Formation of the Cold Tongue and ENSO in the Equatorial Pacific Basin

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ABSTRACT

This paper proposes a mechanism that explains how coupled dynamics alone can spontaneously give rise to a realistic west–east asymmetric mean state and an ENSO-like interannual variability without requiring the existence of an external preexisting west–east asymmetry in circulation. The essence of the newly proposed mechanism is that the basinwide ocean–atmosphere coupling acts to reduce the effective restoring force. As a result, the coupled oceanic waves travel more and more slowly within the equatorial ocean basin as the coupling strength increases. When the coupling strength reaches a critical value, the zonally leveled thermocline becomes unstable as a result of the weakening of the effective restoring force, at which the theoretical limit of the traveling timescale would be infinite without nonlinearity. Due to nonlinearity in the coupled system, the primary air–sea interaction instability leads to a west–east asymmetric mean state in which the atmosphere has a prevailing easterly and the ocean basin has a deep-in-west–shallow-in-east thermocline with a warm-west–cold-east sea surface temperature. The direction of the west–east asymmetry in the mean state is dictated by a planetary factor of the earth, namely, that the Coriolis parameter changes sign at the equator. As the coupling strength further increases, the asymmetry in the mean state amplifies and the phase speeds of the coupled equatorial oceanic waves begin to decrease gradually toward an asymptotic limit equal to the full speed in the uncoupled situation.

Using the coupling coefficient that is consistent with the observation, the fully coupled model can produce a realistic mean state in which the basinwide SST (thermocline depth) difference is 4.2°C (116 m) and the westward wind stress at the central Pacific basin is 0.54 dyn cm⁻². The self-sustained oscillation has a primary period of 3.7 yr. The SST in the west (east) oscillates between 27.5°C and 28.5°C (between 25.2°C and 22.5°C).

1. Introduction

The equatorial Pacific ocean–atmosphere climate system is characterized by a zonally asymmetric mean state accompanied by an El Niño–Southern Oscillation (ENSO) of 3–4 yr. The mean thermocline in the west is much deeper than that in the east. The sea surface temperature (SST) has a warm pool in the west and a cold tongue in the east with a mean west–east SST difference of about 4–5°C. On top of the zonally asymmetric mean ocean state is a prevailing mean easterly wind of 6–7 m s⁻¹, which is nearly as large as the easterly trade wind that peaks at a few degrees of latitudes off the equator. The ENSO variability itself also exhibits a strong west–east asymmetry. In particular, the oscillation has larger amplitude at the eastern boundary (about 2°–3°C in SST and 30–40 m in thermocline depth) than that at the west (less than 1°C in SST and about 20 m in thermocline depth).

The center piece of the existing theories for ENSO lies upon the existence of a west–east asymmetric time mean state and is built in the framework of anomaly coupled dynamics (e.g., Cane and Zebiak 1985; Zebiak and Cane 1987; Schopf and Suarez, 1988, 1990; Suarez and Schopf 1988; Battisti and Hirst 1989; Philander 1990; Cane et al. 1990; Münich et al. 1991; Neelin and Jin 1993; Jin and Neelin 1993a,b; Wang and Fang 1996; Jin 1997a,b; Kang and An 1998; An and Kang 2000). Recently, the fully coupled dynamics has been studied to explain the west–east asymmetric mean state in the equatorial Pacific basin. In these studies, one of the two types of preexisting west–east circulation asymmetries has been incorporated in a fully coupled simple model to explain the observed mean climate state in the equatorial Pacific basin. The first one is a preexisting easterly wind residing over the equatorial Pacific basin (Neelin and Dijkstra 1995; Dijkstra and Neelin 1995; Cai 1995; Jin 1996; van der Vaart et al. 2000). As summarized in Dijkstra and Neelin (1999), the preexisting easterly wind acts to “seed an asymmetry” and the coupled dynamic feedbacks then act to amplify the asymmetric mean state to the level that is comparable to the observation. The second one is to consider thermocline feedback only at the eastern boundary (Sun and Liu 1996; Jin 1996; Liu 1997; Liu and Huang 1997). Cai (1995), Jin (1996), Dijkstra and Neelin (1999), and van der Vaart et al. (2000) further addressed the interannual ENSO-like variability embedded within the
self-generated asymmetric climate state (aside from the small portion that is attributed to a preexisting weak easterly wind). Cai (1995, hereafter C95) derived an analytic solution of a coupled system consisting of the first two equatorial oceanic waves and a conceptual atmospheric model. His analytic solution is essentially a damped oscillation about a mean state exhibiting an eastward-shoaling thermocline beneath a prevailing atmospheric easterly at the equator. The oscillation period of this solution is about 25 months when the asymmetry of the self-generated climate state is comparable with the observation. Using a simple box model of the coupled system, Jin (1996) found an oscillatory equilibrium solution that has a period of about 4 yr. Moreover, both the mean value and the amplitude of Jin’s oscillation solution are very reasonable compared to the observation. Dijkstra and Neelin (1999) and van der Vaart et al. (2000) used a two-step approach. The first step is to obtain the nonlinear equilibrium state of the fully coupled intermediate equatorial ocean–atmosphere model as in Neelin and Dijkstra (1995) and Dijkstra and Neelin (1995). The second step is to do an instability analysis on the self-generated equilibrium climate state. Their instability analysis reveals an unstable ENSO-like oscillatory mode superimposed onto the self-generated west–east asymmetric equilibrium state.

The question we attempt to address in this paper is the following: In the absence of an external preexisting west–east asymmetry in circulation, can the coupled equatorial ocean–atmosphere dynamics alone spontaneously give rise to both a realistic west–east asymmetric mean state and an ENSO-like interannual variability? Obviously, these two types of preexisting west–east asymmetries used in the literature are consistent with observations. The objective of this study is to propose a theory that explains the observed west–east asymmetry in the mean state and its accompanying ENSO-like interannual variability without taking advantage of “knowing” any observational factors that could directly attribute to a west–east asymmetry within the basin other than those planetary factors of the earth. By doing so, the “external asymmetric factors,” such as “the stronger thermocline feedback/coupling in the east,” would be a part of the factors to be explained by the theory itself. The two planetary factors that cannot be disregarded and may imply a west–east asymmetry are (i) that the Coriolis parameter changes its sign at the equator and (ii) that the equatorial Pacific basin is confined by two walls in the west and east.

The presentation of this paper is organized as follows. A brief description of the coupled system is presented in section 2. Section 3 reports the numerical solution of the fully coupled system. Discussed in section 4 are the mechanisms that are responsible for the self-generated coupled climatology and self-sustained oscillation in this fully coupled model. A summary follows in section 5. The sensitivity of the numerical solution of the fully coupled model to the numerical values of the key model parameters is discussed in the appendix.

2. The model

The coupled equatorial Pacific ocean–atmosphere model used in this study consists of (i) a ½ layer reduced-gravity linear shallow-water equatorial β-plane model with the long-wave approximation, (ii) a “stripped-down” version of the equatorial SST equation, and (iii) a conceptual basinwide Walker circulation atmospheric model. There are only two forms of nonlinearity in this coupled system. One is in the thermocline feedback and the other one is the Ekman layer feedback via a Heaviside function.

The governing equations for the ocean dynamics are

\[
\frac{\partial u}{\partial t} - \beta y v + g \frac{\partial h}{\partial x} = \frac{\tau}{\rho H}, \tag{1a}
\]

\[
b \beta y u + g \frac{\partial h}{\partial y} = 0, \tag{1b}
\]

\[
\frac{\partial h}{\partial t} + H \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) = 0, \tag{1c}
\]

with boundary conditions

\[
\int_{-L}^{L} u \, dy = 0 \quad \text{at} \quad x = 0, \quad \text{and} \quad \int_{-L}^{L} u \, dy = 0 \quad \text{at} \quad x = L; \tag{1d}
\]

where \( u \) and \( v \) are the depth-averaged zonal and meridional currents of the upper-layer ocean; \( L \) is the basin length; \( h \) is the departure thickness of the upper-layer ocean from a constant mean depth \( H \); and \( \tau \) is the zonal wind stress to be determined from the atmospheric model. The model parameters are \( \beta \) (meridional gradient of Coriolis parameter evaluated at the equator), \( g \) (reduced-gravity parameter), and \( \rho \) (mean density of the upper-layer ocean).

As in C95, a simple “conceptual” atmospheric model is used to predict the zonal wind stress \( \tau \):

\[
\tau = \tau_0 \sin \left( \frac{\pi x}{L} \right) \exp \left( -\left( \frac{y}{Y} \right)^3 \right), \tag{2}
\]

where

\[
\tau_0 = -\frac{\alpha A_0}{L} \left[ T_{\text{west}}(t) - T_{\text{east}}(t) \right]. \tag{3}
\]

In (3), \( \alpha \) is the nondimensional coupling coefficient and the product of \( \alpha A_0 \) is the coupling coefficient with dimension. Because the conceptual atmospheric model (2)–(3) only needs SST at the western and eastern boundaries, it suffices to solve the stripped-down equatorial SST equation (Neelin and Dijkstra 1995) at those two locations only, which can be written as...
$$\frac{dT_{bdn}}{dt} = 2\delta r_0 \left[ T_{bdn} - T_{so} - (T_0 - T_{so}) \tanh \left( \frac{h_{bdn}}{H^*} \right) \right] \times \mathcal{H}(-\tau_o) - \mu(T_{bdn} - T_0) = 0,$$

where, the subscript “bdn” stands for either “west” or “east.” The variables \( h_{bdn} \) and \( h_{east} \) are the thermocline displacements at the western and eastern boundaries, respectively, and they are predicted by (1c). The Heaviside function \( \mathcal{H}(\cdot) \) is defined as \( \mathcal{H}(\cdot) = 1 \) only if \( \tau_o \) < 0 and otherwise \( \mathcal{H}(\cdot) = 0 \). The factor of “2” in (4) is a result of using a zonally averaged wind stress in evaluating the upwelling effect. This particularly implies that the upwelling rate itself has been deliberately set to be identical in the west and east, excluding the potential west–east asymmetry due to the upwelling rate itself. But the effects of the upwelling can be different if a west–east-sloped thermocline emerges as a result of air–sea interaction. The remaining unknown, \( \tau_o \), is predicted by (3). We here add that keeping that \( \partial T/\partial t \neq 0 \) in (4) actually yields similar results. We invoke this simplification primarily for eliminating the SST modes a priori.

The numerical version of (1) used in this study is the one developed by Cane and Patton (1984) and adopted in Zebiak and Cane (1987) that solves (1) on a rectangular domain with a grid size of 0.5° lat × 1° lon. The model domain is rectangular and extends from \( Y_w = 28.75°S \) to \( Y_e = 28.75°N \) and from 124°E to 80°W (\( L = 17,316 \) km). The numerical values of the model parameters are: \( g' = 0.003 \) g (\( g \) is the earth gravity equal to 9.8 m s\(^{-2}\)), \( H = 145 \) m, and \( \beta = 2.28 \times 10^{-11} \) m\(^{-1}\) s\(^{-1}\). With these chosen values, the Kelvin wave speed is 2.07 m s\(^{-1}\) and the oceanic Rossby deformation radius is 300 km. We note that in principle the constant part of the thermocline depth, \( H \), should be predicted by the model instead of being specified as an external parameter. In other words, \( H \) should be determined by the collective effects of oceanic dynamics and thermodynamics. Understanding what the factors are that determine \( H \) in the equatorial Pacific basin is beyond the scope of this study. We here argue that the ocean–atmosphere coupled dynamics dictates the west–east asymmetry in the mean thermocline depth, but not the general level of the thermocline depth. The results to be presented below are not sensitive to the value of \( H \) as long as the corresponding free oceanic wave speeds are reasonable (the timescale of the ENSO is somewhat sensitive to \( H \) but not the strength of the asymmetry in the mean state).

The conceptual atmospheric model only predicts the wind stress in the central equatorial Pacific, representing a basinwide Walker circulation driven by a basinwide SST difference. The prescribed meridional structure of wind stress is intended to imply that the conceptual atmospheric model (2)–(3) is valid only within the equatorial latitude band. The parameter \( Y \) measures the meridional scale of the part of the wind stress that is coupled with the equatorial Pacific basin. Table 1 lists the standard value of \( Y \) used in this study. We will report the sensitivity with respect to the numerical value of \( Y \) in the appendix. The exact form of the prescribed zonal profile of the zonal wind stress is not critical as long as it has a broad basinwide scale. For example, the calculations with a zonally constant zonal wind stress basically reveal the same results, as one may expect. The dimensional portion of the coupling coefficient \( A_s/L \) has a preset value. We will present the solution to (1)–(4) as a function of coupling strength by varying the non-dimensional coupling coefficient \( \alpha \). The preset value of \( A_s/L \) is chosen in such a way that we will only need to scan \( \alpha \) from 0 (the uncoupled case) to 1 (the strongest coupled case). Physically, the coupling coefficient \( \alpha(A_s/L) \) measures the efficiency of the atmospheric engine fueled by a basinwide SST difference. With \( A_s/L \) equaling to 0.2 dyn cm\(^{-2}\) °C\(^{-1}\), we need to set \( \alpha \) equal to 0.65 to fit the observed mean wind stress 0.55 dyn cm\(^{-2}\) for a mean basinwide SST difference equal to 4.2°C. We will see in section 3, the fully coupled system indeed is able to produce a realistic climatology and an ENSO-like oscillation with this choice of \( \alpha \).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Values</th>
<th>Comments</th>
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<tbody>
<tr>
<td>( \delta )</td>
<td>10 (3 months × dyn cm(^{-2}))(^{-1})</td>
<td>For mixing layer depth of 50 m and 1 dyn cm(^{-2}) wind stress, this would yield an Ekman layer vertical motion of 0.006 cm s(^{-1})</td>
</tr>
<tr>
<td>( \mu )</td>
<td>1 (50 months(^{-1}))</td>
<td>Thermal damping rate*</td>
</tr>
<tr>
<td>( T_0 )</td>
<td>30° and 25°C</td>
<td>The range of SST is 20–30°C</td>
</tr>
<tr>
<td>( H^* )</td>
<td>145 m (= ( H ))</td>
<td>Control the sensitivity of thermocline feedback*</td>
</tr>
<tr>
<td>( A_s/L )</td>
<td>0.2 dyn cm(^{-2}) °C(^{-1})</td>
<td>Dimensional part of the coupling coefficient</td>
</tr>
<tr>
<td>( Y )</td>
<td>7.5° lat</td>
<td>Rossby deformation radius of equatorial atmosphere*</td>
</tr>
</tbody>
</table>

* Sensitivity to the numerical value of this parameter is reported in the appendix.

There are five parameters, namely, \( T_{so}, T_{so}, \delta, \mu, \) and \( H^* \) in (4). The parameters \( T_{so} \) and \( T_{so} \), respectively, are the zonally uniform sea surface and subsurface equilibrium temperature at the equator in the absence of any dynamics/coupling. The \( \delta \) is a thermocline feedback parameter measuring the upwelling strength produced by a unit westward wind stress. The \( \mu \) is the thermal damping rate, measuring the relaxation timescale of SST toward the prescribed equilibrium sea surface temperature \( T_{so} \). Here \( H^* \) is a reference depth controlling the sensitivity of the nonlinear dependence of subsurface temperature on thermocline depth. The standard numerical values of the five parameters used in this study are given in Table 1. The term associated with the Heaviside func-
tion represents the SST change due to Ekman upwelling dynamics. The change of SST due to meridional surface Ekman current is neglected for simplification. But note that all variables in (1)–(4) are total fields rather than anomaly fields. The observed total wind stress is westward almost everywhere along the equatorial Pacific basin, implying that the equatorial meridional Ekman currents are poleward at most places in compensating the upwelling there. Because poleward surface Ekman currents by themselves cannot alter the SST at the equator, the impact of neglecting meridional advection is expected to be minimal. The zonal advection is also neglected in (4) as in Neelin and Dijkstra (1995). The zonal advection term links the SST in the west with that in the east. The Ekman layer dynamics links the SST locally to the thermocline underneath. The change in thermocline depth is dictated primarily by the equatorial wave dynamics and is thought of far more importance than the advection term.

Note that the simple relationship between the thermocline depth and subsurface temperature used by Jin and Neelin (1993a) for an anomaly model and subsequently used by Dijkstra and Neelin (1995, 1999) in a fully coupled model has been modified in this study. Specifically, the argument of the hypertangential function in Jin and Neelin (1993a) is set to be \( (h' + h_0)/H^* \), where \( h_0 = 30 \) m, \( H^* = 25 \) m, and \( h' \) is the departure from a zonally asymmetric mean thermocline depth. The parameter \( h_0 \) is introduced to control the asymmetry of thermocline feedback. In this study, all model variables are the total fields. In that regard, it is not necessary to impose an external asymmetry factory in the parameterized thermocline feedback because the asymmetry in the feedback would be implicitly expressed by the variable \( h \) itself. Nevertheless, the results obtained by including \( h_0 = 15 \) m in (4) are qualitatively the same as the results with \( h_0 = 0 \). To avoid the potential confusion about the source of the west–east asymmetry in the mean state, we only report the results with \( h_0 = 0 \). As indicated in Table 1, we have also used a larger value of \( H^* \) in (4) because \( h \) is the thermocline depth departure from the preset constant thermocline depth \( H \). As a result, \( h/H^* \) in this study has the same magnitude as \( (h' + h_0)/H^* \) in Jin and Neelin (1993a) to retain the relation between thermocline depth and subsurface temperature numerically. As reported in the appendix, different values of \( H^* \) do not change behaviors of the solution significantly.

Equations (1)–(4) form a closed set of equations for the coupled equatorial Pacific ocean–atmosphere system. The only equilibrium solution of the system without coupling is the one in which the atmosphere is windless and the ocean has a zonally leveled thermocline with no current plus \( T_{\text{west}} = T_{\text{east}} = T_0 \) (the trivial solution). Because all model parameters are constant, particularly implying that they are zonally invariant, we can indisputably attribute the west–east asymmetric solution of (1)–(4), if it exists at all, to a form of coupled instability of the trivial solution. All numerical integrations with (1)–(4) are carried out to a minimum of 520 yr starting from the initial conditions summarized in Table 2.

### 3. Results

#### a. Overview of the solution

Figure 1 summarizes the behaviors of the coupled system as a function of the coupling coefficient \( \alpha \). There exist two critical values of the coupling coefficient that mark significant changes in the characteristics of the coupled solution. The first critical value is \( \alpha = \alpha_p \approx 0.575 \). Plotted in Figs. 2a–c are the numerical solutions with three representative values of \( \alpha < \alpha_p \). When air–sea interaction is disabled (\( \alpha = 0 \); Fig. 2a), the basinwide SST difference (dotted curve) dies out immedi-

<table>
<thead>
<tr>
<th>Model component</th>
<th>Initial conditions</th>
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<tbody>
<tr>
<td>Atmosphere*</td>
<td>( \tau_0 = 0 )</td>
</tr>
<tr>
<td>SST equation</td>
<td>( T_{\text{west}} = T_{\text{east}} = T_0 = 30^\circ \text{C} )</td>
</tr>
<tr>
<td>Ocean</td>
<td>Motionless ((u \text{ and } v = 0))</td>
</tr>
<tr>
<td></td>
<td>Constant depth ((h + H = 145 \text{ m}))</td>
</tr>
</tbody>
</table>

* A constant wind stress equal to \(-0.05 \text{ dyn cm}^{-2}\) is used initially for 3 months to “kick off” the system. This constant wind stress is reset to be zero for the remaining integration.

![Figure 1](http://journals.ametsoc.org/doi/abs/10.1175/1520-0442(2003)016<0144:FOTCTA>2.0.CO;2?md5=444e41f60e1f4b8ad144c420030f28d2)

**FIG. 1.** A summary diagram of the fully coupled model solution as a function of the (nondimensional) coupling coefficient (the abscissa). The curve with open circles is the oscillation period in months. The curve with solid circles is the time mean of \((h_{\text{west}} - h_{\text{east}})/2\) in meters. The curve with “+” is the amplitude of the self-sustained oscillation, measured by the difference between the maximum and minimum of \((T_{\text{west}} - T_{\text{east}})\) in 0.1°C. The dotted portion of the curves corresponds to a damped oscillation and the solid portion represents a self-sustained oscillation. The time series of the solution at the values of the coupling coefficient labeled “a,” “b,” “c,” “d,” “e,” and “f” are displayed in Fig. 2. The details of the solution with the coupling coefficient equal to the value labeled “O” are displayed in Figs. 3–5.
Fig. 2. The fully coupled model solution. (a)–(f) Cases with $\alpha = 0.0, 0.3, 0.5, 0.58, 0.62,$ and 1.0, respectively. The solid curve is the time series of $(b_{\text{west}} - b_{\text{east}})$ in meters. The dotted curve in (a)–(c) is the time series of $(T_{\text{west}} - T_{\text{east}})$ in (1/30)$^\circ$C.

Figures 2a and 2c illustrate the ocean–atmosphere system results. The basinwide thermocline depth difference (solid curve) displays a damped oscillation with a period of about 1 yr, an uncoupled timescale of equatorial oceanic waves within the Pacific basin. The decay of the oscillation in this frictionless ocean model can be easily explained by the loss of energy to the two boundaries (Cane and Moore 1981). The asymptotic behaviors of the solutions displayed in Figs. 2b and 2c are identical to Fig. 2a because the ocean–atmosphere system becomes uncoupled after a sufficiently long time due to the disappearance of a basinwide SST difference. This particular implies that the zonally symmetric state (or the trivial solution) is the sole equilibrium state for $\alpha < \alpha_0$. Nevertheless, the coupled dynamics results in a systematic change to the oceanic waves. It is seen in Fig. 1, the oscillation period (the curve with open circles) increases with the coupling coefficient for $\alpha < \alpha_0$. For example, the periods of the coupled system for $\alpha = 0$ (Fig. 2a), 0.3 (Fig. 2b), and 0.5 (Fig. 2c) are approximately equal to 1, 1.9, and 3.5 yr, respectively. As $\alpha$ approaches 0.58, the timescale becomes longer and longer (approximately up to 17 yr as revealed by the numerical solution) whereas the trivial solution remains to be the stable state of the coupled system (Fig. 1).

When $\alpha > \alpha_0$, the zonally symmetric state is unstable, resulting in a new equilibrium state in which the thermocline (SST) is deep (warm) in the west and shallow (cold) in the east, and the atmosphere has a prevailing easterly at the equator. The thermocline depth difference across the basin (the curve with solid circles in Fig. 1) increases as the coupling coefficient increases. In contrast to the cases of $\alpha \leq \alpha_0$, the oscillation period now decreases from its peak value at $\alpha = \alpha_0$, as the coupling coefficient increases. The second critical value of $\alpha$ is $\alpha = \alpha_1 \approx 0.61$, above which the oscillation is self-sustained (indicated by solid portion of the curves in Fig. 1). The amplitude of the self-sustained oscillation (the curve with cross in Fig. 1), measured by the difference between the maximum and the minimum of $(T_{\text{west}} - T_{\text{east}})$, is about 4$^\circ$C and changes little with the coupling coefficient soon after $\alpha$ exceeds $\alpha_0$. Figure 2d represents the solution between the two critical values of $\alpha$, characterized with a weak asymmetric mean state and a damped oscillation with a period of about 17 yr. Two examples of the self-sustained oscillation solutions for after $\alpha > \alpha_1$ are shown in Fig. 2e ($\alpha = 0.62$) and Fig. 2f ($\alpha = 1.0$). The primary periods of the solutions at these two values of $\alpha$ are 4.2 and 2.3 yr, whereas the mean thermocline differences across the basin are about 96 and 330 m, respectively.

b. The solution for the case of $\alpha = 0.65$

In section 2, we estimated that using $\alpha = 0.65$ in (3) would best fit the observed relation between the mean zonal wind in the central Pacific and the mean SST difference across the equatorial Pacific basin. Because of the fully coupling approach, we expect this value of $\alpha$ should yield a realistic reproduction of the prominent features of both the observed annual mean state and its accompanying ENSO variability in the equatorial Pa-

1 Note that we have intentionally set $\mathcal{H}(-\tau_e) = 1$ in Figs. 2b,c to estimate the oscillation timescale. Using $\mathcal{H}(-\tau_e) = 1$ only for $\tau_e < 0$, as stated in (4), we can estimate the timescale only from the first half cycle of the oscillation. This is because once the upwelling is turned off ($\tau_e \geq 0$), the SST at the two boundaries are immediately equal to $T_0$ as required by (4). As a result, the subsequent oscillation becomes uncoupled. Therefore, only the first half of the cycle records the effects of coupling on the oscillatory behavior of the oceanic dynamics. The coupled oscillation timescale can then be estimated alternatively as the twice of the first half-cycle timescale. This estimate yields nearly identical results. But we do not need to set $\mathcal{H}(-\tau_e) = 1$ in estimating the timescale of the oscillation after the coupled mean state emerges (e.g., Figs. 2d–f).
Fig. 3. Time series derived from the fully coupled model solution for \( \alpha = 0.65 \). (a) Wind stress at central Pacific basin (dyn cm\(^{-2}\)); (b) SST (°C) in the western (open circles) and eastern (solid circles) basins, and zonal mean thermocline depth (thin curve; its scale is omitted for brevity); (c) thermocline depth (m) in the western (open circles), in the eastern (solid circles) basins, and the zonal mean thermocline depth (thin curve, scale omitted); (d) zonal mean zonal current at the equator (solid) and the western boundary meridional current at 2.25°S (dashed) in m s\(^{-1}\).

The solution of the coupled model (1)–(4) with \( \alpha = 0.65 \) (Fig. 3) reveals that the mean SST in the west and in the east are 28.2° and 24°, respectively, resulting in a basinwide SST difference of 4.2°. The mean thermocline depth difference across the basin is 116 m. The corresponding mean wind stress in the central Pacific is -0.54 dyn cm\(^{-2}\). Superimposed on this coupled climate state is a self-sustained finite-amplitude oscillation with a primary period of about 3.7 yr. The basinwide thermocline depth (the thin solid curve in Figs. 3b,c) leads the depth/SST anomalies in the east by about 4(5) months during a cold (warm) phase. The latter, in turn, leads depth/SST anomalies in the west with an opposite sign by 3(5) months during a cold (warm) phase. The zonal mean zonal current leads the zonal mean depth by 2 months during a cold phase and increases up to 7 months in a warm phase (solid curve in Fig. 3d). The poleward meridional current at the western boundary lags the zonal mean zonal currents by 3 months for both warm and cold phases (the dashed curve in Fig. 3d).

The power spectral analysis reveals that the self-sustained oscillation is not a simple harmonic oscillation due to the nonlinearity (Fig. 4). In addition to the primary harmonics that has a period of 3.7 yr, there are subharmonics with periods of 1.85, 1.23, 0.925 yr, and so on. There also appears a timescale peaking at 37 yr in the spectrum diagram. Such a nonlinear oscillator forced with annual cycle may lead to chaotic behavior (Jin et al. 1994).

The amplitude of the zonal wind stress anomalies at the equator is about 0.23 dyn cm\(^{-2}\) (Fig. 5a). The equatorial thermocline depth anomalies (Fig. 5b) exhibit a quasi-stationary oscillation pattern with the node point about 20° west of the central Pacific basin. Moreover, the depth anomalies in the east lead the anomalies with opposite sign in the west by 3–5 months and the former are about twice as large as the latter (40 versus 20 m). Another point worth mentioning is that whereas the depth anomalies appear to have an eastward “propagation,” the zonal current anomalies (Fig. 5c) exhibit a
westward propagation as found in Cane and Sarachik (1981) and Philander and Pacanowski (1981).

4. Mechanisms of the coupled climatology and self-sustained oscillation

We now turn our attention to discuss the mechanisms for the coupled climate state and the accompanying self-sustained oscillation exhibited in the solution of (1)–(4).

a. What leads to the coupled zonally asymmetric climate state?

The answer to this question is the centerpiece of our newly proposed theory for the west–east asymmetry in the mean state of equatorial Pacific basin. This question is paired with the accompanying question of why the timescale of the coupled oceanic waves first increases and then decreases with the coupling coefficient after the coupled zonally asymmetric climate emerges, as illustrated in Fig. 1. The analytical solution reported in C95 reveals a similar dependency of the oscillation timescale on the coupling coefficient and the self-generated asymmetric mean state. Here we wish to use the analytical solution reported in C95 to address this question.

The analytical solution of C95 can be reproduced by introducing three approximations to the coupled model (1)–(4): (i) only considering the first two equatorial ocean modes of the ocean model, (ii) assuming that $(T_{\text{west}} - T_{\text{east}})$ is proportional to $\tanh[(h_{\text{west}} - h_{\text{east}})/H]$, and (iii) using a symmetric boundary condition instead of the asymmetric one, Eq. (1d), which is derived from the long-wave approximation. With these three approximations, Eqs. (1)–(4) become

$$ \frac{\partial K}{\partial t} + \frac{2X\hat{\alpha}}{\rho H\sqrt{c^2}\pi^{3/4} \cosh^2 \left( \frac{\Delta h}{H} \right)} \frac{\partial K}{\partial x} = 0 $$

$$ \frac{\partial R}{\partial t} - \frac{2X\hat{\alpha}}{\rho H\sqrt{c^2}\pi^{3/4} \cosh^2 \left( \frac{\Delta h}{H} \right)} \frac{\partial R}{\partial x} = 0 $$

where, $K$ and $R$, respectively, stand for the Kelvin wave and the first Rossby mode; and $X$ is the nondimensional basin length equal to $L/\sqrt{c^2}$, and $\Delta h = (h_{\text{west}} - h_{\text{east}})$, representing the equilibrium solution of the simplified coupled model. Other notations remain unchanged except that the variables “$x$” and “$t$” are nondimensional length and time, respectively. We use a different notation, $\hat{\alpha}$, to denote the coupling coefficient with a dimension of $[N \text{ m}^{-2}]$. Details of the derivation of (5) can be found in C95.

The term inside the bracket in (5a) for the Kelvin mode is identical to that for the Rossby mode (5b), which is expressed as

$$ G = \left[ 1 - \frac{2X\hat{\alpha}}{\rho H\sqrt{c^2}\pi^{3/4} \cosh^2 \left( \frac{\Delta h}{H} \right)} \right]. $$

The first term in (6) measures the natural restoring force of the uncoupled ocean, which has been nondimensionalized to be equal to 1. The second term is related to ocean–atmosphere coupling. We interpret $G$ expressed in (6) as the effective oceanic (nondimensionalized) restoring force toward the equilibrium state $\Delta h$. For a zonally leveled thermocline ($\Delta h = 0$), which is the sole equilibrium solution for a sufficiently small $\hat{\alpha}$ the effective oceanic restoring force becomes weaker and weaker ($G$ is smaller and smaller) as the coupling strength increases, causing coupled Kelvin and Rossby waves to travel slower and slower. This explains why
the coupled oscillation period increases with increasing coupling coefficient for a zonally leveled thermocline as shown in Fig. 1. The process of weakening the effective restoring force eventually leads to an instability as the coupling coefficient reaches a critical value, at which point the oscillation period would be infinite without nonlinearity. This critical value of the coupling coefficient can be determined by letting \( G = 0 \) and \( \Delta \bar{h} = 0 \) in (6), resulting in

\[
\bar{\alpha}_c = \frac{\rho H c^3 \beta \pi^{5/4}}{2X} = \frac{\rho H c^3 \pi^{5/4}}{2L}.
\]  

(7)

It is seen that this critical coupling strength is solely determined by the ocean model parameters that determine the equatorial oceanic waves' speeds.

Beyond the critical coupling strength, the coupled system, in the absence of nonlinearity, would no longer be able to oscillate but would grow exponentially because (6) yields \( G < 0 \) for \( \Delta \bar{h} = 0 \) and \( \bar{\alpha} > \bar{\alpha}_c \). The nonlinear development of air–sea interaction instability results in a zonally sloped thermocline equilibrium states (\( \Delta \bar{h} \neq 0 \)), as seen in Fig. 1. For the zonally asymmetric equilibrium solution \( \Delta \bar{h} \neq 0 \), \( G \) is always positive. This particularly implies that the equatorial oceanic waves now travel within the basin with a finite timescale along the sloped thermocline. The corresponding traveling timescale is a function of the climate state as well. According to (5), the effective restoring force of the coupled system increases toward the natural restoring force of the uncoupled system as the strength of the mean asymmetry amplifies. This explains why

the coupled oscillation period starts to decrease with an increasing coupling coefficient after the asymmetric state emerges as shown in Fig. 1.

It is of importance to point out that prior to the instability of the zonally symmetric solution shown in Fig. 1, the fully coupled system behaves just like an anomaly model. In this sense, the trivial solution is equivalent to an “anomaly-coupled solution,” which is a decayed oscillator prior to the instability and becomes an unstable stationary solution afterward. The slower propagation of coupled equatorial waves has been shown in the literature in the context of anomaly coupling (e.g., Hirst 1986). Jin and Neelin (1993a) further showed that the timescale of the ocean basin mode increases with the increasing coupling coefficient and eventually becomes infinite (Fig. 13c in Jin and Neelin 1993a). They described this behavior as the merging of a decay oscillatory basin mode with the stationary mode due to air–sea coupling. In the process of merging with the stationary mode, the ocean basin mode’s timescale becomes longer and longer. After the merging, the stationary mode continues to be unstable as the coupling coefficient further increases. Therefore, the growing portion in their Fig. 13c corresponds to the unstable portion of the trivial solution, which has a growth rate proportional to the absolute value of \( G \) with \( \Delta \bar{h} = 0 \).

Because we consider the problem in the context of the fully coupled system, we can further show that the instability of the anomaly-coupled solution leads to a birth of the zonally asymmetric mean state. Moreover, we can then examine how the timescale of the ocean basin modes is modified when they travel along the self-generated zonally sloped thermocline and how the otherwise decayed oscillation becomes self-sustained finite-amplitude oscillation.

b. Why is an opposite west–east asymmetry in the mean state not observed?

There are three ways that the SST can be changed by large-scale oceanic dynamics: (i) Ekman upwelling, (ii) meridional advection, and (iii) zonal advection. In this coupled model, we only consider the Ekman upwelling effects in the SST equation. Because the Coriolis parameter changes sign at the equator, a westward (eastward) stress would create an upwelling and poleward (downwelling and equatorward) Ekman drift along the equator. It follows that the SST is directly coupled with the thermocline change underneath only for an easterly wind at the equator. In other words, by definition, the air–sea interaction in the presence of Ekman layer dynamics favors an easterly wind over a westerly in the coupled system. As pointed out in the introduction, the fact that the Coriolis parameter changes sign at the equator is a planetary factor of the earth that implies a west–east asymmetry. Considering only this planetary factor and the planetary factor that the lands in the west and east confine the equatorial Pacific basin, we here have demonstrated that the atmosphere–ocean coupling alone can give rise to an equilibrium state that has the same west–east asymmetry configuration as in the real world.

c. Why does a damped oscillation become a self-sustained finite-amplitude oscillation?

When the coupling coefficient is smaller than the second critical value, the solution is a damped oscillation toward either a zonally level thermocline (prior to the instability) or an eastward-shoaling thermocline (after the instability). We observe that the damped oscillation has the same “turnaround” signal as the self-sustained finite-amplitude solution; namely, that the change in the zonal mean thermocline depth (or the zonal mean zonal current) leads to the depth anomaly in the east that, in turn, is lagged by the change in the depth in the west.

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2 Their coupled basin mode becomes unstable prior to its merging with the stationary mode. Our anomaly-coupled solution has an infinite period when it becomes unstable. This minor difference could be due to either differences in the model parameters or the fact that their coupled basin mode oscillates about a prescribed zonally asymmetric state whereas ours is about a zonally symmetric state that is the trivial solution of the model.
This suggests that the turnaround signal alone would not explain why there is a self-sustained oscillation.

Currently, we do not have an explanation that can be confirmed explicitly on what causes an otherwise damped oscillation to become a self-sustained finite-amplitude oscillation as the west–east asymmetry in the mean state becomes sufficiently strong. We attempt to attribute the existence of a self-sustained finite-amplitude oscillation in the coupled model to the delay oscillation mechanism. After the self-generated coupled west–east asymmetric thermocline state emerges, the thermocline feedback becomes more sensitive in the region where the mean thermocline is shallower, resulting in a further amplified west–east asymmetry in terms of the sensitivity of the oceanic response to wind stress. Based on delay oscillator theory (Suarez and Schopf 1988), we conjecture that this asymmetry in coupling would naturally give rise to a self-sustained oscillation via the delay oscillation mechanism. Direct proof of this conjecture requires additional work and is currently under way.

5. Summary

This paper presents a simple theory that explains how the basinwide ocean–atmosphere coupled dynamics can spontaneously give rise to the observed west–east asymmetric climate state in the equatorial Pacific basin and its accompanying interannual ENSO variability. This simple theory does not hinge upon a preexisting external factor in circulation that can be directly attributed to a west–east asymmetry. Between the two planetary factors of the earth that would imply a west–east asymmetry, the fact that the Coriolis parameter changes sign at the equator favors the observed west–east asymmetry configuration when the ocean is coupled with the atmosphere. The planetary factor, in which the landmass in the west and east confines the equatorial Pacific basin, plays a neutral role for the orientation of the west–east asymmetry although it is essential for the existence of a zonally asymmetric coupled mean state.

The coupled model system (1)–(4) consists of three components: (i) a linear reduced-gravity shallow-water equation on equatorial $\beta$ plane with the long-wave approximation (Cane and Patton 1984); (ii) a stripped-down equatorial SST equation governed by the Ekman layer and thermocline feedbacks (Neelin and Dijkstra 1995); (iii) a basinwide Walker circulation that is linearly proportional to a basinwide SST difference and the nondimensional portion of the proportional factor $\alpha$ is referred to as the coupling coefficient. In addition to the Heaviside function associated with the Ekman upwelling, the only other form of nonlinearity in this coupled system is the thermocline feedback. In this coupled system, the windless atmosphere, the zonally leveled thermocline with no current plus $T_{\text{out}} = T_{\text{out}} = T_0$ (the trivial solution) is the sole equilibrium solution of the system without coupling (where $T_0$ is the zonally uniform equilibrium temperature equal to 30°C). Because there are no external thermodynamic/dynamic factors of any kind that would introduce a preexisting west–east asymmetry, we can indisputably attribute the west–east asymmetric solution of (1)–(4) to the coupled instability of the trivial solution.

This coupled model is capable of reproducing the most salient aspects of the observed climatology and ENSO of the equatorial Pacific ocean–atmosphere system in the neighborhood of $\alpha = 0.65$. This value of the coupling coefficient linearly fits the relationship between the mean wind stress and the basinwide SST difference in the observation. For $\alpha = 0.65$, the coupled climate state has a basinwide SST (thermocline depth) difference equal to 4.2°C (116 m) and a westward wind stress at the central Pacific equal to 0.54 dyn cm$^{-2}$. The corresponding self-sustained oscillation has a primary period equal to about 3.7 yr. The SST at the west oscillates between 27.5°C and 28.5°C whereas the SST at the east oscillates between 25.2°C and 22.5°C. The thermocline depth difference across the basin oscillates from 68 m during an El Niño event to 175 m during a La Niña event. Accompanying the oscillation in the ocean, the maximum westward wind stress changes from 0.32 to 0.78 dyn cm$^{-2}$ from an El Niño to a La Niña event. The development of an El Niño (or La Niña) event can be traced back to a change in the zonal mean thermocline depth. About 4–5 months later, the SST/depth anomalies in the eastern basin start to reverse sign, which is followed by a change in wind stress and then a change in the western SST/depth anomalies with opposite sign. The self-sustained oscillation also exhibits a significant mismatch in space. The quasi-stationary depth anomalies have a node point about 20o west of the central Pacific basin. Moreover, the eastern SST/depth anomalies are 2–3 times as large as the anomalies in the western basin.

The cornerstone of this simple theory is the reduction of the effective restoring force due to basinwide air–sea coupling, as first suggested in C95. The weakening of the effective restoring force due to air–sea coupling is equivalent to the phenomenon that the ocean basin mode gradually merges with the stationary mode and becomes unstable as the coupling coefficient increases, as first described in Jin and Neelin (1993a). Both their studies and this one reveal that in the context of anomaly coupled dynamics, the timescale of ocean basin modes increases as the coupling coefficient increases leading to unstable ocean basin modes with an infinite oscillatory timescale. In the framework of a fully coupled model, we further demonstrate that this air–sea interaction instability leads to a west–east asymmetric solution. And the (damped) ocean basin mode can then travel along the sloped thermocline with a timescale much longer than the uncoupled one. As the coupling coefficient increases further, the west–east asymmetry in the mean state amplifies whereas the oscillation timescale starts to decrease from its peak value at the instability point.
toward an asymptotic limit equal to the uncoupled timescale. When the amplitude of the self-generated west-east asymmetric mean state is small, the oscillation about the coupled climate state is still a damped oscillation. But just slightly increasing the coupling strength, which results in a slightly stronger west-east asymmetry of the self-generated climate state, leads to a self-sustained finite-amplitude oscillation.

Neelin and Dijkstra (1995) also found the nonlinear west-east asymmetric equilibrium solution in the absence of a preexisting easterly when the coupling strength is sufficiently and “unrealistically” large in their model. Moreover, the spatial pattern of the nonlinear equilibrium solution is not quite realistic, showing a deep “cold” SST sandwiched between the warm region in the central and eastern part of the basin. Nevertheless, their results are in a good agreement with the results reported here as far as the “symmetry breaking” phenomenon due to air–sea coupling in the absence of a preexisting asymmetry factor is concerned. Neelin and Dijkstra (1995) solved their SST equation and atmospheric model continuously along the equatorial basin whereas we only solve for the SST at the two boundaries and the wind stress at the central basin with a prescribed basinwide zonal profile. In reference to our results, perhaps their preexisting external easterly wind acts more to “seed” a basinwide SST profile so that the coupled nonlinear solution would take place at a realistic coupling strength and its spatial pattern would look realistic. In our case, we have effectively specified a basinwide scale for the SST. By doing so, there is essentially no difference between the solution with/without a preexisting external easterly wind (Fig. 1 versus Fig. A1 to be presented in the appendix).

Currently, a modeling and diagnostic study with a state-of-the-art coupled ocean–atmosphere general circulation model is planned to ratify the results obtained with the simple coupled system. In principle, this theory is also applicable to the equatorial Atlantic basin where a strong west–east asymmetry in the mean state is also observed. We have made calculations with the coupled model for the equatorial Atlantic basin. The results confirm that the same “softening” mechanism of the oscillatory inertia of the equatorial ocean basin due to basinwide air–sea coupling can give rise to a realistic west–east asymmetric mean state over the equatorial Atlantic basin and an oscillatory variability with a timescale about 1–2 yr. Details of the results will be summarized in a separate paper. Obviously, this simple theory would not be applicable to the equatorial Indian basin. Because of the presence of the land–ocean contrast is just several latitudes north of the equator and the dominance of the warm pool is over the western boundary of the equatorial Pacific, it is very unlikely the air–sea interaction within the equatorial Indian basin alone is responsible for the mean climate state there.

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APPENDIX

Sensitivity of Solution to the Key Model Parameter Settings

We can add a constant westward wind stress in Eq. (3) to examine the effect of a preexisting time-independent easterly wind. Figure A1 summarizes the numerical solution of (1)–(4) with a small preexisting time-independent westward wind stress (−0.005 dyn cm⁻²). In the absence of air–sea interaction (α = 0), the Sverdrup balance with this external wind stress results in a basinwide depth difference equal to 1.1 m and the corresponding Ekman upwelling creates an SST difference of 0.03°C across the basin. This small amount of basinwide SST difference corresponds to the “seed” referred to in Dijkstra and Neelin (1995) and Neelin and Dijkstra (1995). The air–sea interaction only marginally strengthens the zonally asymmetric climate state until the coupling coefficient reaches α₀ (=0.58). For α > α₀, the much more pronounced west–east asymmetry in the time mean state is clearly a result of air–sea interaction instability. As α further increases, the zonally asymmetric time mean state amplifies vigorously. The timescale of the coupled oceanic waves first increases

Fig. A1. As in Fig. 1 except that a constant wind stress equal to −0.005 dyn cm⁻² is added in (3). The curve with solid circles is the time mean of (hₘₚ – hₘₜ)/4 in meters.
with \( \alpha \) and reaches the maximum value (equal to 5.6 yr) at \( \alpha = \alpha_0 \). The oscillation period then decreases as \( \alpha \) increases. Comparing Figs. 1 and A1 reveals that the oscillation period shortens as a result of the presence of this weak external westward wind stress. Moreover, the self-sustained oscillation about the coupled climate state exists at a slightly larger value of \( \alpha (\alpha = \alpha_1 = 0.63) \). The numerical solutions with different values of pre-existing easterly wind reveal that the critical values of \( \alpha_n \) and \( \alpha_1 \) tend to increase as the preexisting wind increases. In summary, we conclude that the solution with an external easterly wind behaves essentially the same as what we have shown in Fig. 1.

We have performed a series of numerical integrations to examine the robustness of the coupled solution to the model parameters. The results are summarized as follows.

1) Parameter \( H^* \) in Eq. (4): For fixed values \( T_0 (=30^\circ C) \) and \( T_{90} (=25^\circ C) \), this parameter is the sole parameter determining the effectiveness of the parameterized thermocline feedback. A very large value of \( H^* \) would effectively shut off the thermocline feedback for a given coupling coefficient, while a smaller \( H^* \) enhances the thermocline feedback and effectively increases the coupling strength. The numerical calculations suggest the change in \( H^* \) does not modify the general behaviors of the coupled system other than altering the critical value of the coupling coefficient \( \alpha \) for air–sea interaction instability. Not surprisingly, a larger (smaller) \( H^* \) yields a large (smaller) value of the critical coupling strength.

2) Parameter \( \delta \) in Eq. (4): This parameter determines the strength of the Ekman feedback (or the relation between wind stress and the Ekman upwelling rate in the mixing layer). A very small value of \( \delta \) would disable air–sea interaction. The standard numerical setting for this parameter given in Table 1 implies that for a mixing layer 50-m deep, a westward wind stress of 1 dyn cm \(^{-2} \) would create an equatorial Ekman upwelling of 0.006 cm s \(^{-1} \). This setting is quite comparable with the relation between the observed mean wind stress and upwelling at the equator. The sensitivity experiments show that a \( \pm 100\% \) change in the numerical value of this parameter produces little change to the solution of the coupled system.

3) Newtonian cooling rate \( \mu \) in Eq. (4): Unlike the two previous parameters, this parameter has no effect on either air–sea interaction instability or the existence of the coupled climate state (of course, it cannot be zero for an obvious reason). But a very large value of \( \mu \) can suppress the self-sustained oscillation because it effectively yields a relatively weaker Ekman feedback. It follows that the self-sustained oscillation can revive by just increasing \( \delta \) slightly for a larger value of \( \mu \). For example, for \( \mu = 1/(12 \text{ months}) \), that \( \delta = 12.5/(3 \text{ months}) \) (dyn cm \(^{-2} \)) \(^{-1} \), a 25% increase from the standard value of \( \delta \), would be sufficient for the model to have a self-sustained oscillation solution again. Other than this, the solution of (1)–(4) shows little sensitivity to this parameter.

4) Meridional width of the coupled wind stress \( Y \) in (2): The coupled climate state only marginally depends on the numerical value of this parameter. Broadening of the meridional width of the coupled wind stress (say \( Y > 10^8 \)), however, can suppress the self-sustained oscillation even through it would also increase the timescale of oceanic waves by imparting more Rossby wave energy to higher meridional modes. Narrowing the meridional scale of the coupled wind stress has no effect on the self-sustained oscillation other than reducing the oscillation period slightly.

Recall that the parameter \( A_y/L \) is not an independent model parameter. A different value (as long as it is not zero) merely changes the numerical scale of the abscissa in Figs. 1 or A1. Other model parameters, including \( T_0 \) and \( T_{90} \) in Eq. (4), are directly measurable from observation. The solution of the coupled system is not sensitive to their numerical values as long as they are set in accordance with the observation. In addition, we note that there is no damping term in the ocean dynamic equation (1). Inclusion of a small damping term would not change the equilibrium solution significantly. But the oscillation amplitude is sensitive to the presence of damped oscillation although the oscillation timescale remains nearly unchanged. For example, we found that for a damping rate larger than \( 1/(6 \text{ yr}) \), the self-sustained oscillation becomes a damped oscillation for the standard setting of the parameters listed in Table 1. The self-sustained oscillation solution can be revitalized in the relatively strong damping case by either increasing the parameter \( \delta \) or decreasing the relaxation timescale toward the radiative equilibrium sea surface temperature \( T_0 \).

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