows the boreal summer (July) climatologies of tropospheric circulation patterns. The center of upper tropospheric divergence associated with middle tropospheric ascent is near the region of the WHWP. Two subsiding limbs of the overturning are located over the subtropical western South Atlantic and eastern South Pacific (as manifested by 200-mb divergence and 500-mb descending motion). The Hadley circulation between 110° W–70° W involves air rising around 10° N–15° N, diverging southward in the upper troposphere, descending in the subtropical South Pacific, then crossing the equator at the surface and returning to the Northern Hemisphere convergent region. The atmospheric circulation patterns are consistent with the results of Bosilovich and Schubert (2002), and Hu and Feng (2001) showing that warm pool is a source of moisture for North America.

4.2. Interannual variability

The eastern Pacific warm pool index is calculated for the anomalies of the area enclosed by the 28.5 °C isotherm, as shown in Fig. 16a. During the 50-year period since 1950 (from 1950 to 1999), large warm pools occur in 1957–58, 1969, 1972, 1982–1983, 1987, 1990–1991–1992–1993, and 1997–1998, all corresponding to El Niño events (Fig. 16b). The maximum correlation between the eastern Pacific warm pool area anomalies and the Niño3 SST anomalies is 0.75 at zero lag.

The horizontal structure of the eastern Pacific warm pool during the warm events can be examined by comparing with the climatological warm pool. Since the largest climatological eastern Pacific warm pool occurs in May, we composite the structure of the eastern Pacific warm pool based on the May SST values of 1957, 1969, 1972, 1983, 1987, 1992, and 1997. The result is shown in Fig. 17. As expected, Fig. 17 shows that the size of the

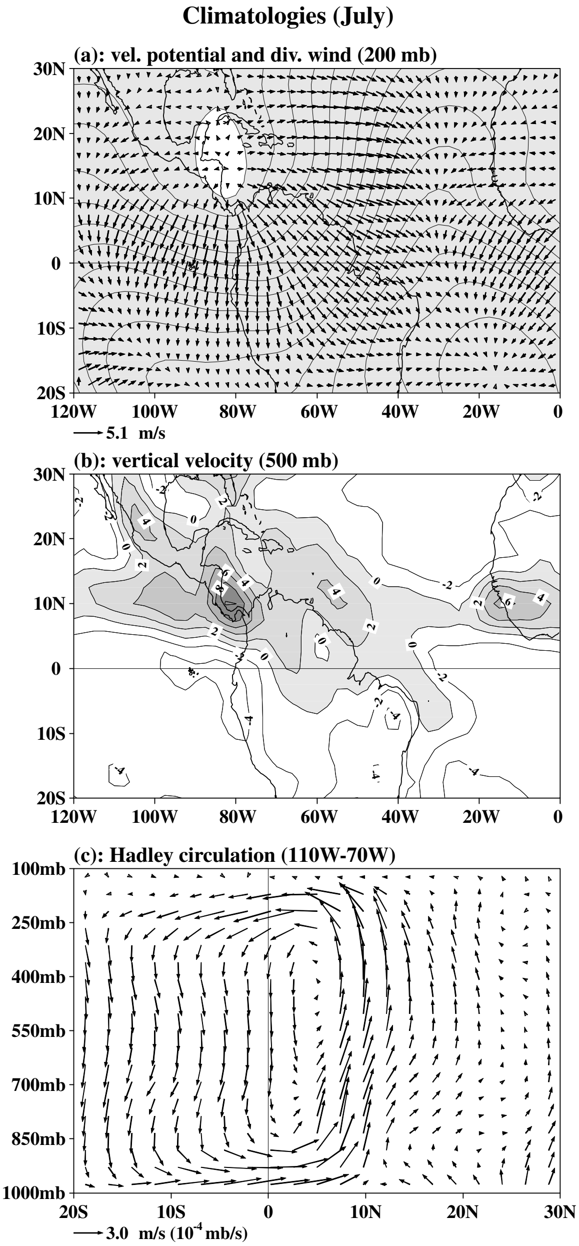

Fig. 15. The boreal summer (July) climatologies of tropospheric circulation patterns. (a) 200-mb velocity potential (10^6 m^2/s) and divergent wind (m/s), (b) 500-mb vertical velocity (10^{-4} mb/s), and (c) meridional-vertical circulation by averaging divergent wind and vertical velocity between 110° W–70° W. The vertical velocity is taken as the negative of the pressure vertical velocity in the reanalysis, i.e., positive values indicate an upward movement of air parcels. Positive values are shaded.

eastern Pacific warm pool for the interannual warm events is much larger than the climatological warm pool (dark contour). All of these suggest that interannual variability of the eastern Pacific warm pool is clearly related to the El Niño events.

What mechanism controls interannual variability of the eastern Pacific warm pool? Like the seasonal cycle, the mechanisms for controlling interannual variability of the eastern Pacific and IAS warm pools seem to be different. Wang (2002b) and Wang and Enfield (2003) showed that during winters preceding large IAS warm pools, there is a strong weakening of the Hadley circulation that serves as a "tropospheric bridge" for

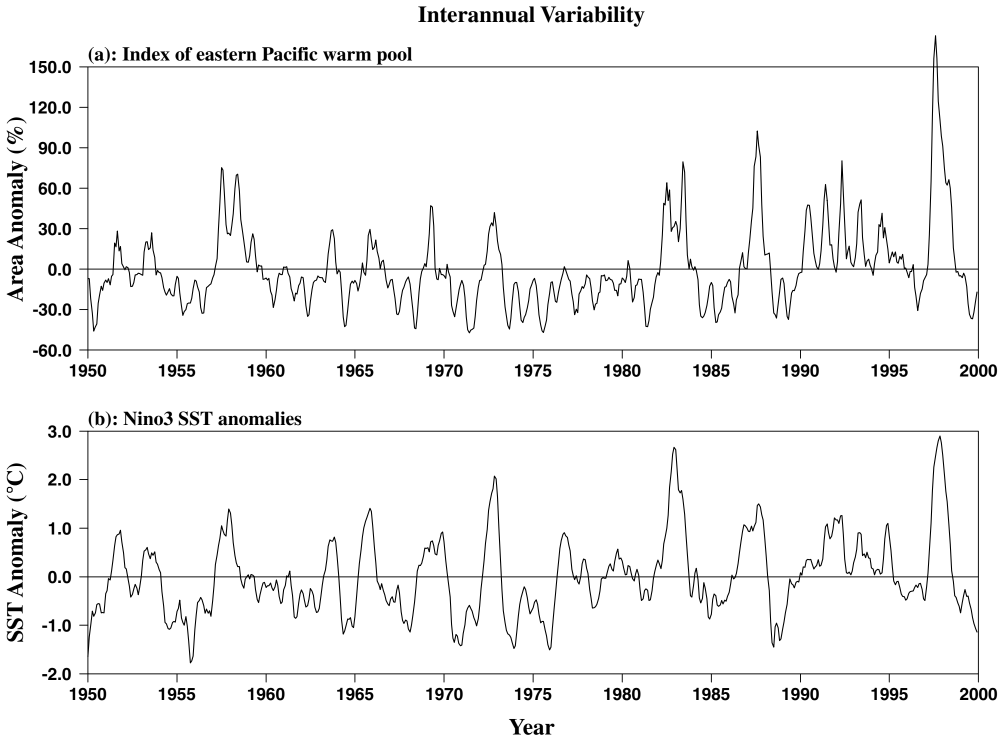

Fig. 16. (a) Eastern Pacific warm pool area anomalies (%) for SST warmer than 28.5 °C, and (b) SST anomalies in the Niño3 region (5° S–5° N, 150° W–90° W). The eastern Pacific warm pool area anomalies (in unit of percentage) are calculated as anomalies of area (⩾28.5 °C) divided by the May climatological warm pool area.

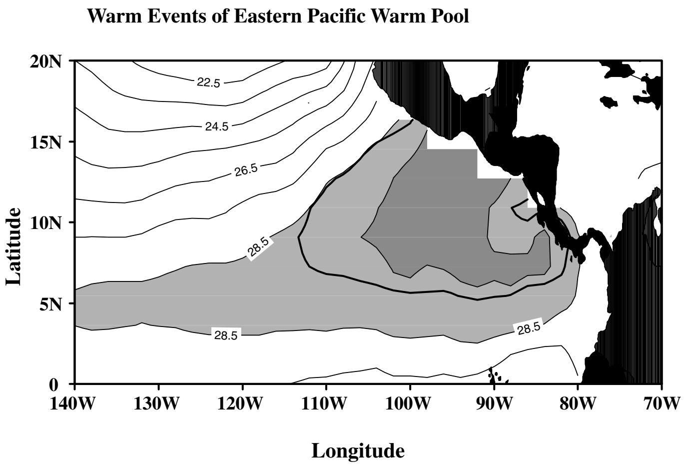

Fig. 17. SST composite for the eastern Pacific warm pool interannual warm events. The composites are calculated by averaging the May SST values over the warm years of 1957, 1969, 1972, 1983, 1987, 1992, and 1997. The shading represents water warmer than 28.5 °C. The dark contour is climatological May eastern Pacific warm pool (SST warmer than 28.5 °C).

transferring Pacific El Niño effects to the Atlantic sector and inducing warming of IAS warm pool. After the ENSO-related warming, a positive ocean–atmosphere feedback operating through longwave radiation is responsible for maintaining or amplifying SST anomaly growth in the region of the IAS warm pool.

Associated with the warm SST anomalies is a decrease in SLP anomalies and an anomalous increase in atmospheric convection and cloudiness. The increase in convective activity and cloudiness results in less longwave radiation loss from the sea surface, which then reinforces SST anomalies. However, this mechanism does not seem to operate in the eastern Pacific warm pool. We calculated correlations among variables in the eastern Pacific warm pool. The eastern Pacific warm pool longwave radiation anomalies are not significantly correlated with the eastern Pacific warm pool SST anomalies, and the eastern Pacific warm pool SLP anomalies are not correlated with the eastern Pacific warm pool cloud anomalies either. All of these suggest that the positive ocean–atmosphere feedback operating through longwave radiation and associated cloudiness is not responsible for the warmings of the eastern Pacific warm pool on interannual scales. Notice that the eastern Pacific warm pool shortwave radiation and latent heat flux anomalies are not significantly correlated with the eastern Pacific warm pool SST anomalies. Interannually, the eastern Pacific warm pool may be due to a direct oceanic connection to the equator, associated with extreme ENSO phases.

5. ENSO biological and ecological variability

ENSO-related changes in winds, insolation, hydrography and circulation in the eastern tropical Pacific are of sufficient magnitude and duration to affect organisms, populations and ecosystems. The species and communities of the region have evolved to persist through the quasi-regular disturbances imposed by ENSO events. Thus, typical or even exceptional events should not result in long-term, fundamental changes (Paine et al., 1998). Biological effects of recent El Niño events in the region have been documented and explained primarily for phytoplankton and commercial fish stocks. For example, Barber and Chavez (1986) summarized the effects of the 1982–1983 El Niño in the eastern equatorial and Peru upwelling systems: deepening of the thermocline, and thus the nutricline, resulted in decreased primary production that ultimately affected survival, reproduction, and distribution of higher trophic level organisms. Sánchez et al. (2000) summarized effects of the 1997–1998 El Niño on the coastal marine ecosystem off Peru: SST rose by up to 9 °C, the thermocline, nutricline and oxycline deepened, nitrate in surface waters decreased, zooplankton volumes decreased, exotic warm-water dinoflagellates and copepods increased, and changes in distribution of fish, cetaceans and seabirds were observed.

In the eastern equatorial Pacific, sampling by ships of opportunity showed that the 1982–1983 El Niño caused a deepening of the thermocline, and a reduction in chlorophyll and copepod abundance (Dessier and Donguy, 1987). Effects of ENSO on phytoplankton and zooplankton in the eastern tropical Pacific are further reviewed in Pennington et al. (2006) and Fernández-Álamo and Färber-Lorda (2006), respectively.

El Niño effects on fish in coastal upwelling systems are well-known. The 1972 El Niño resulted in a recruitment failure for Peruvian anchoveta that, along with overfishing, led to a collapse of the world's largest fishery (Clark, 1977). The 1982–1983 El Niño had a variety of effects on commercial fish stocks in Peru: hake moved down the continental slope to stay in cooler deep water, shrimp and sardines moved southward so that catches in some areas decreased and in other areas increased, jack mackerel moved inshore in search of euphausiid prey and were subject to high predation mortality there, scallop abundance increased due to enhanced reproductive success in warmer water, and the anchoveta population crashed due to reduced food availability for adults and larvae. Many of these stocks recovered rapidly beginning in late 1983 (Barber and Chavez, 1986).

There have been few reports of ENSO effects on animals other than commercially exploited fish in the eastern tropical Pacific, except for the well-known mass mortalities of guano-producing seabirds on the coast of Peru (Wooster, 1960). Seabird populations experienced breeding failures, mass mortalities, and migrations in search of food throughout the tropical Pacific in 1982–1983 (Ainley et al., 1986), although a few species were not affected. Body weight of Galapagos penguins increased during the 1971 La Niña and decreased during 1972 El Niño, indicating short-term response to food availability, but the population suffered 77% mortality during the 1982–1983 El Niño and had not recovered by 1997 (Dee Boersma, 1998). Blue-footed booby reproductive attempts failed and breeding colonies were abandoned during the 1986–1987 El Niño, apparently in response to reduced availability of sardines, but several other species were not affected (Anderson, 1989). Piscivorous seabirds in coastal Peru have consistently experienced adult mortality and decreased reproductive success during El Niño events; these are short-term population effects, resulting from reduced availability of anchoveta (Tovar et al., 1987; Crawford and Jahncke, 1999). All fifteen species of seabirds nesting on the Galapagos Islands stopped breeding or experienced reduced reproductive success during the 1982–1983 El Niño, but resumed breeding the following year (Valle et al., 1987). Pelagic seabird surveys have shown changes in the relative abundance of less common species, but not the dominant species, during El Niño and La Niña events in the eastern equatorial Pacific (Ribic et al., 1992). Such changes may be explained by shifts in distribution between the equatorial and subtropical water masses covered by these surveys (Ballance et al., 2006). In general, seabirds that forage in equatorial and coastal upwelling areas of the eastern tropical Pacific suffer reproductive failure and mortality due to food shortage during El Niño events. However, other species that forage in areas less affected by El Niño, for example the warm pool, may be relatively unaffected (Ballance et al., 2006).

Mortality and other population effects of El Niño on marine mammals have been observed in coastal ecosystems in the eastern tropical Pacific. Manzanilla (1989) observed a 1983 "El Niño mark" in the teeth of mature female Peruvian dusky dolphins and suggested that the mark resulted from low foraging success for the preferred prey, anchoveta, which became unavailable during the 1982–1983 El Niño. Consistent with this inference, no such marks were observed in other dolphin species that consume other prey. Galapagos pinnipeds suffered great mortality in 1982–1983, especially in younger year classes, and reduced pup production due to reduced food availability (Trillmich and Limberger, 1985). Peruvian pinnipeds were affected by the reduced availability of anchoveta in 1983 (Majluf and Reyes, 1989) and again in 1998 (Soto et al., 2003). Ramirez (1986) observed diet changes and reduced feeding success of Bryde's whales off Peru during 1982–1983 El Niño. ENSO effects on oceanic cetaceans are reviewed in Ballance et al. (2006).

Biological effects of El Niño, such as temporary range shifts, may be benign in the sense that normal or average conditions are quickly restored. Effects of the opposite, La Niña, phase of the ENSO cycle are less well documented, probably because La Niña events do not bear the negative reputation of El Niño events. It has been argued that ENSO variability might be responsible for the high production of the Peru coastal upwelling system by disrupting biological controls such as predation and opening "loopholes" for the success of fecund, small pelagic fishes (Bakun and Broad, 2003). ENSO variability is also subject to low-frequency modulation. See Mantua and Hare (2002) and Miller et al. (2003) for recent reviews of Pacific decadal variability and its effects on ocean ecosystems.

6. Changes in ENSO variability

Changes in the characteristics, or modulation, of ENSO variability over the past one to two centuries have been described in instrumental and proxy records from the tropical Pacific. Mestas-Nuñez and Enfield (2001) found that the late 1970s climate shift that warmed the eastern equatorial Pacific (Niño3 region) by about 0.5 °C was also characterized by increased interannual variance through the 1980s and 1990s. An 1893–1994 coral record from Clipperton Atoll (within the eastern Pacific warm pool area) shows both ENSO and decadal-scale variability closely related to instrumental Southern Oscillation Index (SOI) and Pacific Decadal Oscillation (PDO) records from recent years (Linsley et al., 2000). Reduced ENSO variability is evident between 1925 and the mid-1940s. An and Wang (2000) found that the dominant period of SST variability in the central and eastern equatorial Pacific (Niño-3.4 region) shifted from 3.3 years during 1967–1973 to 4.2 years during 1980–1993. Setoh et al. (1999) found a similar increase in the period of ENSO in the equatorial Pacific during the late 1970s, but also found slight changes in the spatial pattern of the ENSO signal. Spectral analysis of a composite record of ENSO events since 622 A.D. showed that the period of ENSO variability has varied, within a range of 1.5–10 years, in cycles of 90, 50, and 23 years (Anderson, 1992). Philander and Fedorov (2003) suggest that there are multiple modes of ENSO, all weakly damped oscillations sustained by random disturbances, depending on the background state of the tropical Pacific Ocean–atmosphere system.

The irregularity of ENSO is obvious when the magnitude and spacing of individual events are considered in an instrumental time series (cf. Fig. 16b). Wunsch (1999) warns, however, that "before concluding that one is seeing evidence for trends, shifts in the mean, or changes in oscillation periods, one must rule out the purely random fluctuations expected from stationary time series". Solow and Huppert (2003), for example, found no significant nonstationarity in 1876–2002 records of Darwin sea level pressure and Niño3 SST. Rajagopalan et al. (1997) show that the "anomalous" prolonged El Niño of 1990–1995 was within the natural variability of the stationary time series of the preceding century. An as-yet unpublished study by Mitchell and Wallace[1] shows that spatial and temporal patterns of ENSO-related anomalies in global wind, temperature and precipitation fields have changed very little since the late 19th century. Fedorov and Philander (2000) argued that apparent changes in the characteristics of ENSO variability may simply reflect decadal-scale changes in the background state (climatology) against which El Niño and La Niña are measured. Characteristics and mechanisms of scales of variability longer than ENSO are reviewed in Wang and Picaut (2004) and Mestas-Nuñez and Miller (2006).

7. Summary

ENSO shows an interannual variability in both the eastern and western tropical Pacific. During the warm phase of ENSO, warm SST and low SLP anomalies in the equatorial eastern Pacific and low OLR anomalies in the equatorial central Pacific are accompanied by cold SST and high SLP anomalies in the off-equatorial western Pacific and high OLR anomalies in the off-equatorial western Pacific. The off-equatorial anomalous anticyclones in the western Pacific initiate and produce equatorial easterly wind anomalies over the western Pacific. So, while the zonal wind anomalies over the equatorial central Pacific are westerly, those over the equatorial western Pacific are easterly. ENSO is observed to be phase-locked to the seasonal cycle. That is, the maximum (minimum) SST anomalies in the equatorial eastern and central Pacific for each El Niño warm (La Niña cold) event occur during the boreal winter. In spite of this, the El Niño warm events over the last 50 years have evolved differently. For the El Niño events between 1949 and 1976, anomalous surface warming occurs first off the South American coast, and then progresses westward along the equator into the eastern and central Pacific. The initial warming of the El Niño events between 1976 and 1996 occurred in the equatorial central Pacific. The 1997–1998 El Niño developed in both the equatorial central Pacific and the South American coast during the spring of 1997. The 2002–2003 El Niño started in the equatorial central Pacific. However, a common feature for all El Niño events since 1950 is that westerly wind anomalies in the equatorial western Pacific always led the Niño3 SST anomalies by about four months, suggesting that the western Pacific is an important region for ENSO variability in the eastern tropical Pacific.

ENSO has been viewed as a self-sustained, naturally oscillatory mode or a stable mode triggered by stochastic forcing. For both views, ENSO involves the positive ocean–atmosphere feedback over the eastern tropical Pacific hypothesized by Bjerknes in the 1960s. After an El Niño reaches its peak, a negative feedback is required for terminating a continued growth of El Niño. Different negative feedbacks have been proposed since the 1980s associated with a delayed oscillator, a recharge oscillator, a western Pacific oscillator, and an advective-reflective oscillator. The delayed oscillator assumes that wave reflection at the ocean western boundary provides a negative feedback for the coupled system oscillation. The recharge oscillator argues that discharge and recharge of equatorial heat content cause the coupled system to oscillate. The western Pacific oscillator emphasizes equatorial wind in the western Pacific that provides a negative feedback for the coupled system. The advective-reflective oscillator emphasizes the importance of zonal advection associated with wave reflection at both the western and eastern boundaries and of the mean zonal current. As suggested by the unified oscillator, all of these negative feedbacks work together to terminate El Niños, and their relative importance is time-dependent.

The eastern Pacific warm pool, defined by surface water warmer than 28.5 °C, is in the region of the eastern tropical Pacific north of the equator. The seasonal SST of the eastern Pacific warm pool shows double maximum values in May and August that lead the double peaks of precipitation over the southern part of Mexico and Central America by one month. The precipitation peaks are separated by the midsummer drought. Seasonally, ocean advection seems to play an important role for SST variations. Interannually, the eastern Pacific warm pool may be remotely forced by ENSO variability along the equator.

ENSO variability has been shown to affect phytoplankton, invertebrates, fish, birds, pinnipeds and cetaceans. Almost all observations of such environmental effects have been on coastal or island populations. While

[1] Mitchell, T.P., and Wallace, J.M. The instrumental record of ENSO: 1840–2000. http://jisao.washington.edu/wallace/enso_chronology.pdf.

it is true that ENSO effects are more extreme in highly productive coastal environments, other factors may be important. Perhaps such populations are less adaptable or opportunistic. Certainly, these populations are more readily accessible for long-term study. El Niño events often cause changes in distribution of species as the distributions of preferred water masses and prey changes. Population effects are observed on local breeding grounds, but recovery usually occurs rapidly when the El Niño event is over. Populations of marine animals in the eastern tropical Pacific have undoubtedly evolved life history strategies to deal with ENSO-related variability in their environment.

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