The Response of a Stochastically Forced ENSO Model to Observed Off-Equatorial Wind Stress Forcing

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ABSTRACT

This study investigates the response of a stochastically forced coupled atmosphere–ocean model of the equatorial Pacific to off-equatorial wind stress anomaly forcing. The intermediate-complexity coupled ENSO model comprises a linear, first baroclinic mode, ocean shallow water model with a steady-state, two–pressure level (250 and 750 mb) atmospheric component that has been linearized about a state of rest on the β plane. Estimates of observed equatorial region stochastic forcing are calculated from NCEP–NCAR reanalysis surface winds for the period 1950–2006 using singular value decomposition. The stochastic forcing is applied to the model both with and without off-equatorial region wind stress anomalies (i.e., poleward of 12.5° latitude). It is found that the multiyear changes in the equatorial Pacific thermocline depth “background state” induced by off-equatorial forcing can affect the amplitude of modeled sea surface temperature anomalies by up to 1°C. Moreover, off-equatorial wind stress anomalies increased the modeled amplitude of the two biggest El Niño events in the twentieth century (1982/83 and 1997/98) by more than 0.5°C, resulting in a more realistic modeled response. These equatorial changes driven by off-equatorial region wind stress anomalies are highly predictable to two years in advance and may be useful as a physical basis to enhance multiyear probabilistic predictions of ENSO indices.

1. Introduction

El Niño–Southern Oscillation (ENSO) is the dominant mode of interannual variability in the earth’s climate system; it affects the climate over many parts of the world (Ropelewski and Halpert 1989; Philander 1990). Over the last few decades, mechanistic studies with simple and intermediate-complexity coupled models have contributed to our theoretical understanding of the oscillatory (e.g., Cane and Zebiak 1985; Zebiak and Cane 1987; Battisti and Hirst 1989; Schopf and Suarez 1988; Jin 1997a,b) and seasonal phase locking behavior of ENSO (e.g., Tziperman et al. 1997, 1998; Neelin et al. 2000). Through this work, ocean dynamics are shown to be integral to the ENSO mode, and the observed irregular periodicity of ENSO can be explained as either self-sustained chaotic (nonlinear) behavior (e.g., Jin et al. 1994; Tziperman et al. 1994, 1995) or the output of a damped stable system that relies on weather “noise” to introduce irregularity and sustain the ENSO mode (e.g., Kleeman and Power 1994; Penland et al. 2000; Zavala-Garay et al. 2003; Batstone and Hendon 2005). The role of weather noise is supported by the diagnostic analysis of observed sea surface temperatures (SSTs; e.g., Penland and Sardeshmukh 1995), subsurface temperatures (e.g., Kessler 2002), and several recent modeling studies (e.g., Zavala-Garay et al. 2003; Batstone and Hendon 2005).

Although ENSO variability is dominated by interannual signals, interdecadal variations in the magnitude and frequency of ENSO events and their teleconnections are apparent in both observations and models (e.g., Wang 2009 American Meteorological Society

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and Ropelewski 1995; Allan et al. 1996; Power et al. 1999; Fedorov and Philander 2000; Mann et al. 2000; Walland et al. 2000; Arblaster et al. 2002; Vimont et al. 2002; Power et al. 2006). A number of complementary mechanisms have been proposed to account for the interdecadal variability of ENSO. For example, (i) random internally generated atmospheric variability can drive interdecadal variability in ENSO (e.g., Kleeman and Power 1994; Kleeman and Moore 1997; Newman et al. 2003; Power and Colman 2006; Power et al. 2006), (ii) the nonlinear dynamics of the coupled system affects the system on interdecadal time scales (e.g., Jin et al. 1994; Tziperman et al. 1994; Timmermann et al. 2003; Rodgers et al. 2004; Schopf and Burgman 2006), and (iii) variability generated in the extratropics subsequently forces changes in the tropics on interdecadal time scales. Extratropical variability can be transmitted to the tropical Pacific Ocean via the atmosphere (e.g., Barnett et al. 1999; Pierce et al. 2000) via the ocean by changes to the shallow meridional overturning circulation (e.g., Kleeman et al. 1999; McPhaden and Zhang 2002; Nonaka et al. 2002) or by extratropical oceanic Rossby waves driven by wind stress variability (e.g., Lysne et al. 1997; Liu et al. 1999; Capotondi and Alexander 2001; Capotondi et al. 2003; Wang et al. 2003a,b; Li and Clarke 2007).

The Rossby wave theories are of particular interest here. In these theories the westward-propagating extratropical Rossby waves driven by wind stress changes can modulate the equatorial region thermocline depth in two ways. First, these waves can induce a geostrophic transport that allows for exchanges between the off-equatorial and equatorial regions. Second, these Rossby waves can impinge on the Pacific Ocean western continental boundary, where they create coastally trapped waves that propagate equatorward. Upon reaching the equator, these coastally trapped waves excite low-frequency equatorially trapped Kelvin waves that drive changes in the equatorial thermocline depth. The resulting changes in equatorial thermocline depth exhibit very little zonal asymmetry (McGregor et al. 2007). Further, it has been suggested that near-zonal variations in the equatorial thermocline depth affect the magnitude and frequency of ENSO events by modulating the “background” state in which ENSO operates (e.g., Fedorov and Philander 2000).

Numerous investigators have studied the effects of extratropical oceanic Rossby waves on the equatorial Pacific Ocean with models of varying complexity (e.g., Lysne et al. 1997; Liu et al. 1999; Capotondi and Alexander 2001; Hazeleger et al. 2001; Capotondi et al. 2003; Galanti and Tziperman 2003; Luo et al. 2003; Wang et al. 2003a,b; Wu et al. 2003; McGregor et al. 2004, 2007; Tourre et al. 2005). The overall message from these studies is that although there is uncertainty regarding the latitudes of importance for the extratropical Rossby wave propagation and the magnitude of the equatorial region response, extratropical oceanic Rossby waves affect the background state of the equatorial Pacific Ocean. Hence, extratropical ocean transfers can be used to explain at least part of the decadal SST and ENSO variability of the Pacific Ocean. In an elegant study, Wang et al. (2003a,b) went one step further and combined off-equatorial wind stress forcing with ENSO theory to produce a unified theory for decadal variability and ENSO modes. However, the manner in which the changes in Pacific Ocean background state induced by extratropical Rossby waves actually interact with and affect ENSO variability in a more complex setting has not been studied extensively to date.

In this study we use a stochastically forced (SF), intermediate-complexity, coupled ENSO model to investigate how these extratropical-to-tropical transfers along the western boundary affected the equatorial Pacific ENSO variability over the period 1950–2006. We are primarily interested in understanding the magnitude of the effect of off-equatorial forcing on the equatorial thermocline response, whether it can help to explain differences in ENSO variations before and after the mid-1970s noted previously (e.g., Fedorov and Philander 2000), and whether the off-equatorial signals can enhance predictability of the coupled atmosphere–ocean system in the equatorial Pacific. The externally prescribed stochastic forcing (weather noise) estimated from observations is used as forcing in our coupled model because the atmospheric component of the coupled model is actually a steady-state model in which intraseasonal variability, such as the Madden–Julian oscillation (MJO), is otherwise not expressed.

This paper is organized as follows: Section 2 describes the intermediate-complexity ENSO model used. In section 3 we estimate the stochastic forcing from 57 yr of the National Centers for Environmental Prediction National Center for Atmospheric Research (NCEP–NCAR) reanalysis surface winds. Section 4 investigates the role of idealized and realistic off-equatorial region wind stress forcing on the stochastically forced SST and ENSO variability of the equatorial Pacific Ocean. The implications of these results on predictability are investigated in section 5. Finally, section 6 provides an overall discussion of the results and conclusions.

2. The coupled ENSO model

The ocean component of the intermediate-complexity coupled ENSO model is very similar to the shallow water model (SWM) ocean used by McGregor et al.
It is a linear reduced-gravity SWM configured at 1° resolution for the low- to midlatitude Pacific Ocean between 41°S–41°N and 120°E–68°W. The western and eastern Pacific boundaries approximately follow the coastlines of Australasia and the Americas respectively; north and south artificial boundaries are at 41°N and 41°S. The density structure of the 1 1/2-layer system is composed of an active upper layer of uniform density overlaying a deep motionless lower layer of larger uniform density. These density layers are separated by an interface (pycnocline) that approximates the ocean’s thermocline. Motion in the upper layer is driven by the applied wind stress (per unit density) $\tau$ ($m^2 s^{-2}$), and the associated response is displayed by the vertical displacement of the thermocline $h$ ($m$) and the horizontal velocity components ($u$ and $v$) of the flow ($m s^{-1}$).

Unlike in McGregor et al. (2007, 2008), where the SWM uses a first baroclinic mode gravity wave speed of $c_1$ everywhere, in this study the SWM uses geographically varying first baroclinic mode gravity wave speeds following Chelton et al. (1998) (see Fig. 1). The inclusion of the geographically varying first baroclinic mode gravity wave speed $c_1$ allows us to use a specified $H$ (here $H = 300$ m; Tomczak and Godfrey 1994, p. 37) and the relationship

$$c_1 = (g' H)^{1/2}$$

(1)

to calculate a geographically varying value for the reduced gravity $g'$. We chose to geographically vary $g'$ as opposed to $H$ because the model solutions otherwise become unstable around sharp bathymetric features when $H$ is varied. Geographic variations in $g'$ are also more realistic and therefore beneficial in regions of larger density gradients around bathymetric features and in terms of the west–east sloping thermocline and its zonal gradient in stratification.

As in the intermediate coupled model of Kleeman (1993), the SST anomaly (SSTA) equation was simplified by assuming a fixed meridional structure that decays away from the equator with an $e$-folding radius of 10°. Changes in the SSTA on the equator are modeled by the equation

$$\frac{dT}{dt} = \alpha(x) h(x) - \varepsilon T,$$

(2)

where $T$ is the SSTA at time $t$, $\alpha$ is a longitude-dependent constant that relates the modeled oceanic thermocline depth displacement $h$ along the equator to the SSTA, $\varepsilon$ is the Newtonian cooling coefficient ($2.72 \times 10^{-7}$ s$^{-1}$), and $x$ is the longitude. The zonal dependence of $\alpha$ reflects the fact that the equatorial thermocline is deeper and has less influence on SST in the west than in the east ($\alpha = 3.4 \times 10^{-8}$ in the east and reduces linearly west of 140°W to a minimum of $\alpha/5$ at the western equatorial boundary). SSTAs computed from (2) are used to compute heating anomalies that drive changes in the atmospheric component of the model.

The atmospheric component of the model is the steady-state, two–pressure level model (250 and 750 mb) of Kleeman (1991). The model has been linearized about a state of rest on the $\beta$ plane (see, e.g., Gill 1982), and the phase speed of the first baroclinic mode is $\sim 60$ m s$^{-1}$. Rayleigh friction and Newtonian cooling are included in the momentum and thermodynamic equations, respectively. These terms combine to give a damping time scale of approximately 3 days. The atmospheric heating used to drive the model is made up of a direct thermal heating component from the modeled SST anomalies and a latent heating component derived from a vertically integrated steady-state moisture equation. Latent heating and modeled precipitation exhibit a strong tendency to reside in the western equatorial
Pacific where total SST is high. This heating produces a first baroclinic mode response at the 250-mb level that is equal and opposite to that at the 750-mb level. The surface wind anomalies generated in the model are converted to surface wind stresses to force the ocean using the linear stress law:

\[ (\tau_x, \tau_y) = \rho_a C_D g \gamma(U, V). \]  

(3)

Here, \( \rho_a \) is the density of air, \( C_D \) is the dimensionless drag coefficient, and, as in Zavala-Garay et al. (2003), \( \gamma \), which typically represents the mean wind speed (6 m s\(^{-1}\)) in this study, acts as a coupling coefficient between the ocean and atmosphere. For the values of ocean and atmosphere model parameters selected, the coupled ocean–atmosphere system produces a damped oscillation consistent with the ENSO theories outlined previously (e.g., Penland et al. 2000).

Note that the coupled model incorporates two important nonlinearities: (i) latent heating due to deep atmospheric convection is only active when SST > 28.5°C and (ii) the effect of the thermocline displacement \( \eta \) on SST is capped, such that |\( \eta \)| > 37.5 m is equal to 37.5 m. The latent heating nonlinearity is incorporated into the model to represent the fact that observed equatorial region penetrative precipitation is predominantly located in areas where the SST is above approximately 27.5°C–28°C (e.g., Hirst 1986; Graham and Barnett 1987). The main purpose of capping the thermocline’s effect on SST is to prevent a runaway coupled instability. This nevertheless also crudely mimics the nonlinear behavior in the system. For example, when considering a large enough thermocline shoaling (deepening), the temperature of the subsurface water entrained into the mixed layer becomes equal to the mixed layer temperature; hence, no larger negative (positive) SSTAs are possible.

3. Coupled model forcing

a. Data

The atmospheric data used in this study were the daily averaged surface winds for the period 1950–2006 from the NCEP–NCAR reanalysis (Kalnay et al. 1996). Here, the NCEP–NCAR surface winds, which are available on a 2.5° × 2.5° grid, were interpolated onto the 1° × 1° SWM grid. The SST datasets used were the monthly reconstructed SST of Smith and Reynolds (2004) for the period 1950–81 and the weekly Reynolds and Smith (1994) optimum interpolation SSTs for the period 1982–2006. These SST datasets were spatially interpolated onto the 1° × 1° SWM grid and then linearly interpolated in time to produce daily values of SST compatible with the NCEP–NCAR surface winds. Both datasets were restricted to the Pacific domain between 40°S–40°N and 120°E–60°W. The long-term mean and annual cycles were removed to form daily anomalies of SST and surface wind.

To investigate how extratropical forcing affects modeled ENSO variability, we divided the model domain into two regions. These are an equatorial region spanning between 12.5°S and 12.5°N and an off-equatorial region consisting of all latitudes in the domain poleward of 12.5° latitude. In the equatorial case the surface wind anomalies are linearly tapered to zero in the poleward direction from 10° to 15° latitude, whereas in the off-equatorial case the surface wind anomalies are linearly tapered to zero in the equatorward direction from 15° to 10° latitude. This tapering of the surface wind anomalies has an affect on the wind stress curl in the tapper region. However, upon careful analysis of SWM ocean simulations forced by wind stresses with and without this artificial tapering, we are convinced that the artificial changes in the wind stress curl due to tapering of the surface winds have a negligible effect on the modeled equatorial region variability. In the experiments presented here, the surface wind stress anomaly forcing was either applied everywhere or only in the equatorial or off-equatorial regions. As described in section 2, the surface wind anomalies were converted to wind stress anomalies using the linear stress law [Eq. (3)] in preparation for the forcing experiments.

b. Estimation of stochastic equatorial region surface winds

Here, as in Zavala-Garay et al. (2003) and Batstone and Hendon (2005), stochastic surface winds are defined as the component of observed daily atmospheric variability that is not linearly related to the underlying SST at zero lag. The rationale behind this definition is that the atmosphere has much shorter adjustment times (on the order of days) than the adjustment times of the equatorial ocean. Thus, the components of atmospheric variability that are not directly due (coupled) to variations in the underlying SST can be considered stochastic. To identify the component of the equatorial region (recall that it is defined between 12.5°S and 12.5°N) NCEP–NCAR surface winds that is uncorrelated to the underlying SSTs, we use the method that was used in Zavala-Garay et al. (2003). A singular value decomposition (SVD) was carried out on the equatorial region SST and the \( x \) and \( y \) components of the surface winds using the method detailed in Bretherton et al. (1992).

The first SVD mode for the period 1950–2006 is shown in Fig. 2a. The STA pattern resembles the mature phase of El Niño and the surface wind anomalies in
the western/central equatorial Pacific Ocean appear to be representative of the shift of surface winds during an El Niño event. This first mode explains approximately 91% of the covariance between the equatorial region surface wind and SST anomalies. The projection coefficients display temporal variability with large deviations in the ENSO years that occurred during the 1950–2006 period (Fig. 2b; only SSTA projection coefficients are displayed). The higher-order SVD modes explain progressively less of the total covariance and the spatial patterns get more complex in structure. The coupled component of the equatorial region surface wind anomalies is defined as the component of the surface wind anomalies that is linearly related to the underlying SSTA. Here, the first seven modes of the SVD analysis were used to identify the “coupled” (linearly related) surface winds as they combine to explain approximately 99% of the equatorial region surface wind and SST anomaly covariance. We then calculate the stochastic component of the surface winds by simply subtracting the coupled component of atmospheric variability from the original surface wind anomaly field. The estimated stochastic surface wind anomalies are dominated by noise in the western tropical region and the extratropics (cf. Fig. 3); when decomposed using an EOF analysis, the associated temporal variability of the first 50 EOFs (which explain >99% of the covariance) have decorrelation times of less than 9 days. These estimates of stochastic surface wind forcing were then converted to wind stresses using the linear stress law [Eq. (3)] for use in the couple model experiments. We note that it is possible that small components of the stochastic wind stresses may not be independent of SST, as has been shown by Vecchi et al. (2006).

4. Stochastically forced coupled model experiments

a. Experiment I—The standard run

In the first experiment, referred to as the standard run, the coupled model simulation is stochastically forced by the equatorial region stochastic surface wind stress anomalies defined in section 3b. An analysis of the Niño-3 region (Niño-3) SSTA produced by the run suggests that the SF model does a reasonable job in reproducing the larger observed ENSO events [e.g., 1957/58, 1972/73, 1982/83, and 1997/98 (Fig. 4a)]. This is confirmed in Table 1 by the correlation coefficient of 0.47 between the model output and observed Niño-3 SSTA (statistically significant at the 99% level). Note that the statistical significance of all correlation coefficients determined in this study takes account of serial (auto-) correlation in the series based on the reduced effective number of degrees of freedom outlined by Davis (1976). Upon visual analysis, it is clear that the standard run simulates the timing and magnitude of ENSO events post-1980 better than it does prior to 1980. The correlation coefficient between Niño-3 modeled and observed SST prior to 1980 is weaker (0.38; statistically significant at the 95% level), whereas the corresponding post-1980 correlation is 0.68 (statistically significant at the 99% level). Separation of the modeled data into before and after 1980 for comparison is appropriate for two reasons: (i) in observations there was a marked shift in the background state of the Pacific Ocean, ENSO statistics, and variability reported in the late 1970s (e.g., Mann et al. 2000) and (ii) there is a possibility that synoptic-scale disturbances in the reanalysis winds might not be well represented prior to 1980.
because of the lack of assimilated satellite observations (Zhang and Gottschalck 2002).

b. Experiment II—Idealized off-equatorial patch forcing

To test what effect an off-equatorial wind stress anomaly–forced change in the equatorial thermocline depth could have on the ENSO variability in the standard run, we carried out two coupled model experiments. These SF coupled model experiments utilize an idealized wind stress distribution that is located in the off-equatorial region as defined in the earlier work of Wang et al. (2003a). It is essentially a patch of wind stress anomaly that zonally spans 30° and has maximum wind stress anomaly amplitude of ±0.024 N m⁻² that decays meridionally from the maximum with an e-folding scale of 5°. This idealized wind stress anomaly patch produces two centers of wind stress curl of opposite sign north and south of the maximum wind stress location. For the experiments here, this wind stress patch is centered at 20°S and 120°W. This location was selected so the equatorward center of wind stress curl generated by the patch sits at approximately 15°S and 120°W, which is one of the locations identified by McGregor et al. (2007)
as being important for off-equatorially forced Niño-3 thermocline depth variability. When applied alone, this wind stress anomaly patch produces a near-zonal change in equatorial thermocline depth with an amplitude of approximately 0.4 m; the sign of the equatorial response depends on the sign (direction) of the wind stress anomaly used. The equatorial thermocline takes approximately 30 months to adjust to the patch of off-equatorial wind stress anomaly, which is consistent with the time it takes a Rossby wave at 15° latitude to propagate from 120°W to the western Pacific continental boundary and then into the equatorial zone as a Kelvin wave (not shown; see Fig. 3 in Wang et al. 2003a). The equatorial response of 0.4 m to this idealized patch is consistent with expectations from linear theory (e.g., Kessler 1991).

When calculated from the observed mean thermocline depth [estimated as the 20°C isotherm from SODA data for the period 1979–2006 (Carton et al. 2000)], a ±0.4 m change in zonal thermocline can alter the equatorial Pacific Warm Water volume by ~2%. Taken with the associated SSTA response of around ±0.1°C (again depending on the sign of the idealized surface wind anomaly patch), a conservative estimate of the change heat content above the thermocline is ~8 × 10^{20} J in the equatorial Pacific Ocean.

The two experiments presented here utilize equatorial region stochastic forcing and an idealized patch of wind stress anomaly forcing in the off-equatorial region as described above. The difference between these experiments is that experiment IIa utilizes a positive wind stress anomaly forcing patch (a westerly wind anomaly), whereas experiment IIb utilizes a negative wind stress anomaly forcing patch (an easterly wind anomaly). This effectively lowers the equatorial thermocline depth in experiment IIa by 0.4 m and raises it in experiment IIb by 0.4 m. The Niño-3 SSTA produced by both IIa and IIb again suggests that the SF coupled model does a reasonable job because many of the observed more intense ENSO episodes during 1950–2006 are successfully captured (Figs. 4b,c). However, what is clear from these experiments is that the relatively small zonal change in the equatorial thermocline depth (~±0.4 m) has had a large effect on the SSTA variability of the modeled ENSO events. For example, experiment IIa (with an equatorial thermocline depth 0.4 m deeper than normal) has a Niño-3 SSTA variance of 1.2°C^2 whereas experiment IIb (with an equatorial thermocline depth 0.4 shallower than normal) has a Niño-3 SSTA variance of 0.45°C^2 (see Table 1). Prior to 1980, experiment IIb provides a more accurate simulation of Niño-3 SSTA (correlation coefficient of 0.28, significant at the 95% level) than experiment IIa (correlation coefficient of only 0.01). The variance of SSTA variability in IIb prior to 1980 is 0.36°C^2, which compares more favorably with the corresponding observed value of 0.6°C^2 than does IIa (variance is 1.52°C^2). Post-1980, both experiments appear to do a good job of picking up the timing of ENSO events highlighted by the respective correlation coefficients of 0.66 between both modeled and observed Niño-3 SSTA (significant above the 99% confidence level). Despite this correspondence, experiment IIa successfully reproduces the amplitudes of the 1982/83, 1986/87, 1997/98, and 2002/03 events whereas experiment IIb fails significantly short. This is highlighted by the post-1980 variance of the IIa SSTA of 0.89°C^2, which compares much more favorably with the corresponding observed value of 0.8°C^2 than does the IIb SSTA (0.48°C^2).

The experiments indicate that changes in off-equatorial surface wind stress anomaly forcing can change the background state of the equatorial thermocline depth and that these changes in the background state significantly alter the magnitude of the SF ENSO variability produced. Further, from these results we can infer that a SF simulation that incorporates relatively modest deepening in the background mean equatorial thermocline depth around the late 1970s would provide more realistic Niño-3 SSTA variability than a simulation in which such a shift is absent.

c. Experiment III—The off-equatorially forced experiment

We pose two questions in the context of experiment II: 1) What is the equatorial response to observed off-
equatorial wind stress anomaly forcing? 2) Can off-equatorial forcing contributions help to explain the shift in ENSO statistics in the late 1970s? To answer these questions, we force the ocean component of the coupled model with off-equatorial region (those latitudes poleward of ±12.5°) NCEP–NCAR reanalysis surface wind stress anomalies. Hence, in this experiment there is no stochastic wind stress anomaly forcing applied (externally) in the equatorial region (within ±12.5° of the equator). It must be noted that this experiment was carried out with and without ocean–atmosphere coupling in the equatorial region and there was almost no difference between the results. It follows that the off-equatorial wind stress anomaly–forced equatorial region variability is insufficient to initiate an atmospheric feedback that significantly amplifies the original equatorial region variability. Nevertheless, this experiment shows that off-equatorial region wind stress anomaly forcing can produce thermocline depth anomalies in the Niño-3 region that are approximately 1 1/2 times larger than those forced by the idealized wind stress anomaly patch used in experiments IIa and IIb (between ±0.6 m; see Fig. 5). These off-equatorially forced Niño-3 thermocline depth anomalies reveal a strong interdecadal signal throughout the 57-yr simulation. As in the idealized off-equatorial patch experiments above, the equatorial thermocline depth changes are near-zonal, confirmed by the correlation coefficient of 0.85 between the Niño-3 and Niño-4 region thermocline depth anomalies (Fig. 5).

The role of western boundary interactions versus simple Ekman and geostrophic transports in the equatorial anomalies forced by off-equatorial region wind forcing was investigated. To this end, the western boundary damping term developed by McGregor et al. (2007) was used to damp out Rossby waves impinging on the entire western boundary. The inclusion of this term in the model prevents the initiation of coastal Kelvin waves that would otherwise transport the incoming Rossby signals equatorward, effectively allowing us to separate the equatorial wave dynamic signal. Results of this experiment show that more than 80% of the equatorial region variance is transferred via the western boundary (not shown). This is consistent with previous studies that identify Rossby waves incident on the western Pacific boundary as a mechanism linking the extra-equatorial with the equatorial Pacific Ocean (e.g., Capotondi and Alexander 2001; Galanti and Tziperman 2003; Wang et al. 2003a; White et al. 2003; McGregor et al. 2004).

Considering that the addition of an off-equatorial wind stress anomaly patch has been shown to significantly affect the amplitudes of the modeled ENSO variability in our experiments, it might be expected that the addition of observed off-equatorial region wind stress anomaly forcing would also significantly affect the amplitudes of the modeled ENSO events. With regard to whether off-equatorial forcing can help explain the shift in background state in the late seventies, we calculate the mean Niño-3 region thermocline depth anomalies from the off-equatorially forced experiment (experiment III) based on the periods prior to 1978 and between 1978 and 2000. The mean off-equatorially forced Niño-3 region thermocline depth anomaly is −0.06 m prior to 1978, whereas post-1978 the mean Niño-3 region thermocline depth anomaly is approximately +0.2 m. This indicates the occurrence of a moderate deepening in the thermocline depth background state in the late 1970s. As shown in experiment IIa, a deepening of the background state of the equatorial thermocline depth would correspond to larger-amplitude El Niños, consistent with the shift seen in observations. However, we note that the shift seen in the off-equatorially forced Niño-3 region thermocline depth anomalies appears to be more representative of a decadal fluctuation with peak amplitudes around the large El Niño events of 1982/83 and 1997/98.

d. Experiment IV—The dual region forcing experiment

The final coupled model experiment utilizes both the equatorial region stochastic forcing and off-equatorial region surface wind anomaly forcing, hereafter called the dual region forcing (DRF) experiment. The purpose of this experiment is to identify whether the off-equatorially forced equatorial region variability affects the timing
and amplitude of the SF ENSO events during the period 1950–2006. To assess the importance of off-equatorial wind stress anomaly forcing on the modeled ENSO variability during 1950–2006, the results of this experiment (experiment IV) will be directly compared to observations and the results of the standard run (experiment I) presented in section 4a (Fig. 6). As in the standard run (Fig. 6a), the Niño-3 SSTA produced by experiment IV (Fig. 6b) shows that the model does quite a good job of reproducing the observed variability as indicated by the correlation coefficient between this modeled and observed Niño-3 SSTA, which is also 0.46 (statistically significant at the 99% level). This result indicates that off-equatorial forcing has little effect on the timing of Niño-3 variability. However, it is clear that the addition of off-equatorial wind stress anomaly forcing to the coupled model has affected the amplitude of the Niño-3 SSTA variability because there are more significant deviations between the two time series (cf. Figs. 6a,b). For example, the addition of off-equatorial wind stress anomaly forcing in experiment IV has significantly increased the magnitude of the modeled ENSO events of 1982/83 and 1997/98, accounting for differences of up to 0.5°C and 0.7°C respectively, and making the model results more like observations. For the post-1980 period, the addition of off-equatorial forcing increased the SSTA variance from 0.51°C² to 0.85°C², which gives a much better match to the observed value of 0.79°C².

5. Implications for predictability

Based on results from the coupled model experiments in section 4, we have shown that ENSO events produced in our stochastically forced coupled model are sensitive to off-equatorially forced changes in the background state of the equatorial Pacific Ocean. Further, these long-time scale (decadal) off-equatorially induced changes in this background state appear to be due to oceanic Rossby waves, which themselves have long time scales compared to most equatorial region processes. This raises two questions: 1) Is the equatorial response to off-equatorial forcing predictable years in advance as expected? 2) Can these long time scale changes in the equatorial background state be used to improve ENSO predictability?

The potential predictive skill of equatorial thermocline depth anomalies forced by off-equatorial region wind stress anomaly forcing is examined by carrying out a sample of 100 4-yr hindcasts. These hindcasts were initialized with output from the off-equatorially forced model simulation in experiment III every 3 months, commencing January 1982 and ending October 2006. The model was then left to freely evolve (i.e., with no further wind stress anomaly forcing) for 4 yr and the modeled Niño-3 thermocline depth anomalies were compared with the off-equatorially forced modeled values in experiment III. The rationale behind this was that oceanic Rossby waves have relatively predictable propagation characteristics and speeds that depend on the latitude of formation. Thus, an oceanic Rossby wave initiated prior to the model restart would continue to propagate across the Pacific Basin after the restart without further wind stress forcing. These waves would eventually modulate the equatorial Pacific thermocline following their interaction with, and Kelvin wave generation at, the western boundary. The hindcast output
of these 100 simulations was stratified according to the number of months because the model was initiated following the start date, from month 1 through to month 48. Consequently, these predictability experiments produced 48 different Niño-3 time series, each having 100 values spaced 3 months apart. The first of these time series corresponds to the predicted equatorial thermocline depth anomalies with 1-month lead, the second with a 2-month lead, and so on. A corresponding set of values was also taken from the off-equatorially forced simulation in experiment III that we were attempting to hindcast, and for each of the 48 pairs of time series produced we calculated the correlation coefficient and the root-mean-square error (RMSE). In the case of persistence, each of the 48 hindcast model samples was replaced at the particular lag by a sample identical to the off-equatorially forced model simulation in experiment III. The result is that persistence is exceeded after approximately 2 months, which by definition has a correlation of 1.0 at zero lag (Fig. 7). Upon further analysis of the anomaly correlation and the RMSE in Fig. 7, it is clear that there is good predictive skill of the equatorial thermocline depth variability driven by surface wind anomaly forcing in the equatorial region with lead times out to and beyond about 2 yr.

Summing up, we have an off-equatorially forced change in the background state of the equatorial Pacific that can be predicted with good skill up to 2 yr in advance. We expect that adding this variability to our ENSO model could improve the predictive skill by reducing the RMSE, especially on the more intense El Niño events. Conversely, the addition of this variability is not expected to increase the predictable lead times of the modeled ENSO variability, which is governed by stochastic weather noise–forced equatorial wave dynamics. Nonetheless, a prediction of the background state of the equatorial Pacific Ocean can usefully improve our forecasts. In what follows, we briefly explain how this works.

First, consider the likelihood of an ENSO event occurring 1 yr ahead: for this task, we have very little deterministic model skill and the probability density function (PDF) of the Niño-3 SSTA (modeled by the post-1980 PDF curve of the dual region forcing experiment; not shown) suggests a 90% chance that the SSTA will fall between $-1.6^\circ$ and $1.4^\circ$C. In practical terms, this implies an El Niño or La Niña event exceeding, say, $\pm 1.6^\circ$C is equally likely. In the parlance of Bayesian probabilities, this is the prior probability distribution based on historical experience. Now, recall that changes in the background state of the thermocline depth are predictable up to 2 yr in advance, and let us make use of this skill to select a more appropriate PDF for the upcoming year—that is, a PDF that takes account of the background state of the tropical Pacific Ocean. In Bayesian terms, this is the a posteriori PDF. PDFs of SSTAs from the stochastically forced standard run (experiment I) and the two stochastically forced runs of experiments IIa and IIb show that relatively modest shifts in background state of the equatorial thermocline depth induced by off-equatorial wind stress anomaly forcing significantly alter the probability of a given SSTA and the likelihood of an intense El Niño or La Niña event (Fig. 8).

If, for example, the predicted off-equatorially forced equatorial signal is close to zero, then we can use the PDF curve of the standard run (experiment I) corresponding to a 90% chance that the anomaly will fall between $-1.3^\circ$ and $1^\circ$C. In other words, we have narrowed the range of temperature anomalies expected from the original PDF and further reduced the likelihood of an intense El Niño or La Niña event. Now, if instead a 0.4-m increase in thermocline depth background state...
was predicted, then we choose the PDF of experiment IIa, which is skewed toward positive SSTAs, and infer an increased probability of a more intense El Niño event. On the other hand, predicting a 0.4-m decrease in thermocline depth background state leads us to choose the PDF of experiment IIb. This suggests a significant increase in the probability of a negative SSTA in the PDF and thus an increased likelihood of a (more intense) La Niña event. Table 2 lists the probabilities for an El Niño or La Niña event conditional on both the depth of the equatorial thermocline and the SSTA. In conclusion, highly predictable changes in the background state of the equatorial Pacific Ocean due to off-equatorial wind stress anomaly forcing lead to more accurate forecasts of ENSO by allowing us to select the PDF distribution that best represents the range of SSTA that can be expected.

6. Discussion

In this study we have investigated how off-equatorial wind stress anomalies affect the modeled equatorial Pacific variability during the period 1950–2006. Our aim was to examine the impact of off-equatorial wind stress anomalies on the modeled interannual variability and to determine whether these anomalies could be used to help explain the marked shift in ENSO behavior that occurred in the late 1970s. We also sought to identify the implications of this for tropical Pacific predictability due to the relatively long adjustment times of the off-equatorial ocean. To investigate how extratropical-to-tropical transfers affect ENSO we first divided the NCEP–NCAR reanalysis surface winds for the period 1950–2006 into two forcing regions that we defined as “equatorial” and “off-equatorial.” The equatorial forcing includes only the reanalysis wind stress anomalies between 12.5°S and 12.5°N; the off-equatorial forcing includes only wind stress anomalies in the extra-equatorial zone (i.e., those latitudes poleward of 12.5° latitude). Using singular value decomposition analysis we then identified the stochastic component of the equatorial wind stress anomalies as the component of observed atmospheric variability that is not linearly related to the underlying wind stress anomalies in the extra-equatorial zone (i.e., those latitudes poleward of 12.5° latitude). Using singular value decomposition analysis we then identified the stochastic component of the equatorial wind stress anomalies as the component of observed atmospheric variability that is not linearly related to the underlying SSTA at zero lag. This equatorial stochastic forcing was then used to force the intermediate-complexity coupled ENSO model described in section 2 in a series of experiments investigating the role of changes in the equatorial Pacific background state due to off-equatorial wind stress anomaly forcing. The objective was to identify the role of wind stress forcing and the important regions of this forcing on the fundamental large-scale ocean dynamics. We did not attempt to address the origin of the wind stress variability in this study.

In the standard run, which incorporates equatorial region stochastic forcing alone (experiment I), the model was shown to reproduce many of the observed ENSO episodes of the period 1950–2006, such as the events of 1957/58, 1972/73, 1982/83, and 1997/98 (Fig. 4a). This was confirmed by the highly statistically significant (at the 99% level) correlation coefficient of 0.47 between the modeled and observed Niño-3 region (Niño-3) SSTA. Using an idealized patch of wind stress anomaly off the equator and equatorial region stochastic forcing in experiment II, we have shown that a relatively small zonal change in the equatorial thermocline depth (∼±0.4 m) can significantly alter the heat content of water above the thermocline in the central and eastern equatorial Pacific. In the coupled setting, the atmospheric latent heating nonlinearity ensures that this change in heat content has a dramatic affect on the modeled precipitation in the central equatorial Pacific where total SST is warm, which in turn acts to modulate the amplitude of the modeled ENSO events. For example, experiment IIa, in which the idealized wind stress anomaly patch deepens the equatorial background state thermocline depth by 0.4 m, has a Niño-3 SSTA variance.
of $1.2^\circ C^2$. This is significantly larger than the standard run Niño-3 SSTA variance ($0.6^\circ C^2$) and the corresponding SSTA variance of experiment IIb ($0.45^\circ C^2$), which incorporates a 0.4-m shoaling in the thermocline depth background state. Furthermore, these relatively modest shifts in background state of the equatorial thermocline depth induced by the idealized off-equatorial patch significantly alter the probabilities of a given SSTA and thus the likelihood of an intense El Niño or La Niña event (Fig. 8). It appears that changes in the equatorial Pacific background state induced by the idealized off-equatorial patch of wind stress anomaly forcing affects the sensitivity of the coupled model to stochastic forcing. The sensitivity of this result to the temperature threshold for latent heating was tested and it was found that this result is robust for a realistic range of temperature thresholds ($26.5^\circ C$–$28.5^\circ C$).

Observations of Niño-3 SSTA show a well-documented shift in the background state of the Pacific Ocean, ENSO statistics, and variability in the late 1970s (e.g., Rodgers et al. 2004). If the observed Niño-3 SSTA time series is separated into two parts, before and after 1980, the variance post-1980 of $0.8^\circ C^2$ is much larger than for the record prior to 1980, which has a variance of $0.6^\circ C^2$. If we look through the statistics of the standard run and the idealized surface wind patch experiments (Table 1), it is clear that a stochastically forced simulation in the absence of any wind stress anomaly patch prior to, say, 1975 but with a positive idealized off-equatorial wind stress anomaly patch post-1975 (allowing a couple of years for an equatorial response) should provide the most realistic Niño-3 SSTA (in terms of variances and correlations with observed). Therefore, a shift in the equatorial thermocline depth background state induced by wind stress anomaly forcing in the off-equatorial region during the late 1970s could be used to explain the apparent shift in ENSO activity seen in observations. Forcing the ocean component of the coupled model with observed off-equatorial region reanalysis wind stress anomalies produces a significant deepening of the equatorial thermocline around this period, consistent with the proposed change needed to explain the change in ENSO statistics. Thus, extratropical-to-tropical oceanic transfers driven by off-equatorial forcing may help to explain the climatic shift of the late 1970s.

In addition, the two most intense El Niño episodes in the last century (1982/83 and 1997/98) occurred when the slow decadal time scale oscillation apparent in the off-equatorially forced equatorial thermocline depth reached peak deepening (Fig. 5). This raises the question: Would the El Niño events of 1982/83 and 1997/98 have been as intense without the contribution from off-equatorial wind stress forcing? Interestingly, the intense ENSO events of 1982/83 and 1997/98 are significantly larger in our model and have more realistic amplitudes in the stochastically forced coupled model experiment that incorporates off-equatorial wind stress anomaly forcing (section 4d) than just stochastic forcing alone in the standard simulation. In fact, the addition of off-equatorial surface winds has increased the amplitudes of these two particular events by $0.5^\circ$ and $0.7^\circ C$, respectively. We have shown in section 5 that there is good predictive skill with lead times out to 24 months of the equatorial region thermocline depth variability driven by reanalysis wind stress anomalies in the off-equatorial region. Hence, forecasts of off-equatorially forced variability could be used to give an indication of the upcoming background state of the equatorial Pacific. Further, we proposed in section 5 that these highly predictable changes in the background state of the equatorial Pacific Ocean could lead to more accurate probabilistic forecasts of ENSO by allowing us to select the PDF distribution that best represents the range of SSTA that can be expected.

We note that stochastic forcing was imposed as an external forcing that is independent of the state of the system and it is possible that components of the SF variability are not independent of SST, as has been shown by Vecchi et al. (2006). We conclude from this study that stochastic forcing can be used to maintain and explain ENSO and that changes in the background state of the equatorial Pacific thermocline depth induced by off-equatorial wind stress anomaly forcing have a significant effect on the amplitude of the modeled ENSO events. Further, we find that the shift in ENSO statistics of the late 1970s could in part be due to a change in the background state induced by off-equatorial forcing. Thus, the relatively small, potentially predictable, off-equatorial contribution to equatorial region variability appears critical for thresholds of ENSO. As such, we believe that a comprehensive understanding and more accurate seasonal-to-interannual prediction of ENSO variability could be gained by taking extratropical-to-tropical Pacific Ocean transfers into account.

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