The Principal Physical Modes of Variability over the Tropical Pacific

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**ABSTRACT:** The structures of the two principal modes of sea surface temperature (SST) variability were extracted by conducting cyclostationary EOF (CSEOF) analysis and regression analysis on several key variables. The CSEOF analysis extracts two dominant modes of SST variability that are distinct in nature. The first CSEOF is stochastic in nature and represents a standing mode of SST variability associated with a basinwide change in the surface wind. The second CSEOF exhibits a strong deterministic component describing a biennial oscillation between a warm phase and a cold phase. The surface wind directional change in the far-western Pacific appears to be instrumental for the oscillation between the two phases. Because of the distinct nature of evolution, dynamical and thermodynamical responses of the two modes are different. Further, the predictability of the two modes is different. Specifically, the biennial mode is more predictable because of the strong deterministic component associated with its evolution. The distinction of the two modes, therefore, may be important for predicting ENSO. The irregular interplay of the two modes seems to explain some inter-ENSO variability, namely, variable duration of ENSO events, approximate phase-locking property, and irregular onset and termination times.

**KEYWORDS:** Ocean/atmosphere interactions, El Niño, Tropical meteorology, Climate dynamics

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1. Introduction

Since the pioneering work of Bjerknes (Bjerknes, 1969), El Niño and La Niña are generally viewed as the oceanic manifestation of the opposite phases of the tropical coupled circulation system called the El Niño and the Southern Oscillation (ENSO). Basic principles of how the coupled ocean–atmosphere system works have been proposed and appear to provide reasonable explanations for these phenomena.

The presence of a seemingly biennial oscillation in many observational records, however, has been a source of controversy for a long time. An issue is whether El Niño and La Niña should be viewed as irregular occurrences (events) or as irregular manifestations of otherwise oscillatory behavior of the tropical Pacific Ocean. Rasmusson et al. (Rasmusson et al., 1990) appear to be the first investigators to have taken the latter view. In the surface wind field of the equatorial eastern Indian Ocean, they identified a well-defined standing component of ENSO variability, which is tightly phase-locked with the annual cycle. They proposed that this standing component is part of a larger-scale biennial oscillation over the low-latitude eastern Indian Ocean–western Pacific sector, and that it is one of the fundamental elements of ENSO variability. Although the concept of the biennial oscillation has gained some acceptance, the debate continues. Henceforth, this paper examines the biennial oscillation further using novel analysis tools.

A related issue is how to interpret the irregularity of El Niño and La Niña. From the viewpoint of data analysis, no two El Niños (and also La Niñas) have the same amplitude and evolving patterns. The occurrence of El Niño and La Niña also seems to exhibit irregular behavior. In addition to this interannual variability, the onset and termination times of El Niño and La Niña with respect to the annual cycle appear to fluctuate widely from one event to another. It is not clear whether this irregularity represents inherent physical and dynamical characteristics of the ENSO system or merely irregular manifestations of different physical modes of the ENSO system. Understanding this inter- and intraannual variability is then of great practical value and is key to improving the predictability of ENSO.

In the present study, ENSO variability is investigated on all relevant scales using a technique called the cyclostationary EOF (CSEOF) analysis (Kim and North, 1997). The main focus of this work is to describe in detail the two principal computational modes of variability associated with the ENSO system from the depth of the ocean to the tropopause. To this end, physically consistent patterns of several key variables are derived via a regression analysis. Based on the results in this study, inferences are made regarding the evolution of the general circulation structures during El Niños and La Niñas. This work provides a physical and dynamical interpretation of the biennial oscillation over the tropical Pacific, and offers a likely reason for the irregularity of ENSO on the basis of the analysis results. Understanding the mechanisms of a physical system is inevitably difficult through data analysis alone. Therefore, some of the conclusions drawn in this study may require further confirmation using more elaborate tools.

Section 2 describes the datasets used in this study. CSEOF analysis, regres-
sion analysis, and Kelvin and Rossby waves decomposition are described in section 3 with particular emphasis on the extraction of physically consistent evolutions from different physical variables. The results of the present analysis are discussed in terms of the two primary physical modes in section 4. The conclusions are found in section 5.

2. Data

In this study, we use data from various sources of disparate nature. The sea surface temperature anomaly, sea level pressure anomaly, and surface wind anomaly datasets are the $1^\circ \times 1^\circ$ monthly averages constructed by da Silva et al. (da Silva et al., 1994). The datasets extend from January 1945 to December 1993. These datasets were subsequently interpolated onto $2.5^\circ \times 2.5^\circ$ arrays. The sea surface temperature anomaly dataset was further extended to May 1998 by splicing the Comprehensive Ocean–Atmosphere Data Set [COADS (Woodruff et al., 1987)].

This study also uses the $1^\circ \times 1^\circ$ monthly averaged assimilated subsurface temperature anomaly and zonal current anomaly data from Giese and Carton (Giese and Carton, 1999) and Carton et al. (Carton et al., 2000a; Carton et al., 2000b). These datasets represent the Geophysical Fluid Dynamics Laboratory Modular Ocean Model (MOM) run by assimilating the COADS data modified by da Silva et al. (da Silva et al., 1994). The assimilated ocean datasets cover the period of January 1950–December 1999. The original data were at irregular sampling points in the meridional direction and have been interpolated onto a $2.5^\circ \times 2.5^\circ$ grid.

The monthly sea level height anomaly data are an assimilated dataset and are archived at the Environmental Modeling Center of the National Centers for Environmental Prediction (NCEP). The assimilated data were obtained from a model forced with weekly mean NCEP operational atmospheric analyses of surface winds and heat fluxes. For more information, see Behringer et al. (Behringer et al., 1998).

The geopotential height and wind data at standard vertical levels came from the monthly NCEP reanalysis data (Kalnay et al., 1996). The NCEP reanalysis data were acquired as $5^\circ \times 2.5^\circ$ arrays. Finally, the precipitation rate was obtained from the $2.5^\circ \times 2.5^\circ$ Xie–Arkin data (Xie and Arkin, 1997).

3. Methodology

3.1 Cyclostationary analysis

In this study, CSEOF analyses have been conducted on observational data with the assumption of a 24-month nested period. (The reader is referred to appendixes A and B for more details on the CSEOF technique.) This assumption was primarily motivated by the work of Rasmusson et al. (Rasmusson et al., 1990), Clarke and van Gorder (Clarke and van Gorder, 1999; Clarke and van Gorder, 2000), and Weiss and Weiss (Weiss and Weiss, 1999). Indeed, the covariance function of the Niño-3 time series appears to have a significant 2-yr component
Figure 1. (left) The two CSEOFs of the time series from the event model in Equation (1). (right) The corresponding covariance function, $C(t, t')$, of the time series. As shown on the left-hand side, the odd-year events and the even-year events are separated as two distinct modes under the CSEOF analysis with a nested period of 24 months.

(Kim, 2002). The most convincing evidence resides in the appreciable amount of variance associated with the biennial structure extracted from the observational data, as will be shown below. It should be mentioned beforehand, however, that the biennial mode may not necessarily be an independent physical mode; it could be one facet of a complex physical process.

One might argue that the biennial structure is due to an annual event, whose occurrence and sign are completely independent between two consecutive years. Such data can be constructed as

$$T(t) = \begin{cases} 
a(t)f(t) & t = \text{odd years} 
\end{cases}$$

$$b(t)f(t) & t = \text{even years},$$

where $f(t)$ is an annual structure, $a(t)$ is the amplitude of the annual structure in odd years, and $b(t)$ is the amplitude of the annual structure in even years. If $a(t)$ is independent of $b(t)$, CSEOF analysis separates the odd-year events from the even-year events as separate modes as shown in Figure 1. This separation arises from the fact that the odd- and even-year events are not correlated, as shown on
the right-hand side of Figure 1. Such a separation was not observed in the CSEOF analysis of the Niño-3 time series, the Niño-3.4 time series, the Southern Oscillation index (SOI) time series, and the sea surface temperature anomaly field. Therefore, the so-called event model (Torrence and Webster, 1998, Torrence and Webster, 2000), in which the odd-year and even-year events are independent of each other, cannot explain the seemingly biennial structure.

### 3.2 Regression analysis in the CSEOF domain

Once the evolution pattern of the sea surface temperature anomaly field is determined, consistent patterns of other variables can be found via regression analysis.
Figure 2. A schematic diagram showing the temporal evolutions of two physical variables pertaining to the same physical process (or system). When the system is externally forced, the two consistent patterns are supposed to exhibit the same evolution.

In practice, a consistent pattern is often found by projecting the measurement of a physical variable onto a target time series; that is,

$$\phi(r) = \sum_{r} P(r, t)T(t),$$  \hspace{1cm} (2)

where $T(t)$ is a principal component (PC) time series of a target variable (say, sea surface temperature anomaly) with the corresponding spatial pattern, $\psi(r)$, and $\phi(r)$ is the consistent spatial pattern of the predictor variable (say, surface wind anomaly), $P(r, t)$. The reasoning behind such a practice is that $\psi(r)$ and $\phi(r)$ have the same evolution history, $T(t)$, and therefore they are physically consistent patterns of the two variables (see Figure 2). This so-called projection method is the same as regression analysis in the EOF space:

$$T(t) = \sum_{n} a_{n}P_{n}(t) + \epsilon(t),$$  \hspace{1cm} (3)

where $P_{n}(t)$ are the PC time series of the predictor variable. Then, the consistent pattern of the predictor variable is found from
where $\varphi_n(r)$ are the EOFs of the predictor variable.

According to the argument above, the predictor pattern should change by the same ratio that the target pattern changes by, which certainly is not a general description of the evolution of two physical variables. It should be noted that two different physical variables often exhibit different paths of physical evolution (see Figure 3). For example, the development of a westerly wind anomaly in the western Pacific induces positive Kelvin waves, the effect of which is seen in the eastern Pacific after a few months. In other words, there may exist a lead–lag relationship between two physical variables. Further, physical evolutions of two variables may not be exactly scalable between the two. This implies that different paths of evolution for different physical parameters should be explicitly accounted for in finding consistent physical patterns.

Note that CSEOFs describe the deterministic components of evolution while the PC time series represent the stochastic evolution (see appendix A). This should be compared with the EOF technique, where a physical process is independent of time and the PC time series solely dictates the evolution of a spatial pattern (see Figure 4). Suppose that two deterministic components of evolution are physically consistent; that is, they belong to the same physical process or physical system. In such a case, all we can stipulate is that the stochastic evolution

$$
\phi(r) = \sum_n a_n \varphi_n(r),
$$

Figure 3. A schematic diagram showing different physical evolutions of two physical variables pertaining to the same physical process (or system). When the system is externally forced, the evolutions of the two physical variables are not identical because of the different physical response characteristics.
of the two should be identical (see Figure 3 and Figure 4). Thus, as in Equation (3), we can formulate a regression relationship:

\[ T(t) = \sum_n a_n P_n(t) + \epsilon(t), \]

where \( T(t) \) and \( P_n(t) \) are cyclostationary PC time series. Then, the pattern of the predictor variable, \( \psi(r, t) \), consistent with the target pattern, \( \phi(r, t) \), is determined by

\[ \phi(r, t) = \sum_n a_n \varphi_n(r, t), \]

where \( \varphi_n(r, t) \) are the CSEOFs of the predictor variable. That is, \( \psi(r, t) \) and \( \phi(r, t) \) have the same stochastic evolution history. The deterministic components in \( \psi(r, t) \) and \( \phi(r, t) \), however, are not identical; in fact, they are determined by the physical equation relating the two variables. The patterns of surface wind anomaly, subsurface temperature anomaly, zonal current anomaly, geopotential height anomaly, and wind at standard vertical levels that are consistent with the first two CSEOFs of the sea surface temperature anomaly field are determined according to Equation (6) and are shown in what follows.
3.3 Kelvin and Rossby decomposition using the meridional function

The regressed sea level height anomaly patterns are further decomposed into Kelvin and Rossby modes by employing a standard decomposition method described in Boulanger and Menkes (Boulanger and Menkes, 1995; Boulanger and Menkes, 1999). This method provides a “natural” basis of functions for representing the dynamics of Kelvin and Rossby waves. The meridional structure functions for the zonal current and the sea level are given by

\[ u_0 = h_0 = \Psi_0(y), \]  

for Kelvin waves, and

\[ u_n = \frac{\sqrt{n(n+1)}}{\sqrt{2(2n+1)}} \left( \frac{\Psi_{n+1}}{\sqrt{n+1}} - \frac{\Psi_{n-1}}{\sqrt{n}} \right) \]

\[ h_n = \frac{\sqrt{n(n+1)}}{\sqrt{2(2n+1)}} \left( \frac{\Psi_{n+1}}{\sqrt{n+1}} + \frac{\Psi_{n-1}}{\sqrt{n}} \right), \]

for the \( n \)th Rossby waves. Here, \( \Psi_n(y) \) represent the normalized Hermitian functions. Thus, the regressed sea level height anomaly and surface wind anomaly patterns are decomposed into

\[ h_{\text{reg}}(x, y, t) = \sum_m \alpha_m(x, t)h_m(y) \]  

\[ u_{\text{reg}}(x, y, t) = \sum_m \beta_m(x, t)u_m(y), \]

where \( \alpha_m(x, t) \) and \( \beta(x, t) \) are the expansion coefficients. In this study, the maximum order of the Rossby waves is 10.

4. Results and discussion

4.1 Low-frequency mode

Movie 1 shows the surface and subsurface variability associated with the first CSEOF of the anomaly fields, which explains about 30% of the sea surface temperature variability. The spatial patterns of the sea surface temperature anomaly, surface wind anomaly, subsurface temperature anomaly, and zonal current anomaly associated with this mode do not change significantly throughout the nested period. Weak Kelvin wave propagation is found along the thermocline, contrasting with the strong wave activity associated with the second CSEOF as shown later. In fact, the sea surface temperature anomaly pattern in Movie 1 is very similar to that of the first EOF. The lack of a strong temporal evolution in the deterministic component indicates that this mode is approximately stationary.

Figure 5 shows the PC time series of the low-frequency mode (upper panel). This time series represents the stochastic modulation of the low-frequency mode and resembles the PC time series of the first EOF except for the relatively high-frequency component (2 yr or shorter) of variability. A spectral analysis shows
Movie 1. The low-frequency mode of the surface and subsurface variability. (top) Sea surface temperature and the surface wind anomalies. (bottom) Subsurface temperature and current anomalies along the equator.

See the online version of this paper to view animation.
that the time series has a fairly broad peak at around the 5.3-yr period (Kim, 2002). In the absence of fast physical evolution, the fluctuation of this mode in the observational data is primarily due to the slow stochastic modulation in Figure 5. Therefore, this mode is called the low-frequency mode (Rasmusson et al., 1990).

The PC time series of the low-frequency mode is significantly correlated with the global residual time series of Zhang et al. (Zhang et al., 1997). The sea surface temperature anomaly and surface wind anomaly patterns of the low-frequency mode are generally consistent with the so-called interdecadal mode (Zhang et al., 1997). Indeed, the extension of the low-frequency mode to the North Pacific via a regression analysis reveals a characteristic pattern of the interdecadal mode (Movie 2). This seems to indicate that the interdecadal mode is a long timescale manifestation of the low-frequency mode. In fact, the PC time series of the low-frequency mode exhibits a significant amount of interdecadal fluctuation. The PC time series of the low-frequency mode is moderately correlated with the time series of the Pacific decadal oscillation (Mantua et al., 1997). This correlation of the low-frequency mode with the variability in the North Pacific seems to suggest a dynamical connection between the tropical basin and the North Pacific. According to Barnett et al. (Barnett et al., 1999a; Barnett et al., 1999b), the interdecadal component of the northern Pacific atmospheric variability is strongly connected with the development of the surface wind anomaly in the western Pacific. This low-frequency wind change preconditions the mean state of the thermocline in the equatorial ocean to produce prolonged periods of enhanced or reduced ENSO. In fact, CSEOF analysis of the high-pass filtered (< 6 yr) SST anomaly shows an equatorial sea surface temperature anomaly pattern

Figure 5. The PC time series of (top) the low-frequency mode and (bottom) the biennial mode. The blue plum shading represents positive (solid) and negative (dotted) peaks of the first PC time series while the red plum shading represents peaks of the second PC time series.
similar to that in Movie 1, but its high-latitude counterpart is extremely weak (see also Zhang et al., 1997). Therefore, the North Pacific–equatorial Pacific connection appears to be present only at the low-frequency (> 6 yr) band.

4.2 Biennial mode

Movie 3 shows the surface and subsurface variability associated with the biennial mode (second CSEOF). This mode explains about 16% of the sea surface temperature anomaly variability. As the name implies this mode represents an oscillation of variability with a 2-yr period. The biennial evolution seen in the figure is not due to the assumption of the 2-yr nested periodicity. As is shown later (Movie A1), the physical evolution in the annual cycle [the first CSEOF of the sea surface temperature field (not anomaly)] has a 1-yr period despite the use of the 2-yr nested periodicity. If there were no 2-yr deterministic component, the biennial mode would not be extracted from the data (see also appendix B).

In the winters of odd years, a negative subsurface temperature anomaly is observed in the far-western Pacific (120°–160°E) at the thermocline depth. This negative subsurface temperature anomaly appears to be associated with the west-
Movie 3. The biennial mode of the surface and subsurface variability. (top) Sea surface temperature and surface wind anomalies. (bottom) Subsurface temperature and current anomalies along the equator.

See the online version of this paper to view animation.
erly surface wind anomaly in the western Pacific. Namely, the thermocline shoals in the western Pacific due to the westerly surface wind anomaly, and as a result a negative subsurface temperature anomaly is observed at the thermocline depth. By the end of the year, the westerly surface wind anomaly disappears or significantly diminishes in the far-western Pacific. The elevated thermocline is then pushed back to the normal level, and this triggers the propagation of Kelvin waves along the thermocline. As they propagate eastward, the upwelling (negative) Kelvin waves erode the thermocline structure favorable for the maintenance of the positive sea surface temperature anomaly condition in the central and eastern Pacific. This terminates the warm phase of the biennial oscillation. It takes about 3 months for the Kelvin waves to reach the eastern boundary; then, the colder SST condition is observed in the eastern Pacific. In the winters of even years, the situation reverses. An easterly surface wind anomaly is observed in the far-western Pacific, and henceforth the thermocline deepens. The warmer subsurface water at the thermocline depth propagates eastward as downwelling Kelvin waves as the direction of the surface wind anomaly changes in the western Pacific by the end of the even year. Then, a warmer surface condition is observed in the following spring.

The propagation speed of the Kelvin waves as inferred from Movie 3 ($\sim 1.5 \text{ m s}^{-1}$) is much less than the theoretical speed because the Kelvin waves shown in Movie 3 are forced waves not free waves. In fact, the propagation speed inferred from Movie 3 represents the eastward movement of the region of ocean–atmosphere coupling, which generates the forced Kelvin waves. This speed is consistent with the eastward propagation speed of the surface zonal wind (Clarke and Shu, 2000). The wave mechanism associated with the biennial mode will be visited again in the next subsection.

The stochastic evolution history of the biennial mode is depicted in the lower panel of Figure 5. This time series shows how the biennial physics is modulated stochastically in the observational data. The PC time series of the biennial mode also has a broad spectral peak at the 5.3-yr period (Kim, 2002) implying that, on average, the biennial physics becomes strong at every 5.3 yr. This stochastic timescale should be distinguished from the physical timescale of 2 yr. The amplitude of the corresponding PC time series determines whether the biennial oscillation, more specifically whether the warm phase or the cold phase, is strong. Because of the 5.3-yr peak, a strong warm phase tends to be followed by a strong cold phase separated by 2.6 yr on average. This does not mean, however, that the biennial oscillation is an artifact of this alternate occurrence of the warm and cold phases with an approximately 2.6-yr interval. The lagged correlation of the PC time series of the biennial mode shows a strong and significant correlation ($\rho=0.67$) at the 12-month lag. Thus, there is a strong tendency for a warm phase to be followed by a cold phase.

### 4.2.1 Sea level height change associated with the biennial mode

Movie 4 shows the sea level change associated with the biennial mode. As a sea surface temperature anomaly rises during the warm phase of the biennial evolution, a westerly surface wind anomaly develops in the western and central Pacific
Movie 4. The surface and subsurface variability associated with the biennial mode. (upper right) Surface temperature and surface wind changes and (lower right) subsurface temperature and current changes. (upper left) Kelvin component and (lower left) the first Rossby component of sea level height change associated with the biennial mode.

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(see August, odd year) due to the positive feedback relationship between the two physical variables. This westerly surface wind anomaly produces positive (downwelling) Kelvin waves as shown in the upper-left panel. In fact, forced Kelvin waves are continuously generated as a result of ocean–atmosphere coupling. The deepening of the thermocline in the central and eastern Pacific manifests the continuous generation of forced Kelvin waves (lower-right panel). The westerly surface wind anomaly in the western Pacific also generates negative first Rossby waves as described in the lower-left panel. The first Rossby waves propagate westward and some of wave energy appears to be reflected as negative (upwelling) Kelvin waves from the western boundary (upper-left panel), which consists of the multitude of Indonesian islands. These upwelling Kelvin waves could not have been generated from the central Pacific since Kelvin waves propagate eastward. The presence of the upwelling Kelvin waves is also shown as shoaling of the thermocline in the western Pacific (lower-right panel). The shoaling of the thermocline is also accelerated by the westerly surface wind anomaly in the western Pacific. The evolution of the zonal surface wind anomaly, sea surface temperature anomaly, sea level height anomaly, and the 20° isotherm depth during the 1997/98 El Niño is reasonably similar to Movie 4 (see McPhaden, 1999; McPhaden and Yu, 1999).

As the westerly wind anomaly subsides in November and December, upwelling Kelvin waves propagate eastward and they erode the deeper thermocline
structure associated with the surface warming. This appears to terminate the warm phase of the biennial oscillation. By the spring of the following year, upwelling Kelvin waves reach the eastern Pacific and then a colder surface condition is observed in the eastern Pacific initiating the cold phase of the biennial oscillation (see April, even year). The description above is essentially that of the delayed-action oscillator mechanism. The observational data seem to suggest that the biennial mode is strongly associated with the delayed-action oscillator hypothesis. It should be noted that the wave activity in the low-frequency mode (first mode) is much weaker than that shown in Movie 4. Further, the delayed-action oscillator hypothesis is not seen in the low-frequency mode (figure not shown).

A close examination of Movie 4 indicates that the direction of the surface wind anomaly changes by October in the far-western Pacific (see October, odd year). This easterly surface wind anomaly is capable of producing upwelling Kelvin waves. The generation of upwelling Kelvin waves forced by the easterly surface wind anomaly is the main feature of the so-called western Pacific oscillator hypothesis (Weisberg and Wang, 1997; Wang and Weisberg, 2000). It appears, therefore, that both hypotheses are consistent with the observational data. It is not clear, however, how much of the upwelling Kelvin waves in the western Pacific in the winter of the warm phase is due to the reflection of Rossby waves and how much is due to the direct wind forcing. The magnitude of upwelling Kelvin waves due to direct wind forcing, in theory, can be estimated by using the surface wind forcing. As can be seen in Movie 4, however, the Kelvin and Rossby decomposition of the sea level height anomaly could not be completed between 125° and 145°E because of New Guinea. As will be discussed below, however, the direct wind forcing in the western Pacific seems to be important and distinguishes the biennial mode from the low-frequency mode.

4.2.2 Atmospheric responses associated with the biennial mode

Movie 5 shows the atmospheric circulation change associated with the biennial mode. As sea surface temperature rises in the central and eastern Pacific, two positive geopotential height anomaly patterns, one on each side of the equator, are developed at the 200-hPa level (see December, odd year). This geopotential height anomaly pattern is related to the easterly wind anomaly at the 200-hPa level in the eastern-central Pacific, which, in turn, appears to be associated with the westerly surface wind anomaly in the western Pacific. The vertical structure of wind along the equator shows increased convection in the central and eastern Pacific (upper-right panel). There is a slow downward movement over the western Pacific and the eastern Indian Ocean. During the cold phase, the direction of the circulation anomaly reverses with decreased convection in the central and eastern Pacific and weak upward motion in the western Pacific. It appears that the tropical Pacific sector and the tropical Indian sector are not strongly connected in terms of circulation features, and the western Pacific (130°-160°E) serves as a boundary zone between the two distinct zonal circulation regimes (see upper panels in December in Movie 5). The two physically separated circulation regimes, however, change in concert with each other, an indication that the two share a common physical mechanism, which needs to be identified and studied further. At the
developing stage of the warm and cold phases of the biennial oscillation, the boundary between the two zonal circulation regimes is displaced farther westward over the eastern Indian Ocean (see July, odd and even years; see also Movie 9). As each phase matures, this boundary, which is characterized by a slow downward (upward) movement of air during the warm (cold) phase of the biennial oscillation, migrates eastward into the western Pacific (see December, odd and even years). The eastward movement of this downward (upward) branch of the zonal circulation anomaly cell is also seen in the velocity potential and the divergence of the low-level (700 hPa) wind anomaly during the warm (cold) phase (see Movie 6).

During the mature stage of the warm and the cold phases of the biennial oscillation, the 200-hPa geopotential height anomaly shows two distinct anomaly patterns: a “zonally symmetric” pattern along the equator and a “zonally asymmetric” pattern at around 140°W (Yulaeva and Wallace, 1994). Both the symmetric and the asymmetric patterns develop strongly in winter. The asymmetric pattern, which appears to be the first Rossby structure, culminates in January and disappears in the late spring. It appears that the zonally asymmetric feature is primarily responsible for the physical separation of the two distinct circulation regimes. The symmetric pattern, which seems to be due to Kelvin waves, persists.

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a little longer and appears to lag the maximum SST change in the central and eastern Pacific by about 2 or 3 months (Yulaeva and Wallace, 1994).

The asymmetric response is associated with the upper-level anticyclonic (cyclonic) circulation at around 140°W during the warm (cold) phase, as is also implied in the lower-right panel of Movie 5. The off-equatorial branch of these cyclonic or anticyclonic circulations almost coincides with the axis of the midlatitude jet stream. It is then not difficult to understand that the asymmetric response can be a source of variability of the jet stream in the Western Hemisphere. This appears to be an important mechanism of teleconnection in midlatitudes as shown in the lower-right panel of Movie 5 (see December). It should be noted that the midlatitude response to the equatorial biennial oscillation involves the entire atmospheric column and contains a significant barotropic component. There is also a strong asymmetry between the summer hemisphere and the winter hemisphere; the response is stronger in the winter hemisphere. This asymmetry, together with the seasonal asymmetry of the biennial oscillation, makes its responses much stronger in the boreal winter hemisphere than in the boreal summer hemisphere.

The precipitation anomaly and the diagnostically analyzed diabatic heating (Hoskins et al., 1989) anomaly patterns associated with the biennial mode are plotted in Movie 6. As sea surface temperature rises in summer, a positive diabatic heating anomaly develops in the central Pacific (see July, odd year). The center of the diabatic heating is not over the eastern Pacific, where a maximum SST
change occurs; rather it appears in the central Pacific within the warm pool, where
the sea surface temperature is high enough for deep convection. Also, a negative
diabatic heating anomaly appears in the far-western Pacific and in the eastern
Indian Ocean, which indicates the decreased convective activity in the region.
The precipitation field in the western Pacific and the eastern Indian Ocean is quite
consistent with the diabatic heating anomaly. In the central Pacific, however, the
precipitation field does not respond quickly to the diabatic heating change, and
there is only a weak sign of the increased precipitation in the early stage of the
biennial oscillation.

As the sea surface temperature increases further in the eastern Pacific by
December of odd years, the diabatic heating anomaly strengthens and gradually
spreads toward the western Pacific. The precipitation anomaly pattern is quite
consistent with that of the diabatic heating change. This diabatic heating change
is often associated with the strengthening and the eastward movement of the
westerly surface wind anomaly (Clarke, 1994; Wang and Weisberg, 2000; Clarke
and Shu, 2000), which, indeed, is clear in Movie 4. The maximum deep convec-
tion, to the west of which is observed a strong westerly surface wind anomaly,
is still located in the central Pacific. The strengthening of the westerly surface
wind anomaly appears to be consistent with the increase of the convective activity
in the central Pacific. It is difficult to reason, however, that the eastward movement
of the westerly surface wind anomaly is due to the deep convection. The center
of the deep convection, as postulated from the diabatic heating anomaly pattern,
does not seem to shift toward the eastern Pacific. The center of the outgoing
longwave radiation anomaly field depicted in Wang and Weisberg (Wang and
Weisberg, 2000) also does not show any significant eastward movement in the
evolution stage of the 1997/98 El Niño.

The precipitation anomaly pattern is generally consistent with that of the
heating anomaly over the western Pacific and the eastern Indian Ocean, especially
during the mature phases of the biennial oscillation. During the mature stage of
the warm (cold) phase, the upward (downward) branch of the zonal circulation
is also formed in the western Indian Ocean as well as in the central and eastern
Pacific (December, Movie 6). Thus, a precipitation anomaly tends to have the
same sign in the central and eastern Pacific and in the western Indian Ocean. The
western Pacific and the eastern Indian Ocean maintain the precipitation anomaly
of opposite sign.

The velocity potential and the divergence of the low-level (700 hPa) wind
anomaly described in Movie 6 (right panels) seem to suggest that the increased
sea surface temperature anomaly and the resulting increase in deep convection in
the central Pacific are responsible for the zonal circulation change in association
with the biennial oscillation. In the summer of the warm phase, the negative
velocity potential and the convergence tend to form an upward branch of the
zonal (Walker) circulation in the eastern-central Pacific, where an excessive dia-
batic heating anomaly is observed (July, odd year). The downward branch is
located between the eastern Indian Ocean and the western Pacific Ocean. As the
sea surface temperature anomaly increases, both the upward and the downward
branches move eastward to their seasonal mean position (December, odd year).
Movie 7. The atmospheric variability associated with the low-frequency mode. (lower left) Surface air temperature and surface wind changes and (upper left) geopotential height and wind changes at the 200-hPa level. (upper right) Atmospheric condition in the vertical plane along the equator and (lower right) vertical plane along 160°W, which is the center of action. The vertical velocity is exaggerated by two orders of magnitude.

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This seems to indicate that the seasonal mean zonal Walker circulation is involved in the circulation change over the tropical oceans.

### 4.3 A two-mode view

The circulation change associated with the low-frequency mode is depicted in Movie 7. One of the main features that distinguish the low-frequency mode from the biennial mode is the weaker zonal circulation anomaly in the equatorial plane (upper-right panel; December, odd year). The weakened convective activity at 160°W near the equator is also evident in the lower-right panel (see Movie 5 for comparison). The vertical component of the wind change in the low-frequency mode is much weaker than in the biennial mode specifically in the central Pacific. It appears that the circulation boundary between the two basins is not as clear as in the biennial mode.

Movie 8 also shows the weakened convective activity in the low-frequency mode (December, odd year). A comparison between Movie 6 and Movie 8 indicates that the diabatic heating anomaly in the low-frequency mode is much weaker than the biennial mode both in the central Pacific (upward branch) and in the western Pacific (downward branch). This divergence and the velocity potential of the low-level wind anomaly show the decreased zonal circulation anom-
aly as well. The decreased subsidence in the western Pacific and the eastern Indian Ocean is also reflected in the weakened zonal circulation anomaly in the Indian sector. It is uncertain in the present results why the convective activity significantly weakens in the low-frequency mode while the sea surface temperature anomaly in the central to eastern Pacific is comparable to that of the biennial mode. It should be pointed out that the meridional component of the low-level divergence is also much weaker in the low-frequency mode.

Movie 9 depicts the streamlines of the wind anomaly in the equatorial vertical plane associated with the low-frequency mode and the biennial mode. It is clear that the circulation patterns of the two modes are quite different in the vertical plane. A cyclonic (anticyclonic) circulation pattern with well-defined upward and downward branches is shown in the tropical Pacific during the mature warm (cold) phase of the biennial mode (see winter, odd and even years). On the other hand, an anticyclonic (cyclonic) circulation pattern is observed in the tropical Indian Ocean during the warm (cold) phase. The two circulation regimes are separated between 120°E and 160°E. A well-defined cyclonic circulation, however, is not observed in the low-frequency mode. The vertical component of the wind anomaly in the low-frequency mode is weaker than in the biennial mode in the central and eastern Pacific and in the western Pacific.

Movie 9 also shows that the zonal circulation anomaly is stronger in the biennial mode than in the low-frequency mode. This may be associated with stronger diabatic heating in the biennial mode in the winter (see December pat-
Movie 9. The streamline of the wind anomaly in the equatorial vertical plane (top) associated with the low-frequency mode and (bottom) that associated with the biennial mode. Red represents the westerly wind anomaly whereas blue, the easterly wind anomaly. The darkness of the color is proportional to the magnitude of the zonal wind anomaly.

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terns in Movie 6 and Movie 8). The strong zonal circulation anomaly patterns associated with the biennial mode seem to be responsible for a strong surface wind anomaly in the far-western Pacific, which is instrumental for the generation of Kelvin waves. Movie 6 and Movie 8 show that the lower-level wind in the
far-western Pacific is much stronger in the biennial mode than in the low-frequency mode. The Kelvin waves induced by the surface wind anomaly, in turn, switch the phase of the SST oscillation from warm to cold and vice versa. It is important to note that the boundary of the two zonal circulation anomaly patterns migrates eastward during the winter; this implies that the direction of the surface wind anomaly changes significantly in the far-western Pacific. The increased zonal circulation anomaly also implies that the generation of the first Rossby waves in the western-central Pacific is strong and, as a result, reflected Kelvin waves from the western boundary will also be strong. Even though the low-frequency mode shows the sign of a westerly surface wind anomaly in the far-western Pacific, it is not strong enough to reverse the thermocline structure established in the tropical Pacific and the biennial oscillation will not occur.

In the context of CSEOF analysis, ENSO variability can be represented in terms of two dominant modes: the biennial mode and the low-frequency mode. The two modes have very different physical and dynamical mechanisms of evolution associated with them. As shown above, the biennial mode exhibits a strong physical evolution phase-locked to the annual cycle. Strong wave activity is evident in the extracted patterns of the sea level change and the thermocline fluctuation associated with the biennial mode. The low-frequency mode, on the other hand, does not show any significant physical evolution or a strong phase-locking tendency. The low-frequency mode appears to be a standing mode in the Pacific basin and its magnitude is determined primarily by the associated low-frequency stochastic undulation.

The separation of the two modes has an important implication for understanding the nature of ENSO and its predictability. For example, the biennial mode and the low-frequency mode have different natures of predictability. The deterministic component of evolution (tendency from a warm phase to a cold phase and vice versa) in the biennial mode is readily predictable, whereas the low-frequency mode is of a stochastic nature and its prediction is limited to its inherent predictability. Also, inter-ENSO variability can be explained in terms of an irregular interplay of the two modes. They are computationally independent modes, and henceforth, their evolution is independent of each other. As shown in Figure 5, the relative peak positions in the PC time series of the two modes are rather erratic, suggesting that the contribution of the two modes to SST change varies from one ENSO event to another. If the phases of the two modes are identical, then a stronger than normal ENSO event might be expected. If the two modes are out of phase, then an ENSO event will be weaker. The onset and termination times of El Niño and La Niña are also affected by the irregular interplay of the two modes; this may partially explain why ENSO events are loosely phase-locked to the annual cycle.

5. Summary and conclusions

In this study, the principal modes of SST variability were investigated using CSEOF analysis. The first two dominant modes, the low-frequency mode and the biennial mode, explain slightly less than 50% of the total SST variability in the tropical Pacific, and seem to describe two distinct components of variation as-
associated with the evolution of El Niño and La Niña. The patterns of other variables, which are consistent with the principal sea surface temperature anomaly modes, were determined from a regression analysis in the CSEOF domain. In this way, the physical and dynamical structures of the dominant modes of ENSO were extracted from the depth of the ocean to the tropopause.

The low-frequency mode appears to be a standing mode and does not exhibit any strong evolution during its life cycle. This mode seems to be of stochastic origin in that its temporal variation is primarily associated with its PC time series. The main structure of the low-frequency mode is the increased westerly surface wind anomaly in the western and the central Pacific and the resulting SST increase in the central and the eastern Pacific. The subsurface temperature anomaly and the thermocline structures are consistent with the surface wind anomaly and sea surface temperature anomaly patterns, with the shallower thermocline and henceforth the decreased subsurface temperature at the thermocline depth in the western Pacific and the deeper thermocline and the increased subsurface temperature in the eastern Pacific. Equatorial waves are fairly weak in the low-frequency mode. It appears that the low-frequency mode is the response of the equatorial coupled circulation system with the basinwide change in the surface wind. Although the interdecadal variability in the North Pacific is related to the prolonged change in the surface wind, it is not the sole origin of low-frequency mode. The high-pass filtering (< 6 yr) of the data still shows a significant amount of variability associated with the low-frequency mode.

The biennial mode represents a 2-yr oscillation of the sea surface temperature anomaly condition with the successive change between the warm phase and the cold phase. The development of Kelvin waves in the western Pacific and their propagation toward the eastern Pacific compose the basic mechanism of evolution associated with the biennial mode. The directional change of the surface wind anomaly in the far-western Pacific in the winter is instrumental for the alternation of the two phases of the biennial oscillation. Each phase seems to be tightly phase-locked to the annual cycle. It appears that both the delayed action oscillation hypothesis and the western Pacific oscillator hypothesis are consistent with the evolution of the biennial mode.

Although the spatial patterns of the two modes are rather similar in winter, when the two modes tend to be largest, different evolution separates the two modes. Because of the disparate evolution, the dynamical and thermodynamical responses of the two modes differ in many respects. For example, the biennial mode induces a stronger response dynamically and thermodynamically in winter even though it is weaker than the low-frequency mode in the SST variability. On the one hand, a weak evolution (deterministic component) in the low-frequency mode makes it difficult to predict. On the other hand, the strong evolution in the biennial mode is readily predictable. These considerations seem to be important in the development of an accurate ENSO forecasting technique (Clarke and van Gorder, 2001).

Some of the inter-ENSO variability also can be interpreted in the context of the two-mode view. A loose phase-locking tendency, variable duration of SST warming or cooling, and different onset and termination times of El Niño and La Niña seem to be reasonably explained in terms of an irregular interplay of the two modes.
Finally, it should be mentioned that the two principal computational modes may not necessarily be true physical modes. Significant uncertainty comes from the assumption of the 24-month nested period, which cannot be proven in a rigorous manner. It is possible that the two modes represent two different facets of the same physical process. In light of this uncertainty, it cannot be stated firmly that the biennial mode has an exactly 2-yr period. In this sense, it may be more judicious to call this mode a “biennial tendency.” It is undoubtedly true, however, that this biennial tendency is seen in various physical parameters, and this component explains a significant portion of variability.

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Appendix A: A Cyclostationary Representation of Data

In CSEOF analysis, space–time data are written as

$$T(r, t) = \sum_n T_n(t)B_n(r, t), \quad (A1)$$

where $T(r, t)$ are the data, $B_n(r, t)$ are cyclostationary loading vectors (hereafter, simply CSEOFs), and $T_n(t)$ are the corresponding principal component (PC) time series. Space–time data are linearly decomposed into CSEOFs as in EOF analysis, but the eigenfunctions in the former analysis technique are functions of time as well as space. Eigenfunctions are periodic in time, that is,

$$B_n(r, t) = B_n(r, t + d), \quad (A2)$$

where $d$ is called the nested period. The periodicity of the eigenfunctions comes from the periodicity assumption of the covariance function, namely,

$$C(r, t; r', t') = \langle T(r, t)T(r', t') \rangle = C(r, t + d; r', t' + d), \quad (A3)$$

where $\langle \cdot \rangle$ denotes ensemble averaging. The periodicity of covariance statistics is an assumption although it may not be a bad one for many geophysical variables. It should be emphasized that this periodicity is assumed to be one (in units of sampling time step) in EOF analysis.

It is important to distinguish between two different sources of temporal variability in the CSEOF representation; the time dependence in CSEOFs represents deterministic evolution and the temporal evolution in a PC time series is of stochastic (random) origin (Figure A1). For example, Kim and Chung (Kim and Chung, 2001) analyzed the sea surface temperature field (not sea surface temperature anomaly) in the tropical Pacific with a nested period of 12 months. The first mode, $B_1(r, t)$, describes the annual cycle (see Movie A1) and the corresponding PC time series, $T_1(t)$, shows the stochastic undulation of the seasonal cycle. These two sources of temporal variability are physically and dynamically distinct and should be distinguished in drawing statistical inferences on a dataset. For example, the PC time series of the first EOF of the same SST field depicts
essentially the physical evolution of the annual cycle; the stochastic undulation of the annual cycle is masked by the strong physical evolution (Kim and Chung, 2001). In this simple example, the distinction of the deterministic component of evolution from the stochastic component of evolution is clear and does not pose a serious problem. In many cases, however, drawing such a distinction based on EOF PC time series is difficult.

The presence of time-dependent physical processes in a dataset is reflected in the temporal dependence of the corresponding covariance statistics. This can be shown as follows. Let us suppose that a space–time data can be written in terms of mutually independent physical modes as in Equation (A1). Then, the covariance function of $T(r, t)$ can be written as

$$C(r, t; r', t') = \langle T(r, t)T(r', t') \rangle = \sum_n \sum_m \langle T_n(t)T_m(t) \rangle B_n(r, t)B_m(r, t)$$

$$= \sum_n R(|t - t'|)B_n(r, t + d)B_n(r', t' + d)$$

$$= C(r, t + d; r', t' + d), \quad (A4)$$

where the autocovariance function, $R(t)$, is defined as

$$R_n(|t - t'|) = \langle T_n(t)T_n(t') \rangle \quad (A5)$$

and the last line in Equation (A4) follows from the independence of the PC time series, $T_n(t)$, and the periodicity of CSEOFs. The temporal dependence of the covariance function comes from the time-dependent physical processes and not from the PC time series. The autocovariance function of the PC time series, which is regarded as purely stochastic, is not a function of time but is a function of lag, $\tau = |t - t'|$, only. Therefore, the time dependence of the covariance statistics implies that there are time-dependent physical processes. Equation (A4) also shows that the covariance function is periodic if all the physical processes are periodic with the same period, $d$; this period is called the nested period.

The periods of physical processes may not be identical in reality. Therefore, the nested period, $d$, should be understood as the least common multiple of all
Movie A1. The annual cycle of the tropical Pacific sea surface temperature data. This represents the first CSEOF with a nested period of 24 months. Note that the physical evolution is essentially identical in the first year and the following year indicating that the physical period of the mode is 12 months instead of 24 months.

See the online version of this paper to view animation.
the periods associated with the physical processes responsible for the data. For example, Movie A1 shows the first CSEOF of the tropical Pacific sea surface temperature with a 24-month nested period. The evolution of the spatial patterns indicates that this mode is the annual cycle. Since the spatial pattern in the first year is identical with that in the following year, the physical period of the mode is indeed 12 months instead of 24 months. Prior to a CSEOF analysis, physical periods are typically unknown, and the only guidance is the nested period. Because of the stipulation of the nested period, all CSEOFs have the same apparent period, but their physical periods are not necessarily the same.

Appendix B: Some Sensitivity Tests

Let us consider artificial data:

\[ T(t) = B_1(t)S_1(t) + B_2(t)S_2(t), \]  

(B1)

where \( B_1(t) \) and \( B_2(t) \) are two deterministic functions and \( S_1(t) \) and \( S_2(t) \) are two stochastic time series. Two experiments are carried out in which deterministic functions are given by

\[ B_1(t) = 1 \quad B_2(t) = \cos(2\pi t/12), \quad \text{experiment 1}; \]
\[ B_1(t) = 1 \quad B_2(t) = \cos(2\pi t/24), \quad \text{experiment 2}. \]  

(B2)

Note that the two deterministic components are orthogonal to each other. The stochastic time series are identical with those shown in Figure 5. Note, in the first experiment, that both \( B_1(t) \) and \( B_2(t) \) are periodic with a period of 12 months; therefore, the covariance function of Equation (B1) should also be periodic with the same period. In the second experiment, the period of covariance statistics is 24 months. Then, CSEOF analysis was conducted on both datasets with a nested period of 24 months.

The first two CSEOFs of each experiment are shown in Figure B1. The first two modes explain almost all of the variance (> 97%). As shown in Figure B1, the structures of \( B_1(t) \) and \( B_2(t) \) were reasonably extracted by the leading CSEOFs. The resulting PC time series are highly correlated (\( \rho > 0.97 \)) with the corresponding stochastic time series \( S_1(t) \) and \( S_2(t) \).

As demonstrated in these experiments, CSEOF analysis is not a filtering technique. Essentially all of the variance in the raw data was captured by the leading two modes as should be expected. Note also that the 24-month nested period does not force CSEOFs to have a 2-yr period. As shown in Figure B1 (left panel), the first mode is essentially flat (period of 1 month), and the second mode has a 12-month period. This point has been addressed in detail in appendix A.

Indeed, it is difficult to find a true period for the covariance function from limited observational data. Neither is it possible to argue that the covariance function should be periodic for geophysical parameters. Many physical processes are most likely not periodic, especially on short timescales. The periodicity of the covariance statistics is simply an assumption based on the hypothesis of (quasi-) biennial oscillation (Rasmussen et al., 1990; Clarke and van Gorder, 1999; Clarke and van Gorder, 2000; Weiss and Weiss 1999; among others).
Figure B1. The first two CSEOFs of (a) expt 1 and (b) expt 2.
Figure B2. (a) Cross correlogram and (b) coherence spectrum of the leading two EOF PC time series of the tropical Pacific SSTA data.
The 24-month period of the covariance function of the Niño-3 time series was tested by Kim (Kim, 2002). Figure B2 shows the cross correlogram and the coherence spectrum of the first two EOF PC time series of the tropical Pacific sea surface temperature anomaly data. The cross correlogram shows that the leading two EOF PC time series are fairly correlated at some nonzero lags. The significance of correlation exceeds the 95% level. The coherence spectrum also shows that the two time series are highly correlated in the neighborhood of a 24-month period. Although these tests, including Kim (Kim, 2002), indicate some periodic variability in the neighborhood of 24-month period, test results are certainly not conclusive, specifically in terms of the true period of oscillation.

References


