Oceanic Rossby Wave Dynamics and the ENSO Period in a Coupled Model

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ABSTRACT

Tropical ocean wave dynamics associated with the El Niño–Southern Oscillation cycle in a coupled model are examined. The ocean–atmosphere model consists of statistical atmosphere coupled to a simple reduced gravity model of the tropical Pacific Ocean. The statistical atmosphere is simple enough to allow for the structure and position of the wind stress anomalies to be externally specified. In a control simulation, where the structure of the wind stress anomaly is determined from observations, the model produces a regular 5-yr oscillation. This simulation is consistent with the so-called delayed oscillator theory in that subsurface wave dynamics determine the slow timescale of the oscillation and surface-layer processes are found to be of secondary importance. Kelvin and Rossby wave propagation is detected along the equator, with periods considerably shorter than the simulated oscillation period. The way in which these relatively fast waves are related to the simulated 5-yr oscillation is discussed.

In order to understand the mechanism responsible for the 5-yr period in the control simulation, two sets of sensitivity experiments were conducted. The first set of experiments focused on how the meridional structure of the wind stress anomaly influences the model ENSO period. Relatively broad (narrow) meridional structures lead to relatively long (short) periods. While the gravest Rossby wave appears to be important in these simulations, it is found that the maximum variability in the thermocline is associated with off-equatorial Rossby waves (i.e., Rossby waves that have a maximum amplitude beyond ±7° of the equator). The second set of sensitivity experiments was designed to examine how these off-equatorial Rossby waves influence the ENSO cycle. Without the effects of the off-equatorial Rossby waves at the western boundary, the model produces a 2-yr oscillation regardless of the meridional structure of the wind stress anomaly. The mechanism by which these off-equatorial Rossby waves influence the ENSO period is described. Based on these experiments, it is shown that the reflection of the gravest Rossby wave off the western boundary is required to produce oscillatory behavior in the model, but the period of the oscillation is determined by the off-equatorial Rossby waves and the latitude at which they are forced.

1. Introduction

On interannual timescales, the variability in the tropical Pacific is dominated by the El Niño–Southern Oscillation (ENSO) phenomenon (Rasmusson and Carpenter 1982; among others). During the warm phase of this oscillation, there is a dynamic adjustment of heat and mass between the western and eastern tropical Pacific, producing a positive sea surface temperature anomaly (SSTA) in the eastern Pacific. Associated with this dynamic adjustment, precipitation is displaced eastward from the climatological warm pool region toward the date line, and the normally easterly trade winds weaken or even become westerly. During the cold phase, the eastern tropical Pacific SSTA is negative, the trade winds strengthen, and the precipitation is tightly confined to the warm pool region of the western Pacific. Typically, the time between warm events is around 3 to 5 yr; however, there is also considerable modulation of the ENSO cycle on decadal timescales (Wang 1995; Zhang et al. 1997).

Bjerknes (1966, 1969) was the first to postulate that large anomalies in the tropical Pacific are due to coupled ocean–atmosphere interactions. The coupled instability mechanism described by Bjerknes (1969) works as follows. If positive eastern equatorial SSTA exists, the temperature gradient between the eastern Pacific and the western Pacific is reduced, which then produces a weakening of the easterly trade winds, augmenting the warming in the eastern Pacific. The additional warming in the east further weakens the trade winds, leading to a coupled ocean–atmosphere instability. A reversal of the argument explains the growth of a cold event. This positive feedback between the SSTA and the atmospheric wind anomaly leads to a continually growing anomaly, although it does not provide an explanation for what causes the transition from one extreme state to the other.

Two competing theories have emerged to describe the transition between warm and cold ENSO states. Schopf
and Suarez (1988), Suarez and Schopf (1988), and Battisti and Hirst (1989) have suggested the “delayed oscillator” mechanism to explain the slow timescale of the quasi-regular oscillation between warm and cold ENSO states. According to the original delayed oscillator theory of Schopf and Suarez (1988) and Battisti and Hirst (1989), a westerly zonal wind stress anomaly in the central Pacific excites a downwelling equatorial Kelvin wave that propagates rapidly to the east and a somewhat slower Rossby wave that propagates to the west. As the Kelvin wave propagates eastward into a region where the thermocline is shallow, the downwelling thermocline anomaly produces a positive SSTA, giving rise to local air–sea coupling. In this coupling region of the eastern Pacific, the disturbance grows as a coupled local instability and any migration develops into a standing, growing, atmosphere–ocean instability. This local growth is due to the Bjerknes instability (see also, e.g., Philander et al. 1984). The original Rossby wave propagates westward from the forcing region in the central Pacific into a region in which the thermocline is relatively deep, so that the thermocline anomaly has little effect on the SSTA and local air–sea coupling. Once the Rossby wave hits the western boundary, it reflects, forming an upwelling Kelvin signal that ultimately reaches the coupled disturbance in the east, negatively interfering with the downwelling and leading to the termination of the warm event and the initiation of the cold phase.

In the above scenario, if the dominant westward propagating signal is carried by a first meridional mode Rossby wave (the gravest mode), then twice the transit time for the Kelvin and Rossby waves is substantially shorter than the observed ENSO period. Cane et al. (1990), using a linear delayed oscillator equation, argued that the oscillation period is longer than twice the delay because the signal in the east is growing due to local instability and, as a consequence of this local growth, the upwelling Kelvin wave that is the result of the western boundary reflections does not completely eliminate the initial downwelling. Additional Rossby wave propagation into the western boundary and the associated reflected Kelvin waves are required to reverse the oscillation. The competition between local growth in the eastern Pacific and the efficiency of western boundary reflections has also been discussed by Battisti and Hirst (1989) and Schopf and Suarez (1990). Battisti and Sarachik (1995) contains a summary of the delayed oscillator theories and a discussion of how well the theories explain the observed characteristics of ENSO.

Neelin (1991), Neelin and Jin (1993), Jin and Neelin (1993a, b), and Hao et al. (1993) suggest an alternative mechanism for the slow ENSO timescales, where ocean wave dynamics play a minimal role and oceanic surface-layer processes provide the “memory” of the coupled system. Using a modified version of the Zebiak and Cane (1987) coupled model, Neelin (1991) identified an unstable and oscillatory “slow SST–fast wave” mode, which contrasts with the “fast SST–slow wave” mode of the delayed oscillator theory. In the slow SST–fast wave limit, the adjustment timescale of the Kelvin and Rossby waves is short compared to the adjustment time of the SST. The slow adjustment time of the SST is a result of local air–sea coupling and advective processes. In this slow SST–fast wave limit, the thermocline anomaly is in equilibrium with the wind stress anomaly and the slow timescale is determined at the surface. On the other hand, in the fast SST–slow wave limit, the SSTA is in equilibrium with the thermocline anomaly and the slow timescale is determined by Kelvin and Rossby wave dynamics in the subsurface. While these two theories suggest contradictory physical mechanisms, Neelin and Jin (1993) have found that they can both be obtained using the same model with different values of the tunable parameters.

In order to determine which mechanism is operating in the real ocean, Schneider et al. (1995, hereafter referred to as SHS) compared two uncoupled ocean general circulation model (OGCM) simulations with prescribed wind stress forcing. First, a control integration was carried out, in which the ocean is forced by observed wind stresses. In the second integration, the same wind stress forcing was used, except that it was applied backward in time. By comparing these two integrations at the same “forcing time,” SHS showed that the heat content was not in equilibrium with the forcing and that time lags due to heat storage in the mixed layer were unimportant for the SST. The primary imbalance or delay between the wind stress anomaly and the ocean fields was in the zonal mean heat content anomaly across the basin.

The results of SHS are consistent with the delayed oscillator theory and the fast SST–slow wave limit; however, SHS derived an empirical delayed oscillator equation based on their OGCM simulations and found that time delays due to equatorially trapped Kelvin and Rossby waves alone could not explain the low-frequency ENSO timescale. In the SHS empirical delayed oscillator equation, the characteristic equation for the complex frequency depends on the amplitude of the thermocline anomaly at the western boundary. Once the amplitude of the thermocline anomaly at the western boundary is permitted to be frequency dependent, then low-frequency ENSO oscillations can be found. SHS used an interference argument as the off-equatorial Rossby waves propagate from the central Pacific, where they are forced by the curl of the wind stress to the western boundary to explain this frequency dependence. This interference argument was based on the zonal extent of the wind stress anomaly. McCreary (1983) also argued for the importance of these off-equatorial Rossby waves in determining the ENSO period, but his formulation for the atmospheric wind stress anomaly was unrealistic, and Battisti (1989) found that off-equatorial Rossby waves in his simulations were not always a precursor to ENSO events.
The mechanism for the slow timescale in the SHS delayed oscillator is different from that of the earlier delayed oscillator theory (Battisti and Hirst 1989; Schopf and Suarez 1988; Suarez and Schopf 1988; Schopf and Suarez 1990; Cane et al. 1990). According to the earlier theory, the ENSO period is determined by a competition between the time delay established by the ocean wave dynamics and the growth of the local instability in the east. While the excitation of higher meridional mode Rossby waves can lead to longer delays and longer periods, the fundamental process of the delayed oscillator involves the graviest Rossby mode. In the SHS empirical delayed oscillator, there is no explicit local instability in the east and the period is influenced by off-equatorial Rossby waves as they strike the western boundary. In this paper, we reexamine the delayed oscillator mechanism by making a distinction between the equatorially trapped Rossby waves and the off-equatorial Rossby waves forced by the curl of the wind stress. Strictly speaking, all the Rossby waves discussed here are equatorially trapped in that the propagation is completely zonal. However, we refer to Rossby waves that are forced by wind stress curl beyond \( \pm 7^\circ \) of the equator as equatorial or subtropical Rossby waves. Rossby waves forced within \( \pm 7^\circ \) of the equator are referred to as equatorially trapped.

A coupled model is presented in which the structure and location of the wind stress anomaly are externally controlled, but the amplitude of the anomaly is determined by the SST in the eastern Pacific. The ocean component of this coupled model is identical to the ocean component used in the Zebiak and Cane (1987) coupled model. This coupled model includes the effects of local instability in the east, equatorially trapped Rossby and Kelvin waves, and off-equatorial Rossby waves. The ocean model SST variations include the effects of horizontal and vertical temperature advection and Newtonian cooling. In a control simulation, where the spatial structure of the wind stress anomaly is determined directly from the ocean model SSTA, the ENSO state is consistent with the fast SST–slow wave limit in that wave dynamics in the subsurface determine the oscillation period. There are short time lags (approximately 1 month) between the SST and the thermocline depth anomaly; however, these time lags do not explain the oscillation period in the model. The transition from one extreme ENSO state to another is consistent with the delayed oscillator theories in that equatorially trapped Rossby wave reflections off the western boundary are required for the model to change ENSO states.

The question of primary interest here is why the coupled model selects a 5-yr oscillation period. Two sets of sensitivity experiments presented here demonstrate that this relatively long timescale is due to the influence of off-equatorial Rossby waves. In the first set of sensitivity experiments, the meridional structure of the wind stress anomaly is made both broader and narrower. Consistent with Cane et al. (1990) and Schopf and Suarez (1990), relatively broad (narrow) structures produce relatively long (short) periods. In the second set of sensitivity experiments, the effects of equatorially trapped Rossby waves at the western boundary have been filtered out. In this case, a 2-yr period results, regardless of the meridional breadth of the wind stress anomaly. Finally, the mechanism by which the off-equatorial Rossby waves lengthen the ENSO period in the coupled model is described.

2. Experimental design

In these experiments, we use a coupled ocean–atmosphere model that consists of a very simple statistical atmosphere coupled to the ocean model used in the Zebiak and Cane (1987) coupled model. The details of the ocean model and the parameter values used here are discussed in Zebiak and Cane (1987). The dynamics of the ocean model are described by the linear reduced gravity equations, which simulate thermocline depth anomalies and depth-averaged baroclinic currents. A shallow frictional layer of constant depth (50 m) is added to the top of the model to simulate the surface intensification of wind-driven currents. The equation governing the SST includes the effects of horizontal and vertical advection of temperature by both mean and anomalous currents. The annual cycle is included in the model by the prescribed mean current, temperature, and thermocline depth. The SST equation also includes an anomalous surface heat flux that acts to damp the SST to zero.

There are several ways of designing a statistical atmosphere for coupled ENSO simulations with various degrees of sophistication (Barnett et al. 1993; Balness et al. 1994; among others). In general, these statistical atmospheres attempt to capture as much of the observed seasonal to interannual wind stress variability as possible. However, since our focus is on how the structure of the wind stress anomalies affect the ENSO period, the statistical atmosphere used here is simple enough so that the magnitude of the wind stress anomaly is determined directly from the ocean model SST, while the structure of the wind stress anomaly is externally prescribed. To some degree, the assumption of a spatially fixed structure is supported by observations of the zonal wind stress anomaly along the equator, which is largely dominated by a standing oscillation.

With the above constraint in mind, the atmosphere is modeled as

\[
\tau_x = \alpha(x, y)(\text{NINO3}) \quad \text{and} \quad \tau_y = \beta(x, y)(\text{NINO3}),
\]

where NINO3 is the SST in the coupled model averaged over the so-called NINO3 region (\(150^\circ\text{W} - 90^\circ\text{W}, 5^\circ\text{S} - 5^\circ\text{N}\)), and \(\tau_x\) and \(\tau_y\) are the zonal and meridional
wind stress anomalies, respectively. The structure functions $a$ and $b$ are independent of time and are externally prescribed in the coupled model simulations.

For the control simulation, the structure functions $a$ and $b$ are determined by linearly regressing the observed time series of NINO3 SSTA from 1964–94 onto The Florida State University (FSU) analyzed pseudostress (Goldenberg and O’Brien 1981). As with the Zebiak and Cane (1987) ocean model, the FSU pseudostress is converted into a wind stress by assuming a constant drag coefficient of $2 \times 10^{-2}$ and a constant air density of $1 \text{ kg m}^{-3}$. The argument for the “superdrag” coefficient in the empirical model is presented in Barnett (1984).

The control structure of the zonal and meridional wind stress anomalies determined from the linear regression is shown in Figs. 1a and 1b. In plotting the wind stress in Figs. 1a and 1b, we have assumed a NINO3 anomaly of $1.35^{\circ}C$. For comparison, Figs. 2a and 2b show a composite warm event zonal and meridional wind stress anomaly calculated from the FSU data. The composites are calculated by taking the average of the mean anomaly from December–February (DJF) of 1965–66, 1972–73, 1976–77, 1982–83, 1986–87, and 1991–92.

In terms of the general structural features, the control (or regressed) wind stress anomalies are similar to the observed composites. There is a large-scale westerly zonal wind stress anomaly in the western and central equatorial Pacific. In the observed composite, the center of the anomaly is south of the equator and is somewhat stronger, and there are easterly anomalies to the north and south that do not appear in the control zonal wind stress anomaly. The broad features of the meridional wind stress anomaly are also similar, but the minimum meridional stress along $5^{\circ}N$ in the control is shifted to the east by about $20^{\circ}$ longitude compared to the observed composite.

The primary advantage of this statistical atmosphere is its simplicity. Multicentury integrations are easily made, and the sensitivity to various parameters may be explored. This simplicity is also the primary disadvantage of the model. While the observed wind stress anomaly along the equator has a large standing component, the spatial structure of the anomaly is not time independent. Propagating components in the wind stress anomaly may also be important, as observed during the 1982–83 ENSO event. Nevertheless, the control simulation shown in the next section captures some of the large-scale features of the observed ENSO cycle.

The sensitivity experiments are formulated by modifying the structure of $a$ and $b$ in the control wind stress anomaly. In the first set of sensitivity experiments, the meridional extent of the anomaly is made broader or narrower by modifying $a$ and $b$. The procedure for modifying the structure of the wind stress anomaly is described in the appendix. The intent of these sensitivity experiments is to assess how and if the structure of the wind stress anomaly in the central Pacific affects off-equatorial and equatorially trapped Rossby wave propagation and the corresponding ENSO period in this coupled model. It should be noted that in modifying $a$ and $b$, the forcing of both the equatorially trapped and the off-equatorial waves has been changed. However, in an additional sensitivity experiment, discussed in section

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1 Using the same composite procedure, the observed composite NINO3 temperature anomaly is $1.35^{\circ}C$. 

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Fig. 1. The control (a) zonal wind stress anomaly and (b) meridional wind stress anomaly. A NINO3 temperature anomaly of $1.5^{\circ}C$ is assumed, and the contour interval is 0.1 dynes cm$^{-2}$.

Fig. 2. The FSU observed (a) zonal wind stress anomaly and (b) meridional wind stress anomaly composite. The composite is formed by taking the average anomaly for DJF of 1965–66, 1972–73, 1976–77, 1982–83, 1986–87, and 1991–92. The contour interval is 0.1 dynes cm$^{-2}$. 

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4, it was found that the dominant effect was due to changes in the forcing of off-equatorial Rossby waves.

Figures 3a and 3b show the meridional structure along the date line of the zonal wind stress anomaly and the curl of the wind stress anomaly for the control and the two sensitivity experiments. As in Fig. 1a, we have assumed a NINO3 SSTA anomaly of 1.35°C. We have omitted showing the meridional wind stress anomaly because, through a number of different experiments, we have found that it has little effect on the dominant ENSO period in this model. It is clear from Fig. 3a that as the zonal wind stress anomaly becomes narrower, the amplitude along the equator becomes larger. Again, this changes the forcing of both the equatorially trapped and off-equatorial Rossby waves. Nonetheless, in the experiments presented here, the most important difference is that the latitude of maximum curl shifts to higher latitudes with a broader meridional structure (Fig. 3b). Throughout the remainder of the paper, the sensitivity experiments with these two modified meridional structures will be referred to as the narrow and the broad cases, respectively.

The intent of the second set of sensitivity experiments is to determine the relative roles of the off-equatorial and the equatorially trapped Rossby waves in the simulated ENSO cycle. This is accomplished by filtering out the effects of the off-equatorial Rossby waves at the western boundary, while leaving the structure of the wind stress anomaly unmodified. The filtering procedure and the results of the filtered experiments are described in section 4.

We have performed a large number of different experiments with idealized structures for a and b. We have also examined the sensitivity to multiplying a and b by amplification (or coupling strength) factors. The subset of sensitivity experiments presented here accurately characterizes this larger set of experiments.

3. Control simulation

All of the simulations shown here were integrated for 200 simulated yr, and the results were analyzed for the last 100 yr. The initial conditions for the simulations were identical and were taken from an extended integration of the ocean model forced with observed FSU wind stress.

ENSO variability

The SSTA variability for the observed data and the control simulation is shown in Figs. 4a and 4b. Figures
Fig. 5. The shading shows time–longitude cross sections along the equator of (a) the control thermocline anomaly and (b) the control thermocline anomaly minus the zonal mean anomaly across the basin. (c) The thermocline anomaly due to all the Rossby wave components. (d) The thermocline anomaly due to the Kelvin wave component. In all four panels, the zonal wind stress anomaly along the equator is superimposed with a contour interval of 0.1 dynes cm$^{-2}$. The units for the thermocline anomaly are in m.

4a and 4b show time–longitude cross sections of the SSTA along the equator, where the observed field is plotted for 1982–92 and the control field is plotted for simulation years 188–198. The control simulation captures some of the basic features of the observed ENSO characteristics. The SSTA is strongest in the eastern Pacific, with little or no anomaly to the west of the date line. The control ENSO events reach their peak amplitude during the boreal winter season, much like the observed signal. The control SSTA is primarily a standing oscillation. Although it is difficult to detect in Fig. 4b, there is a rapid eastward migration of the SSTA during the boreal winter and spring. The observed SSTA is also primarily a standing oscillation, except in 1982–83, when there was an eastward migration of the SSTA. While the durations of the control warm events are too long compared with the observed data, the asymmetry between warm and cold events is similar to the observed ENSO cycle. This asymmetry can be seen in the transition from one extreme ENSO state to another. In both the control and the observed data, the transition from the warm phase to the cold phase is more rapid than the transition from the cold phase to the warm phase.

Similar to SHS, we find that, in the control simulation, the SSTA and the zonal wind stress anomaly along the equator are largely in equilibrium with each other. There is approximately a 1-month phase lag between the thermocline anomaly and the SSTA, but this phase difference does not appear to be important in terms of determining the dominant ENSO period in the model. Ocean wave dynamics along the thermocline provide the mechanism for the model to oscillate from one extreme ENSO state to another and for determining the period of the oscillation.

Using the same format as Figs. 4a and 4b, Figs. 5a and 5b show the zonal wind stress anomaly along the equator as functions of time. In both panels, the zonal wind stress variations along the equator are superimposed on the thermocline variations. In plotting Figs. 5a and 5b, we have decomposed the thermocline anomaly into $H = H^* + \bar{H}$, where $\bar{H}$ is the zonal mean thermocline anomaly across the
basin and $H^*$ is the variation about the zonal mean. In Fig. 5a, the shaded contours indicate the total thermocline anomalies ($H$), and in Fig. 5b, the shaded contours indicate the thermocline anomalies once the zonal mean thermocline across the basin is removed ($H^*$). The control $H$ is dominated by a zonal dipole, which changes sign with the ENSO cycle. The superposition of the zonal wind stress anomaly highlights the fact that $H$ is not in equilibrium with the zonal wind stress anomaly. This imbalance appears as a phase lag between $H$ and the zonal wind stress anomaly, which can be easily seen in the central Pacific (Fig. 5a), with positive $H$ leading the westerly wind stress anomaly. The phase difference between $H$ and the zonal wind stress anomaly disappears once the zonal mean ($\bar{H}$) is removed, which indicates that $H^*$ (Fig. 5b) is largely in equilibrium with the zonal wind stress anomaly.

This equilibrium condition between $H^*$ and the zonal wind stress anomaly results from the approximate balance in the zonal momentum equation along the equator (Cane 1992). Put simply, the anomalous thermocline gradient balances the zonal wind stress anomaly along the equator. The past history of the wind stress anomaly or the memory of the coupled system is detectable in variations of $\bar{H}$. The SSTA is in equilibrium with $H^*$ (but not $H$), which gives standing oscillations at the surface. This result is consistent with the results of Wyrtki (1985), Zebiak and Cane (1987), Philander and Hurlin (1988), Zebiak (1989), and SHS.

During any simulated ENSO event and outside of the forcing region in the central Pacific, a continuous stream of wave pulses or packets can be identified in Figs. 5a and 5b. Compared to the timescale of the simulated ENSO events, these wave packets propagate away from the forcing region relatively quickly. Given the large degree of equilibrium between $H^*$ and the zonal wind stress anomaly, the fast propagation speeds of these wave pulses, and the fact that the slow timescale memory appears in $\bar{H}$ variations, it is easy to see how one might conclude that ocean wave dynamics are unimportant in determining the ENSO period. Nevertheless, we find that ocean wave dynamics and the associated time delays provide the mechanism for the transitions between extreme ENSO states.

Throughout the remainder of this paper, wave packets similar to those identified in Figs. 5a and 5b will be identified as Kelvin and Rossby waves. Strictly speaking, these wave pulses are not individual Kelvin and Rossby waves but consist of continuously forced packets of Kelvin and Rossby signals. When we refer to the gravest Rossby wave, we are referring to a packet of Rossby waves whose phase speed is dominated by the gravest mode. Similarly, when we refer to off-equatorial Rossby waves, we mean a packet of off-equatorial Rossby waves forced at a particular latitude by the curl of the wind stress anomaly. These off-equatorial Rossby wave packets have decreasing phase speed with increasing distance from the equator (e.g., McCreary 1983).

While relatively fast equatorial Rossby and Kelvin wave propagation along the equator can be detected in Figs. 5a and 5b, in order to better diagnose the wave propagation and reflection off the western and eastern boundaries, it is necessary to separate $H$ into the different Rossby and Kelvin wave components. Figures 5c and 5d show time–longitude cross sections of $H$ variations due to all of the Rossby wave components and the Kelvin wave component, respectively. The Kelvin wave component (Fig. 5d) is calculated by projecting $\bar{H}$ onto the appropriate meridional structure. The component due to all of the Rossby waves (Fig. 5c) is calculated by subtracting the Kelvin wave component from $\bar{H}$. Rossby wave propagation off the eastern boundary as the result of reflecting Kelvin waves is readily apparent. At the western boundary, each incoming Rossby wave reflects as a Kelvin wave, and even though the Rossby component includes all of the meridional modes, the phase propagation along the equator is dominated by the gravest mode. There appears to be destructive interference as the reflected waves reenter the forcing region in the central Pacific. It is also apparent from Fig. 5d that the meridional structure of $\bar{H}$ primarily projects onto the meridional structure of the Kelvin wave.

Eastern boundary reflections also appear to be important in Fig. 5c, but, as shown in section 4, the variability associated with these reflected Rossby waves is considerably smaller than the Rossby waves generated by the wind stress anomaly in the central Pacific.

The evolution and alternation between the warm and cold events along the equator in the control simulation appear to be consistent with the delayed oscillator theories. For example, a westerly wind stress anomaly excites a relatively strong downwelling Kelvin wave that propagates away from the forcing region into the eastern boundary, where it reflects as Rossby waves. As this Kelvin wave propagates through the eastern Pacific, the SSTA increases due to local ocean–atmosphere coupling (see Fig. 4b). Just to the west of the westerly wind stress anomaly, a gravest meridional mode Rossby wave propagates into the western boundary and reflects as an upwelling Kelvin wave, which reduces as it propagates into the eastern Pacific but does not reverse the local downwelling anomaly. As the ENSO event develops, there is a continuous stream of these Kelvin and Rossby waves propagating away from the forcing region in the central Pacific. Eventually, the Kelvin waves reflected off the western boundary have sufficient amplitude to reverse the sign of the SSTA in the eastern Pacific.

The above explanation of the transition from the warm to the cold ENSO state is completely described by equatorially trapped modes. It also appears that the phase speed of the gravest Rossby wave determines when the Kelvin waves, which act to damp the anomaly in the east, are reflected off the western boundary. In the delayed oscillator theory, the amplitude of these Kelvin waves at the western boundary plays a critical role in determining the period of the oscillation. For
instance, in the transition from the warm to the cold state, if the amplitude of the upwelling Kelvin wave is reduced, then it will be less effective at reversing the downwelling wave in the equatorial region. If the amplitude at the western boundary is reduced enough, it will take additional Kelvin waves to reverse the oscillation, thereby increasing the period. In this model, the amplitude of the Kelvin wave at the western boundary is determined by all the Rossby waves striking the western boundary. Moreover, in some experiments, the periods of the oscillation are 2 yr, regardless of the latitude. It should be noted that this result is different from that obtained by Battisti (1988) in different experiments, in which the equatorial Rossby waves that filtered from the coupled model. In the Battisti (1988) experiment, there was a slight decrease in the period in which the equatorial waves were filtered from his coupled model simulation.

4. Sensitivity experiments

The sensitivity experiments presented in this section are designed to determine why the control simulation selects a 5-yr ENSO oscillation period. In the first set of experiments, it is shown that the period is sensitive to the meridional breadth of the wind stress anomaly. In the second set of sensitivity experiments, by removing the impact of equatorial Rossby waves, we find that the oscillation period is 2 yr, regardless of the meridional breadth of the wind stress anomaly. The implication is that time delays due to the equatorially trapped waves alone cannot explain the ENSO period in the model. It is also shown how equatorial Rossby waves reduce the amplitude of the Kelvin waves at the equatorial region, and it is found that this reduction in amplitude lengthens the period of the oscillation. Based on these experiments, it is argued that the period in the control simulation is strongly influenced by equatorial Rossby waves. In particular, because the phase speed of equatorial Rossby waves decreases with increasing distance from the equator, the latitude at which these waves are forced determines the ENSO period.

a. Narrow and broad experiments

The narrow and broad simulations were made in the same way as the control simulation. The meridional structure of the zonal wind stress anomalies for these experiments was shown in Fig. 3, and the procedure for calculating them is described in the appendix. There is a clear separation between the dominant ENSO period in all three simulations. Using the broader wind stress anomaly, the period is about 3 yr, in contrast to the 5-yr period in the control simulation.

Figures 6a–c show time–longitude cross sections of the thermocline anomaly and the wind stress anomaly along the equator for the broad, control, and narrow cases. While the periods are markedly different, there are characteristics of the simulated ENSO events that are uniform across all three experiments. For example, the variability in $H$ is dominated by a standing oscillation in all three simulations. There is a phase lag between $H$ and the zonal wind stress anomaly, which is most evident in the control simulation. Although not shown, $H^*(not H)$ is in equilibrium with the wind stress anomaly and the SSTA. In the forcing region, equatorially trapped Kelvin and Rossby waves are excited, which propagate to the east and west on timescales that are fast compared to the ENSO oscillation.

Even though the period changes among the different experiments, the time delays associated with the Kelvin and gravest Rossby waves do not change. The phase speeds of these waves are unaffected by the meridional structure of the wind stress anomaly. The question remains as to why the model oscillates with a longer (shorter) period in the broad (narrow) simulation. According to Schopf and Suarez (1990), for example, the broader meridional structure in the wind stress anomaly leads to longer periods because higher meridional mode Rossby waves (i.e., with slower phase speeds) would imply longer delays.

Another possible explanation for the changes in the period is that changes in the structure of the wind stress anomaly mostly affect the gravest Rossby mode, which ultimately changes the period. This possibility comes from the fact that in broadening and narrowing the structure of the wind stress anomaly, the forcing of the gravest Rossby mode has changed considerably. In order to separately consider the changes to the wind stress forcing of the equatorially trapped and the off-equatorial Rossby waves, we have performed an additional experiment in which the control wind stress structure is used between $\pm 7^\circ$ of the equator and the broad structure is used beyond $\pm 7^\circ$ of the equator. The intent is to increase the forcing of the off-equatorial Rossby waves, while leaving the forcing of the gravest Rossby mode unchanged. In this experiment, an 8-yr period results, compared to the 9-yr period in the broad and 5-yr period in the control simulations, respectively. This result shows that the off-equatorial Rossby waves have a significant impact on the ENSO period in this coupled model. Specifically, the ENSO period in the model is determined by the relative forcing of the equatorially trapped and the off-equatorial Rossby waves, and is sensitive to the latitude at which these off-equatorial Rossby waves are forced.

The dominant latitude at which the off-equatorial Rossby waves are forced in the model can be clearly seen in the meridional structure of the temporal variance shown in Figs. 7a and 7b. In Fig. 7a, the variance is averaged over the western and central Pacific ($120^\circ$–$180^\circ$), and in Fig. 7b, the average is taken over the
eastern Pacific (120°–80°W). In the eastern Pacific, the
temporal variance is small compared to that of the west-
ern and central Pacific, is symmetric about the equator,
and is largest at the equator. In the central and western
Pacific, the variance is larger and is asymmetric about
the equator, and the latitude of maximum variance shifts
corresponding to the structure of the wind stress anom-
aly. In the broad simulation, the maximum variance
occurs at 12°S, and in the control simulation, the var-
iance is broader and centered along 8°S. In the narrow
simulation, the variance is narrower and centered at 5°S.
Weaker maxima at somewhat lower latitudes can be
found in the Northern Hemisphere. The latitude of max-
imum variance corresponds with the latitude of maxi-
mum absolute wind stress curl seen in Fig. 3b.

The dominant variability in the thermocline is focused
on a fairly narrow band of off-equatorial latitudes for
each simulation. In going from the narrow simulation to
the broad simulation, this narrow band of latitudes
corresponds to decreasing Rossby wave phase speed
with increasing latitude. In the central and western Pa-
cific, the variance associated with the relatively fast
equatorially trapped Rossby wave propagation is small
compared to that associated with the lower-frequency
off-equatorial Rossby waves. The variability at the east-
ern boundary is quite small compared to that of the
central and western Pacific, and it is unlikely that the
Rossby waves reflecting off the eastern boundary are
important in determining the simulated ENSO period.

The fact that the off-equatorial Rossby waves have a
different phase speed than the equatorially trapped
waves and the relative unimportance of the Rossby
waves reflected off the eastern boundary can be seen in
Figs. 8a–c. Figures 8a–c show the thermocline depth
anomaly at the latitude of maximum variance and the
wind stress anomaly at the equator for the broad, con-
trol, and narrow simulations, respectively. Comparisons
with Figs. 6a–c (note that the contour interval in Figs.
8a–c is larger than the contour interval in Figs. 6a–c)
show that the off-equatorial westward propagation is
FIG. 7. Temporal variance as a function of latitude in (a) the western and central Pacific (120°E–180°) and (b) the eastern Pacific (120°–80°W). The broad simulation is shown in the continuous curve, the control simulation is shown in the long-dashed curve, and the short-dashed curve shows the narrow simulation.

slower than the westward propagation along the equator. The Rossby waves reflected off the eastern boundary are damped as they approach the forcing region in the central Pacific. Rossby waves of the opposite sign are forced in the central Pacific by the curl of the wind anomaly, and it is these waves that give the relatively large variance seen in Fig. 7a.

It is also interesting to consider how these simulations correspond to the linear delayed oscillator of Battisti and Hirst (1989) and the nonlinear delayed oscillator of Suarez and Schopf (1988). In the Battisti and Hirst (1989) model, the fundamental physics are linear interactions between the local instability and the delayed wave effects; nonlinearities play a secondary role in the oscillations. In the Suarez and Schopf (1988) model, the wave delay effects play a secondary role and the dominant balance is between linear and nonlinear local instability. Both the broad and the control simulations appear to be in the nonlinear regime of Suarez and Schopf (1988) because the time delays due to the equatorial modes appear to be of secondary importance in determining the oscillation period. Moreover, these two simulations appear to oscillate between warm and cold stable states, which is predicted by the nonlinear delayed oscillator (Battisti and Hirst 1989). In the narrow simulation, it is more difficult to determine whether the model is in a nonlinear or linear regime. However, the thermocline variability is more bimodal, suggesting that the model is in the nonlinear regime. Interestingly, the filtered experiments described below appear to be more consistent with the linear regime, where the time delay established by the equatorial modes is of primary importance in determining the oscillation period.

b. Influence of off-equatorial Rossby waves

In this model, the influence of off-equatorial Rossby waves striking the western boundary is felt in determining the amplitude of the Kelvin wave reflected off the western boundary. In turn, the amplitude of the reflected Kelvin waves is important for determining when the oscillation alternates. In the long-wave approximation (Cane and Sarachik 1977), the amplitude of the Kelvin wave at the western boundary (\( K \)) is given by

\[
K(Y_s, Y_n) = -2 \int_{Y_n}^{Y_s} u_R(x = \text{western boundary}, y) \, dy,
\]

where \( u_R(x = \text{western boundary}, y) \) is the zonal velocity due to all of the Rossby waves at the western boundary, and \( Y_s \) and \( Y_n \) denote the northern and southern boundaries. It is possible to explicitly filter out the influence of the off-equatorial Rossby waves by restricting the integration in the above equation to include only those latitudes in the oceanic equatorial waveguide. This method for filtering the effects of off-equatorial Rossby waves at the western boundary was also presented by Battisti (1989) and Graham and White (1991). We have repeated the broad, control, and narrow simulations; except in this case, the amplitude of the Kelvin wave at the western boundary is determined only by the Rossby waves between \( Y_s = 7^\circ S \) and \( Y_n = 7^\circ N \). These experiments will be referred to as filtered-broad, filtered-control, and filtered-narrow simulations, respectively. The intent is to remove the effects of the off-equatorial Rossby waves only at the western boundary, and, based on the variance seen in Figs. 7a and 7b, this seems to be a reasonable choice of latitudes for the filter.

Figures 9a–c show time–longitude cross sections of the thermocline anomaly at the equator for the filtered-broad, filtered-control, and filtered-narrow simulations. The dominant period in all three simulations (approximately 2 yr) is considerably shorter than the simulations that include all of the Rossby waves at the western boundary (Figs. 6a–c). While the filtered-narrow simulation shows slightly more irregularity than the filtered-broad and filtered-control simulations, all three simu-
lations oscillate on a timescale that is consistent with a single reflection of the gravest Rossby wave. During each phase of the ENSO cycle, there is only one Kelvin wave reflection off the western boundary before the oscillation changes sign. This is consistent with the interpretation of the delayed oscillator presented by SHS and Battisti and Hirst (1988), and indicates that, when using wind stress structure functions calculated from observations, time delays due to the equatorially trapped modes alone are not sufficient to explain the ENSO period in the model.

The impact of the off-equatorial Rossby waves at the western boundary can be seen in Figs. 10a–c. Figures 10a–c show the amplitude of the Kelvin waves at the western boundary with and without the effects of the off-equatorial Rossby waves. Both curves of each panel are from the filtered simulations, where the dashed curve gives the amplitude of the Kelvin wave at the western boundary when only the equatorially trapped modes are included. This is the Kelvin amplitude that is felt by the model in the filtered simulations. The continuous curve is calculated from the filtered integrations, except that all latitudes are used to determine the amplitude of the Kelvin wave at the western boundary. The timescale of the oscillations in the Kelvin wave amplitude at the western boundary agrees with the thermocline variations seen in Figs. 10a–c. In all three simulations, the amplitude at the western boundary is larger without the effects of the off-equatorial Rossby waves.

The reduction in the amplitude of the Kelvin wave at the western boundary is due to the slower phase speed of the off-equatorial Rossby waves compared to the equatorially trapped waves. For example, if a gravest Rossby wave and an off-equatorial Rossby wave at 8°S were excited in the central Pacific at the same time, the gravest Rossby wave would reach the western boundary approximately 18 months earlier than the off-equatorial Rossby wave. In this coupled model, the slower phase speed of the off-equatorial Rossby waves results in a phase lag between the thermocline anomaly at the equator and at off-equatorial latitudes. Therefore, according to (1), the amplitude of the Kelvin wave reflected off...
the western boundary is reduced when the impact of off-equatorial Rossby waves is included (i.e., in the unfiltered experiments). This reduced amplitude implies that the reflected Kelvin waves are less efficient at reversing the signal in the eastern Pacific, and a longer oscillation period results. This phase difference is highlighted in Figs. 11a and 11b, where the thermocline anomaly at the western boundary is plotted for the broad and narrow experiments. In Fig. 11a, the solid curve gives the thermocline anomaly at the equator and the dashed curve gives the amplitude at 12°S. Similarly, in Fig. 11b, the solid curve gives the anomaly at the equator and the dashed curve gives the anomaly at 5°S. The phase lag between the equator and the off-equatorial latitudes is clearly seen in both simulations. The phase difference in the broad simulation is considerably larger than that in the narrow simulation, which is consistent with more destructive interference in determining the amplitude of the reflected Kelvin waves and a longer oscillation period.

These filtered experiments argue that the change in the period is not due to merely increasing the delay by exciting higher meridional mode Rossby waves, as, for example, in the Schopf and Suarez (1990) model. If the time delay were the primary factor in modifying the period, an increase in the period would be expected when the filter is applied. The reason for this is that if the dominant westward propagating signal is carried by higher meridional mode Rossby waves, then the filter would act to damp the signal striking the western boundary and reduce the amplitude of the reflected Kelvin waves, thereby increasing the period. In contrast, it is argued here that the filter removes the impact of the off-equatorial Rossby waves at the western boundary, which, because of the phase speed difference between the equatorially trapped and the off-equatorial Rossby waves, increases the amplitude of the Kelvin waves reflected off the western boundary and shortens the oscillation period.

Finally, we have also performed an experiment in which only the effects of the off-equatorial Rossby waves are used to determine the amplitude of the Kelvin waves at the western boundary. In this case, the model does not oscillate, regardless of the coupling strength. There are no Kelvin waves reflected off the western boundary without the influence of the gravest Rossby waves, and therefore there is no mechanism for the model to oscillate.
FIG. 10. Time series of the amplitude of the Kelvin wave at the western boundary for the (a) filtered-broad, (b) filtered-control, and (c) filtered-narrow simulations. The dashed contour indicates the amplitude when only latitudes between 5°N and 5°S are used, and the solid contour indicates the amplitude when all latitudes are used.

FIG. 11. Temporal variability of the thermocline anomaly at the western boundary in the (a) broad simulation at the equator (solid curve) and at 12°S (dashed curve). In the (b) narrow simulation, the solid curve gives the anomaly at the equator and the dashed curve gives the anomaly at 5°S.

5. Discussion and concluding remarks

Since the pioneering work of Schopf and Suarez (1988) and Battisti and Hirst (1989), there has been substantial debate regarding the delayed oscillator mechanism as an explanation for the ENSO cycle. The central confusion regarding the theory is that the gravest Rossby and Kelvin wave propagation along the equator is too fast to explain the observed ENSO timescale. Cane et al. (1990) clarified some of this confusion by showing that, due to local instability in the east, multiple Rossby wave reflections off the western boundary were required to reverse the oscillation. While changing the structure of the wind stress anomaly can lead to changes in the oscillation period, Schopf and Suarez (1990) argued that the gravest equatorially trapped Rossby wave was the primary agent producing the oscillation. SHS derived an empirical delayed oscillator equation without any Bjerknes instability in the east, where, in order to get the slow ENSO period, the amplitude of the thermocline displacements at the western boundary needed to be frequency dependent. SHS concluded that time delays due to the equatorially trapped modes alone could not produce the observed timescale. It was proposed that Rossby waves propagating into the western boundary forced off the equator by the curl of the wind stress were responsible for this frequency dependence and the corresponding slow ENSO period.

The motivation for this paper was to test the delayed oscillator hypothesis, with the intent of resolving the role of off-equatorial Rossby waves in modulating the ENSO period in the model. In this study, we referred to Rossby waves forced by wind stress curl beyond ±7° of the equator as off-equatorial Rossby waves. The Rossby waves that are forced within ±7° of the equator were referred to as equatorial Rossby waves. A simple statistical atmospheric model was coupled to the ocean model used in the Zebiak and Cane (1987) coupled model. The design of the statistical atmosphere allowed the structure and position of the wind stress anomalies to be externally prescribed. Using wind stress anomaly structures derived from FSU observations, a regular 5-yr ENSO oscillation was found in the coupled model. This simulation appeared to be consistent with the nonlinear delayed oscillator of Suarez and Schopf (1988).

The thermocline anomaly along the equator in the coupled simulation has a strong east–west dipole structure that changed sign with the ENSO cycle. Moreover, this dipole structure in the thermocline anomaly was largely in equilibrium with the zonal wind stress anomaly, demonstrating that the zonal wind stress anomaly is in equilibrium with the thermocline gradient anomaly along the equator. The zonal mean thermocline anomaly across the basin ($\bar{H}$), however, was not in equilibrium with the wind stress anomaly. This imbalance along the equator was best seen in the central Pacific, where the thermocline anomaly leads the wind stress anomaly. The control experiment indicates that time delays due to surface-layer processes are of secondary importance in
determining the timescale of the oscillation and that the memory of the coupled system resides with subsurface wave dynamics.

Eastward propagating Kelvin and westward propagating Rossby waves along the equator were detected in the control simulation. A continuous stream of Rossby wave reflections off the western boundary was associated with the transitions between extreme ENSO states. When the meridional structure of the wind stress anomaly was made to be broader (narrower), the simulated ENSO period increased (decreased). Consistent with changes in the curl of the wind stress anomaly in these experiments, the variability of the thermocline anomaly also shifted to higher (lower) latitudes. This relationship between the thermocline variance and the curl of the wind stress anomaly leads to the speculation that off-equatorial Rossby waves make a significant contribution to the simulated ENSO cycle.

A second set of sensitivity experiments was presented in which the effects of the off-equatorial Rossby waves at the western boundary were filtered out of the simulation. In these experiments, regardless of the structure of the wind stress anomaly, the model produces a 2-yr-period oscillation. It also appears that in these simulations, the model has shifted to a fundamentally linear delayed oscillator regime from the nonlinear regime in the unfiltered models.

In the delayed oscillator theory, the amplitude of the Kelvin waves reflecting off the western boundary is a critical component in defining the period of the oscillation. Reducing the amplitude of the Kelvin waves increases the period, and increasing the amplitude decreases the period of the oscillation. In the experiments presented here, the off-equatorial Rossby waves reduce the amplitude of the Kelvin waves at the western boundary, which increases the model ENSO period. Changes in the structure of the wind stress anomaly leads to changes in the oscillation period. This sensitivity is because the phase speed of the off-equatorial Rossby waves, which depends on the latitude at which they are forced, is considerably slower than that of the equatorially trapped waves. This phase difference leads to destructive interference, due to the off-equatorial Rossby waves, in determining the amplitude of the reflected Kelvin wave at the western boundary. Broader meridional structures lead to slower off-equatorial Rossby wave phase speeds, which increases the period of the oscillation. The strong dependence of the simulated period on the meridional structure of the wind stress anomaly leads to the conclusion that the atmospheric equatorial radius of deformation and the ocean basin geometry are important factors in determining the characteristic ENSO frequency in this model. While the off-equatorial Rossby waves have a large influence on the simulated period, the equatorially trapped modes are required to produce oscillatory behavior.

As with many modeling studies, the results and conclusions presented here must be interpreted within the context of the limitations of the model employed. In this case, there are a number of model deficiencies beyond the oversimplified statistical atmosphere. For instance, the ocean model includes only one baroclinic mode, so that all the energy of the wind stress forcing must project onto this single mode, whereas in the real ocean, the wind stress forces a continuous spectrum of vertical modes. Given the simplified horizontal boundaries and boundary conditions, it is possible that the model overemphasizes the wave reflections off the eastern and western boundary. In fact, some observational studies question whether there is a clear meridional boundary beyond 8° of the equator (e.g., Clarke 1983). The ocean model also employs linear dynamics and the long-wave approximation, and it is not clear whether nonlinearities are important in the real ocean. There are also observational studies that indicate that the importance of the off-equatorial Rossby waves found here may be overestimated (e.g., Kessler 1991). In order to address some of these issues, we are in the process of investigating the role of off-equatorial Rossby waves using a more realistic ocean component in the coupled model.

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APPENDIX

Wind Stress Modifications

The modifications to the spatial structure functions $\alpha$ and $\beta$ for the broad and narrow experiments are described below. For the broad experiments, the modified structure functions are given by

$$\alpha_{\text{broad}}(x, y) = (y^2/\lambda^2 + 1)e^{-C_1r^2/\lambda^2}\alpha(x, y)$$

and

$$\beta_{\text{broad}}(x, y) = (y^2/\lambda^2 + 1)e^{-C_1r^2/\lambda^2}\beta(x, y),$$

where $y$ is the distance from the equator, $\lambda$ is the equatorial radius of deformation, and $C_1 = 0.1$. The structure functions $\alpha_{\text{broad}}$ and $\beta_{\text{broad}}$ are then normalized to preserve the meridional standard deviation at each longitude. The final structure functions are shown in Figs. 3a and 3b. For the narrow experiments, the structure functions are given by

$$\alpha_{\text{narrow}}(x, y) = e^{-r^2/\lambda^2}\alpha(x, y)$$

and

$$\beta_{\text{narrow}}(x, y) = e^{r^2/\lambda^2}\beta(x, y).$$
As in the broad experiment, the structure functions \( \alpha_{\text{broad}} \) and \( \beta_{\text{broad}} \) are then normalized to preserve the meridional standard deviation for each longitude. These structure functions are also shown in Figs. 3a and 3b.

REFERENCES


