Interdecadal Sea Surface Temperature Variability in the Equatorial Pacific Ocean. Part II: The Role of Equatorial/Off-Equatorial Wind Stresses in a Hybrid Coupled Model

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ABSTRACT

Many modeling studies have been carried out to investigate the role of oceanic Rossby waves linking the off-equatorial and equatorial Pacific Ocean. Although the equatorial ocean response to off-equatorial wind stress forcing alone tends to be relatively small, it is clear that off-equatorial oceanic Rossby waves affect equatorial Pacific Ocean variability on interannual through to interdecadal time scales. In the present study, a hybrid coupled model (HCM) of the equatorial Pacific (between 12.5°S and 12.5°N) was developed and is used to estimate the magnitude of equatorial region variability arising from off-equatorial (poleward of 12.5° latitude) wind stress forcing. The HCM utilizes a reduced-gravity ocean shallow-water model and a statistical atmosphere derived from monthly output from a 100-yr Australian Bureau of Meteorology Research Centre (now the Centre for Australian Weather and Climate Research) coupled general circulation model integration. The equatorial region wind stress forcing is found to dominate both the interannual and interdecadal SST variability. The equatorial response to off-equatorial wind stress forcing alone is insufficient to initiate an atmospheric feedback that significantly amplifies the original equatorial region variability. Consequently, the predictability of equatorial region SST anomalies (SSTAs) could be limited to ~1 yr (the maximum time it takes an oceanic Rossby wave to cross the Pacific Ocean basin in the equatorial region). However, the results also suggest that the addition of off-equatorial wind stress forcing to the HCM leads to variations in equatorial Pacific background SSTA of up to almost one standard deviation. This off-equatorially forced portion of the equatorial SST could prove critical for thresholds of El Niño–Southern Oscillation (ENSO) because they can constructively interfere with equatorially forced SSTA of the same sign to produce significant equatorial region ENSO anomalies.

1. Introduction

The El Niño–Southern Oscillation (ENSO) phenomenon is the dominant mode of interannual variability in the earth’s climate system, and it affects the climate over many parts of the world (Ropelewski and Halpert 1987; Philander 1990; Wang et al. 2003a). While ENSO variability is dominated by interannual signals, an interdecadal variation in the magnitude and frequency of ENSO events and their teleconnections is apparent in both observations and models (e.g., Wang and Ropelewski 1995; Allan et al. 1996; Power et al. 1999; Mann et al. 2000; Walland et al. 2000; Arblaster et al. 2002; Vimont et al. 2002; Kiem and Franks 2004; Power et al. 2006). There is also an ENSO-like interdecadal variation of SST present in observations (e.g., Mantua et al. 1997; Zhang et al. 1997; Folland et al. 1999; Power et al. 1999; Folland et al. 2002) and models of the Pacific Ocean (e.g., Walland et al. 2000; Arblaster et al. 2002; Power and Colman 2006; Power et al. 2006). This interdecadal SST pattern is referred to as either the Pacific decadal oscillation (PDO) when the emphasis is on the North Hemisphere (Mantua et al. 1997) or the interdecadal Pacific oscillation (IPO) if both hemispheres are considered (Power et al. 1999). Evidence has been
produced to show that the interdecadal variations of ENSO could be linked with the interdecadal SST variability of the IPO (Power et al. 1999; Arblaster et al. 2002). However, Power et al. (2006) have shown that the relationship between the IPO and ENSO could be due to the nonlinear behavior of ENSO teleconnections.

At the present time, the mechanisms responsible for the Pacific Ocean’s interdecadal SST and climate variability are unclear, and there is still some debate as to whether the IPO operates independently of ENSO. However, many hypotheses have been proposed (e.g., Latif 1998; Miller and Schneider 2000; Power and Colman 2006). For example, instabilities in the atmosphere can drive internal atmospheric variability and heat flux variability on time scales of up to and beyond a decade, which may alter the surface climate (e.g., Frankignoul and Hasselmann 1977; James and James 1992; Power et al. 1995; Frankignoul et al. 1997). Interdecadal variability can also arise in association with long-time-scale changes in ENSO activity, which can be generated in the tropical Pacific either through nonlinear mechanisms (e.g., Jin et al. 1994; Tziperman et al. 1994; Timmermann et al. 2003; Rodgers et al. 2004; Schoop and Burgman 2006), or by the reddened oceanic response to the stochastic nature of ENSO events (e.g., Newman et al. 2003; Power and Colman 2006). Variability generated in the extratropics can also influence the tropics on interdecadal time scales in a number of different ways. This extratropical variability can be transmitted to the tropical Pacific Ocean via the atmosphere (e.g., Barnett et al. 1999; Pierce et al. 2000). This extratropical variability can also be transmitted to the tropics via the ocean by (i) changing the temperature of the water advected into the tropics in the meridional overturning circulation (Gu and Philander 1997; Zhang et al. 1998), (ii) changing the rate at which water in the meridional overturning circulation is advected into the tropics (Kleeman et al. 1999; McPhaden and Zhang 2002; Nonaka et al. 2002), or (iii) exciting oceanic Rossby waves by variations in the extratropical wind stress (e.g., Lysne et al. 1997; Liu et al. 1999; Capotondi and Alexander 2001; Capotondi et al. 2003; Wang et al. 2003a,b).

In theories detailing the role of oceanic Rossby waves linking the extratropical and tropical Pacific Ocean, westward-propagating extratropical Rossby waves 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 modulate the equatorial thermocline depth via equatorial Kelvin waves that propagate eastward back across the basin. This Rossby wave theory has recently been expanded and included in theories of interdecadal variability that are explained in a similar context to the recharge–discharge ENSO paradigm (Wang et al. 2003b), and the delayed oscillator mechanism for ENSO (White et al. 2003; Tourre et al. 2005). There have been numerous modeling studies investigating the role of oceanic Rossby waves in linking the extratropical and tropical Pacific Ocean (Lysne et al. 1997; Liu et al. 1999; Capotondi and Alexander 2001; Capotondi et al. 2003; Hazeleger et al. 2001; Galanti and Tziperman 2003; Luo et al. 2003; Wang et al. 2003a,b; Wu et al. 2003; McGregor et al. 2004, 2007, hereafter Part I; Tourre et al. 2005). These studies have been carried out using models of varying complexity from simple analytical models and intermediate-complexity models through to sophisticated general circulation models. There is uncertainty regarding the latitudes of importance for the extratropical Rossby wave propagation and the magnitude of the equatorial region response. For example, in Part I we identify Rossby waves at 16°N and 14°S as being important, while Hazeleger et al. (2001) highlight the importance of Rossby waves poleward of 30° latitude. However, these studies suggest that extratropical wind stress–forced oceanic Rossby waves play a role in affecting the tropical Pacific SST on interdecadal time scales.

In this study we address whether the relatively small equatorial ocean response to off-equatorial wind stress forcing is sufficient to initiate a coupled response that amplifies the original equatorial response. To address this question we develop a hybrid coupled model (HCM) of the equatorial Pacific. The HCM incorporates a linear reduced-gravity shallow-water model (SWM) ocean with the Zebiak and Cane (1987) thermodynamic SST equation, and a statistical atmosphere derived from the analysis of output from a 100-yr coupled GCM (CGCM) simulation. A series of hybrid coupled model experiments are then performed to investigate the role of off-equatorial wind stress forcing (and oceanic Rossby waves) on the interannual and interdecadal SST variability of the equatorial Pacific Ocean.

This paper is organized as follows. Section 2 describes the CGCM [the Australian Bureau of Meteorology Research Centre (BMRC; now the Centre for Australian Weather and Climate Research) coupled atmosphere–ocean–sea ice GCM version 2.2 (BCM2.2)] and the century time-scale simulation. Section 3 provides a description of the SWM and the BCM2.2 wind stress–forced SWM simulation. Section 4 describes the HCM, and section 5 details the HCM experiments along with the presentation of results. A discussion and conclusions from the study are provided in section 6.
2. CGCM and century-scale simulation

The CGCM used in this study is BCM2.2, which is designed for seasonal prediction and climate change research with a focus on the tropical Pacific (Power et al. 1998). It has been shown to exhibit realistic interannual variability in the tropical Pacific and realistic Rossby and Kelvin wave propagation (Power et al. 1998; Bettio 2007). BCM2.2 incorporates the BMRC unified an atmospheric general circulation model, which is a spectral model configured at the horizontal resolution of rhomboidal wave 21 (R21), with 17 vertical levels. The ocean model component is a global version of the Princeton University Geophysical Fluid Dynamics Laboratory Modular Ocean Model (Pacanowski et al. 1991; Power et al. 1995). The meridional grid spacing increases from 0.5°, within 7° latitude of the equator, to a maximum of 5.8° near the North Pole. The zonal spacing is constant at 2° over the entire globe. Full details of the model’s atmospheric, ocean, and sea ice components and some of the interdecadal variability evident in the model are described by Power et al. (2006) and Power and Colman (2006).

A century time-scale simulation using BCM2.2 was performed where the model simulation was left to freely evolve. Monthly averages of the wind stresses from the 100-yr simulation were interpolated onto the 1° × 1° SWM grid to allow us to force the intermediate-complexity SWM (see section 3). Monthly averages of the SST, and vertically averaged temperature (VAT) to 300-m depth from the simulation were also interpolated onto the 1° × 1° SWM grid and have been analyzed using empirical orthogonal functions (EOFs) to determine the important modes of interannual and interdecadal SST variability. These analyses were presented in Part I where it was shown that the model produces Pacific Ocean ENSO-like interannual variability and IPO-like (PDO-like) decadal variability.

3. The reduced-gravity ocean model

The ocean model used in our series of experiments comprises a linear reduced-gravity SWM at 1° resolution configured 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, while north and south artificial boundaries are at 41°N and 41°S. The SWM dynamic model consists of 1½ layers, with the active upper layer of uniform density overlaying a deep, motionless lower layer of larger uniform density. These density layers are separated by an interface called the pycnocline that approximates the ocean’s thermocline. The active upper layer of the SWM is further divided into a shallow frictional layer of constant depth (the upper 50 m) and the lower active layer containing the depth-averaged baroclinic currents. This allows us to incorporate the thermodynamics of Zebiak and Cane (1987) to describe the evolution of SST anomalies in the ocean’s upper surface layer. In Part I we showed that the SWM forced with the CGCM wind stresses could produce Pacific Ocean interannual and interdecadal variability with similar spatially and statistically significant temporal characteristics to that of the CGCM. Hence, it was concluded that at least a portion of the interdecadal VAT and SST variability of the CGCM Pacific Ocean variability must also rely on ocean dynamic processes that are consistent with shallow-water dynamics. In this study the SWM is forced with anomalous (from the long-term monthly mean) zonal component wind stresses from the century time-scale BCM2.2 simulation [hereafter the zonal wind stress control (ZWSC) simulation]. The BCM2.2 monthly zonal wind stress anomalies were interpolated temporally on to the SWM 2-h time step when forcing the simpler model. The use of only the zonal component of the BCM2.2 anomalous wind stress forcing for the forced SWM simulations simplified the development of the hybrid coupled model (see section 4a) and reduced the computational resources required.

To identify the consequences of neglecting the meridional component wind stress forcing, the results of the ZWSC simulation were compared to the corresponding SWM control simulation results (see McGregor et al. 2004; Part I), which incorporated both the zonal and meridional BCM2.2 wind stress components. Here, the raw (unfiltered) thermocline (pycnocline) depth anomalies of the ZWSC simulation are projected onto the first-mode EOF of the unfiltered SWM control simulation results of Part I (seen here in Fig. 1a). The 13-yr low-pass Butterworth-filtered (hereafter simply filtered) thermocline depth anomalies of the ZWSC simulation are projected onto the first-mode EOF of the filtered SWM control simulation results from Part I (seen here in Fig. 1b). A comparison of the unfiltered (filtered) time series produced by this projection with the first-mode unfiltered (filtered) time series of the control simulation reveals a striking similarity (not shown here) that is quantified by the correlation coefficient of 0.992 (0.997), which is significant above the 99% (99%) confidence level. We note that the statistical significance of all correlation coefficients reported in this study take account of serial (auto-) correlation in the series based on the reduced effective number of degrees of freedom outlined by Davis (1976). The very strong and highly significant correlation between the
original EOF time series and the time series produced by this projection confirm that neglecting the meridional component wind stresses in the SWM ZWSC simulation produces a negligible change to the leading unfiltered and filtered EOFs from the SWM results.

4. The equatorial HCM

In Part I we showed that both equatorial and off-equatorial wind stress forcing can modulate the equatorial thermocline displacement on interannual through to interdecadal time scales, and that equatorial wind stress forcing appeared to be the dominant mechanism. However, the question raised here is whether the small equatorial ocean response of the SWM to off-equatorial wind stress forcing is sufficient to initiate a coupled atmosphere–ocean interaction that will significantly increase the original equatorial region variability. In an attempt to address this question we first divide the zonal component BCM2.2 wind stresses into two forcing regions that we define as “equatorial” and “off-equatorial.” Our definition of equatorial region wind stress forcing here includes only ZWSC simulation between 12.5°S and 12.5°N (the wind stress anomalies are tapered to zero linearly from 10° to 15° latitude). The off-equatorial region wind stress forcing includes only wind stresses in the extraequatorial zone (i.e., those latitudes poleward of 12.5° latitude). In this case, the wind stress anomalies are tapered to zero in the equatorial direction from 15° to 10° latitude. We then develop an HCM of the equatorial Pacific region that will allow us to evaluate the importance of off-equatorial wind stress forcing on the coupled equatorial Pacific Ocean variability.

The equatorial region HCM utilizes the dynamic ocean SWM coupled with the Zebiak and Cane (1987) thermodynamic SST equation (as introduced in section 3) and a statistical atmosphere derived from the singular value decomposition (SVD) of the cross covariance between the two BCM2.2 output fields—the equatorial region SST and the overlying zonal wind stresses. This SVD method of the cross-covariance matrix has been described in detail by Bretherton et al. (1992) and Syu et al. (1995). The SVD of the cross-covariance matrix, from two datasets, identifies pairs of spatial patterns that explain as much as possible of the mean-squared temporal covariance between the two fields (Bretherton et al. 1992). Here, the SVD analysis is of the monthly BCM2.2 equatorial region SST anomalies (SSTAs) and zonal wind stress anomalies (both anomalies are from the long-term monthly mean, i.e., the seasonal cycle is removed), giving us a covarying pattern of equatorial SSTA and zonal wind stress ($\sigma^2$) for each calendar month. These covarying patterns will be referred to as the SSTAs SVDs and the $\sigma^2$ SVDs. Considering the short atmospheric adjustment times relative to the adjustment times of the equatorial ocean, we make the assumption that the SVD analysis identifies variations in the overlying zonal wind stresses that are in response to changes in the equatorial SST in the ZWSC simulation.

The spatial structure of the first SVD modes of SSTA
and $\tau^x$ for every other calendar month (February, April, June, August, October, and December) are presented in Fig. 2. Visually, the first SVD mode of SSTA for each month is characteristic of the seasonal evolution of ENSO in the CGCM. The first SVD mode of $\tau^x$ vectors for each month is consistent with the breakdown (amplification) of the zonal Walker circulation with a built-in time-dependent propagation similar to
that of observations (Kirtman 1997). The ENSO-like nature of these dominant first-mode SSTA contours and $\tau^s$ vectors (cf. Fig. 2) further confirms that this statistical approach is both appropriate and useful in representing the coupled variability of the equatorial region. The use of a statistical atmosphere whose properties vary with calendar month allows us to simulate an ENSO that is phase locked to the annual cycle, as is the case for the real-world ENSO (although highly simplified).

When developing a statistical atmosphere there are typically two ways to identify how many modes should be retained. In the first method an empirically determined number of modes is retained beyond which the explained variance of the singular values drops off sharply [e.g., the “scree test” of Cattell (1966)]. In the second approach, modes are retained in order to account for approximately 95% (or some other arbitrary choice) of the total covariability of the system. This second approach has resulted in the inclusion of as few as two modes (e.g., Moore and Kleeman 2001), but more often than not has resulted in the retention of around seven to eight modes (e.g., Zavala-Garay et al. 2003). Here, evaluation of the number of modes retained is constrained by the fact that we couple the statistical atmosphere (based on BCM2.2 output) with SST from the intermediate-complexity ocean SWM. Thus, our statistical estimate of the atmospheric coupling is based on the relevant number of SVD modes of SST that the SWM reproduces well.

To identify the more sophisticated BCM2.2 SST SVD modes of covariability that the simpler SWM SST represents well, we projected the results of the SWM ZWSC simulation (detailed in section 3) onto the first eight SVD modes for each month. The correlation coefficients between the BCM2.2 SST SVD time coefficients and the projected SWM SST time coefficients are evaluated (Table 1). It is clear that we should retain the first-mode SVD for each calendar month because the highly significant correlations (all correlation coefficients are significant at the 99% level) confirm that the SSTs of the SWM zonal control simulation are highly representative of the BCM2.2 first SVD mode SSTs. It is tempting to retain mode 2 in our statistical atmosphere because the correlations are reasonable for each of the 12 months, with each being statistically significant above the 95% confidence level (cf. Table 1). However, there is no consistent spatial pattern similarity between all of the 12 corresponding second BCM2.2 SVD modes (not shown here), indicating that a consistent physical process is not represented in the second SVD of each calendar month. In addition, using the criterion of North et al. (1982), 5 of the 12 calendar month second SVDs can be considered random mixtures of true eigenvectors (cf. Fig. 3). In regards to third and higher-order modes, there is almost no consistent SWM skill for all 12 months (cf. Table 1). Thus, considering this information and the fact that the only drop-off point that is consistent for each calendar month falls after the first SVD mode (Cattell 1966; cf. Fig. 3), we base our statistical atmosphere on the first SVD mode for each calendar month. As discussed earlier, these 12 first modes of the equatorial region zonal wind stresses appear to be representative of variations in the Walker circulation in quadrature (cf. Fig. 2). Hence, our HCM would be expected to give a good representation of equatorial region variability. This expectation is supported in the results from a previous study by Moore and Kleeman (2001) who showed, using a statistical atmosphere based on the first two modes from an SVD analysis (the two modes seemed to represent ENSO in quadrature), that their hybrid coupled model did a good job of reproducing the variability in the more sophisticated coupled dynamical model used in their study.

One of the major advantages of the HCM’s simple statistical atmosphere is that it allows us to identify and represent the coupled components of the BCM2.2 zonal wind stresses as those linearly related to equatorial SST anomalies. Further, by knowing the coupled component of the equatorial region wind stresses, then the noise component, that is, the component that does not adjust to time changes in the equatorial SST, can be obtained by subtracting the coupled component from the original wind stress anomaly field. Here, we estimated both the coupled and the noise components of the zonal equatorial region wind stresses estimated us-

<table>
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<th>Month</th>
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<th>Mode 2</th>
<th>Mode 3</th>
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ing this approach. The standard deviation (SD) of the noise component of the surface wind stress forms a chevron shape that surrounds the low wind stress variability of the eastern tropical Pacific region (Fig. 4b), while the SD of the coupled wind stress component (Fig. 4a) is concentrated in the western/central equatorial Pacific, consistent with the wind stress response to ENSO variability (Dijkstra and Burgers 2002). Power spectra of the coupled and uncoupled (noise) components of the zonal wind stress anomalies at 170°W, shown as an example, reveal that there is a strong peak at the shorter end of the 2–7-yr ENSO window (~0.15–0.5 cpy) in the coupled signal, while there is no clear defining spectral peak in the equatorial region noise (Fig. 5). The realism of the SD spatial plots for both the coupled and noise (uncoupled) components, along with the fact that the estimates of the equatorial region wind stress noise have no prominent spectral peak, suggests that this simple statistical approach provides an appropriate representation of the fundamental coupled variability in the equatorial region.

5. HCM experiments

a. Experiment I: The standard HCM simulation

To examine the seasonal dependence of the HCM to an initial perturbation we carry out a series of six HCM simulations. Each of these simulations is forced with the same randomly selected anomalous wind stress pattern for a 2-week period, and then model coupling is switched on and the model is left to freely evolve. The bimonthly simulations were respectively initiated in January, March, May, July, September, and November. The HCM response for each of the six simulations ex-
hibits damped oscillatory changes in both the thermocline depth anomalies and SST. The level of damping depends on the calendar month in which the HCM wind stress forcing is initiated. The time series of the Niño-3 region (5°N–5°S, 150°–90°W) SSTA from each of the six simulations are shown in Fig. 6. Overall, the amplitudes and damping response in the HCM are consistent with other dynamical models of ENSO (e.g., Zavala-Garay et al. 2003).

b. Experiment II: The ENF HCM experiment

Experiment II, or the equatorial noise forcing (ENF) HCM experiment, is carried out to identify whether the HCM forced with equatorial region noise can produce realistic interannual and interdecadal variability in the tropics. To this end, the HCM is forced with randomized spatial patterns of equatorial region noise (described in section 4) for 90 yr. The order of the estimated wind stress noise was randomized to eliminate any possible effects of unidentified wind stress features that rely on the coupled atmospheric memory. To analyze the interannual variability in the ENF experiment, results from the first EOF (EOF1) of the modeled SST and thermocline depth variability and from the second EOF (EOF2) of the modeled thermocline depth variability, and the variance statistics of the Niño-3 and

![Image](https://via.placeholder.com/150)

**Fig. 4.** Standard deviation of the (a) coupled component and (b) noise (uncoupled) component (N m⁻²) of the zonal wind stress.

**Fig. 5.** Power spectra of the equatorial zonal wind anomalies at 170°W for the coupled and uncoupled (noise) signals.

**Fig. 6.** Niño-3 region SSTA from the six standard HCM simulations in expt 1.
Niño-4 region (5°N–5°S, 160°E–150°W) SSTA variability are analyzed (cf. Table 2). EOF1 of the modeled SSTA and thermocline depth anomalies, respectively, accounts for approximately 51% and 15% of the total variances of these fields (cf. Fig. 7), while EOF2 of the modeled thermocline depth anomalies accounts for approximately 9% of the total field variance (not shown). The tongue of SSTA extending from the eastern equatorial Pacific to the central/western equatorial Pacific in EOF1 of SSTA (cf. Fig. 7a) is distinctly similar to the observed SSTA signature of ENSO. Variations of weighting in EOF1 of the thermocline depth anomalies (cf. Fig. 7b) display an out-of-phase relationship between the eastern tropical Pacific and western tropical Pacific, which is consistent with the delayed-action oscillator ENSO mechanism (Schopf and Suarez 1988; Battisti and Hirst 1989). The HCM also appears to be consistent with the recharge–discharge oscillator ENSO paradigm of Jin (1997), because the spatial pattern of the EOF2 thermocline depth displays large off-equatorial signals that lag EOF1 by approximately 6 months. The ENSO-like nature of this variability is further illustrated by the interannual fluctuation of both EOF1 fields with their accompanying expansion coefficient time series (Figs. 7c,d), correlating at 0.94, significant above the 99% level.

To analyze the interdecadal variability that is evident in the experiment II ENF results, we examine both the variance statistics of the filtered Niño-3 and Niño-4 region SST variability (cf. Table 2) and the results of an EOF analysis on the filtered ENF simulation output. The EOF1 spatial pattern of the filtered SST (Fig. 8a) accounts for approximately 83% of the field’s total filtered variability, and it has a similar pattern to the ENSO-like pattern of its unfiltered first EOF of the ENF experiment SSTA (cf. Fig. 7a). However, it is clear that the equatorial weighting of this filtered EOF is somewhat broader meridionally than its unfiltered counterpart, which is consistent with both BCM2.2-modeled variability (Power and Colman 2006) and observations of tropical Pacific Ocean decadal variability (e.g., Zhang et al. 1997). Further supporting the spatial pattern correspondence with observed tropical Pacific Ocean interdecadal variability is the centrally located strong positive weighting as well as the slow interdecadal-scale oscillation of the accompanying expansion coefficient time series (Fig. 8c). EOF1 of the filtered thermocline depth variability (Fig. 8b) explains approximately 37% of the filtered model simulation variance. The spatial pattern of this mode displays positive weighting in the equatorial region, which is broader and extends farther west than its unfiltered counterpart, somewhat similar to that seen in Fig. 1b. The expansion coefficient time series of this EOF also has a slow in-

<table>
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<th>Niño-3 Filtered</th>
<th>Niño-4 Unfiltered</th>
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Table 2. Variance (°C²) of the unfiltered and filtered SST anomalies produced in the various SWM experiments.

Fig. 7. The first EOF spatial pattern of the unfiltered (a) SSTA and (b) thermocline depth anomalies of the ENF HCM simulation. (c), (d) Their respective accompanying expansion coefficient time series are also shown.
terdecadal-scale oscillation, which correlates with filtered SSTA EOF1 time series with a correlation coefficient of 0.97, significant above the 99% level (Fig. 8d). However, unlike the thermocline depth variability highlighted in results from the unfiltered EOF analysis, which explains the mechanisms of ENSO variability (e.g., the delayed-action oscillator), the thermocline depth variability highlighted here does not appear to characterize any well-established or understood process that can help explain the changes in SST. Overall, based on analyses of the unfiltered and filtered ENF experiment results presented here, it is clear that the HCM forced with equatorial region noise produces realistic interannual ENSO-like variability and interdecadal variability that is not inconsistent with what is known to occur with the IPO.

c. Experiment III: The OffEqF HCM experiment

In Part I, we showed that off-equatorial wind stress forcing can modulate the equatorial thermocline depth on interdecadal time scales, and that the majority of the resulting thermocline depth variability appears to be predictable up to 3 yr in advance. However, the amplitude of the off-equatorially forced equatorial region thermocline depth variance is small when compared with direct equatorial region wind stress forcing. The purpose of experiment III, or the off-equatorial forcing (OffEqF) HCM experiment, is to identify whether the small equatorial ocean response of the SWM to off-equatorial wind stress forcing is sufficient to initiate an atmospheric feedback response that will significantly amplify the original equatorial region variance. To this end, a 90-yr HCM integration is forced with randomized spatial patterns of off-equatorial region wind stress forcing. Hence, in this experiment there is no un-coupled wind stress forcing applied (externally) in the equatorial region (within $\pm 12.5^\circ$ of the equator). An EOF analysis of the unfiltered and filtered OffEqF experiment results reveals spatial patterns that are dominated by signals in the off-equatorial region (not shown here), suggesting that the addition of off-equatorial region wind stress forcing to the HCM has done little to produce important additional equatorial region variability. This is confirmed from analysis of the variance statistics of the equatorial region SSTA (Table 2), where it is seen that the unfiltered OffEqF results produce Niño-3 (Niño-4) region variances that are approximately 42 (21) times smaller than those produced in the ENF HCM simulation.

d. Experiment IV: The DRF HCM experiment

The final HCM experiment, using both the “noise” component of the equatorial region wind stress forcing and the off-equatorial region wind stress forcing, is the dual-region forcing (DRF) HCM experiment. The purpose of this experiment is to identify whether the small portion of off-equatorially forced equatorial region of variability is important in terms of interannual and interdecadal variability in the HCM. Again, the order of the estimated equatorial region wind stress noise was randomized to eliminate any possible effects of unidentified wind stress features that rely on the coupled atmospheric memory. To exclude artificial changes in wind stress curl where the forcing regions overlap (i.e.,
the taper regions from 10º to 15º latitude) the same randomization was also carried out on the off-equatorial region wind stress forcing. Analyses of the unfiltered and filtered DRF experiment EOF results (not shown here) reveal that the addition of off-equatorial region wind stress forcing to the ENF HCM has had little effect on the dominant model behavior (identified by the leading EOFs), because the realistic interannual ENSO-like variability and interdecadal IPO-like variability of the model is still produced. However, in order to assess to role of the off-equatorial wind stress forcing on the equatorial Pacific Ocean the results of this DRF HCM experiment will be directly compared to the results of ENF HCM experiment.

We first present the variance statistics of the DRF HCM experiments’ unfiltered and filtered Niño-3 and Niño-4 region SSTA variability for comparison with those from the ENF HCM experiment (cf. Table 2). Analysis of these statistics reveals that although the addition of OffEqF had little effect on the dominant modes of variability produced in the HCM, it nevertheless increased the unfiltered (filtered) variance in the Niño-3 region by approximately 6% (12%; cf. Table 2). To evaluate the relative impact of the OffEqF on the interdecadal variability in the HCM, we plot the time series of the filtered Niño-3 region SSTA produced by the DRF and ENF experiments (Fig. 9). The strong positive correlation that exists between the filtered ENF Niño-3 region SSTA and the filtered DRF Niño-3 region SSTA (correlation coefficient of 0.96, significant at the 99% confidence level) indicates that the interdecadal SST variability of the HCM equatorial Pacific Ocean is primarily a consequence of the higher-frequency equatorial region variability. However, there are small differences in amplitude between the interdecadal SST variability of the DRF and ENF experiment time series that are a result of the addition of off-equatorial wind stresses to the DRF experiment.

To assess what these differences mean for the interannual variability in the HCM we plot the unfiltered time series of the Niño-3 region SSTA produced by both experiments (Fig. 10). It is clear that both time series are very similar, so much so that they are almost indistinguishable. This is quantified by the correlation coefficient of 0.97, which is statistically significant above the 99% level. However, it is the differences in which we are interested because it is likely that this variability has predictable lead times of up to 3 yr (Part I). We calculate the root-mean-squared difference between the Niño-3 region SSTA from the DRF experiment and the ENF experiment to be ~0.1°C, while the maximum root-squared difference is 0.4°C (almost one standard deviation from the mean). A mean difference of 0.1°C is quite small when compared to the standard deviation of the ENF SSTA in the same region (0.5°C; cf. Table 2). However, as can be seen at various times throughout the simulation (e.g., see Fig. 8 at ~1.5, 2, 21, 25, 35.5, 56, 58, 64, 65, and 76 yr), differences larger than 0.2°C in Niño-3 SST resulting from OffEqF occur. As such, it would appear that a comprehensive understanding and more accurate seasonal–interannual prediction potential could be gained by taking this OffEqF variability into account.

6. Discussion and conclusions

In this study we have attempted to identify important physical mechanisms that underpin the interdecadal vertically averaged temperature (VAT) to a depth of 300 m and SST variations in a century-long CGCM simulation. In Part I, we showed that the important ocean dynamic processes of the CGCM’s interannual and interdecadal VAT and SST variability appear to be mostly captured by a linear first baroclinic mode ocean SWM forced with the CGCM wind stresses. It was shown that both the equatorial region (between ±12.5º latitude) and off-equatorial region wind stress forcing (those latitudes poleward of 12.5º) can modulate the equatorial thermocline displacement on interdecadal time scales, but that equatorial region wind stress forcing was the dominant mechanism (Part I). In the study herein, the small equatorial ocean response of the SWM to off-equatorial wind stress forcing was further investigated to evaluate whether it was sufficient to initiate a coupled atmosphere–ocean response that could significantly amplify the original equatorial region variance. To do this we developed a hybrid coupled model (HCM) of the equatorial Pacific. The role of equatorial region wind stress noise, off-equatorial wind stress forcing (and oceanic Rossby waves), and the combined wind stress forcing from both the equatorial and off-equatorial regions were investigated in a series of HCM experiments. We note here that although we seek to identify the role of off-equatorial wind stress forcing on the equatorial region variability, we do not attempt to address the origin of the wind stress variability. The
HCM comprises a linear reduced-gravity ocean shallower-water model (SWM) with the Zebiak and Cane (1987) thermodynamic SST equation coupled with an equatorial region statistical atmosphere derived from the singular value decomposition (SVD) of the BCM2.2 monthly zonal wind stress anomaly and SSTA output. The statistical atmosphere component of the HCM has a built-in seasonal dependence, which is consistent with observed variations of ENSO. Overall, the HCM displays interannual ENSO-like variability and interdecadal IPO-/PDO-like variability in the tropics that are consistent with observations.

The HCM experiments performed in this study have shown that the coupled tropical Pacific response to both equatorial region wind stress noise and the off-equatorial region wind stress forcing have roles to play in modulating the equatorial region SSTA and the thermocline depth anomalies. As expected from theory and previous experiments, the coupled equatorial response to direct (equatorial region) wind stress forcing dominates the higher-frequency (ENSO time scale) Niño-3 and Niño-4 region SSTA and thermocline depth variations. Here, the unfiltered off-equatorially forced HCM (OffEqF) simulation generates a variance that is approximately 2% (5%) of the size of its equatorial wind stress noise–forced (ENF) equivalent in the Niño-3 (Niño-4) region (cf. Table 2). When considering the lower-frequency (filtered) interdecadal time-scale thermocline depth anomalies in the tropical Pacific, the equatorial region wind stresses also dominate. The filtered off-equatorially forced HCM simulation (OffEqF) produces a variance that is 7% (9%) of the size of its ENF equivalent in the Niño-3 (Niño-4) region (cf. Table 2). These results indicate that even with ocean–atmosphere coupling in the tropical Pacific, the off-equatorial wind stress–forced equatorial region variability is insufficient to initiate an atmospheric feedback that significantly amplifies the original equatorial
Fig. 11. The time series of the first EOF of the filtered SWM control simulation thermocline depth anomalies (solid) and the (13-yr low pass) filtered expansion coefficient time series of the first EOF on the unfiltered SWM control simulation thermocline depth anomalies (dashed).

region variability. These results more conclusively confirm the findings in Part I, and now strongly suggest that tropical Pacific interdecadal variability is predominantly driven by equatorially forced variability.

To identify the relationship between the modeled ENSO and the modeled IPO-like variability we low-pass filter (with a 13-yr cutoff) the EOF time series of the ENSO-like mode from the Part I SWM control simulation (seen here in Fig. 1c), and compare it with the time series of the corresponding IPO-like mode (seen here in Fig. 1d). Analysis of these two time series (see Fig. 11) produces a correlation coefficient of 0.98, which is significant above the 99% confidence level. The statistical similarity between the filtered ENSO-like mode time series and the time series of the IPO-like mode supports the theory of Power et al. (2006), which proposes that the random interdecadal variability of ENSO events caused by stochastic forcing, and the red response to such random changes, may dominate the Pacific Ocean’s interdecadal variability. This suggests that the predictable lead times for the majority of the equatorial Pacific Ocean’s interdecadal SST variability is limited to 1 yr or so (the theoretical predictable lead times for ENSO events). However, we must note that because of the use of a regional domain and the relatively simple 1½-layer SWM ocean we cannot categorically rule out the possible impacts of variations in the meridional overturning circulation on predictability.

Our results also suggest that there might be a small component of equatorial region variability that is predictable on multidecadal time scales. We have shown here that the addition of off-equatorial region wind stress forcing to the ENF HCM, in the dual-region wind stress–forced (DRF) experiment simulation, has increased the unfiltered (filtered) variances in the Niño-3 region by ~6% (12%). In Part I, we showed that the variability driven by off-equatorial region wind stress forcing can be predicted up to 3 yr in advance. Now, considering that the ENF HCM experiment produces Niño-3 region SSTA variations with a standard deviation (SD) of ~0.5°C, the addition of a ~0.1°C temperature variation resulting from off-equatorial wind stress forcing is relatively unimpressive, even if it is predicted 3 yr in advance. However, off-equatorial wind stress forcing can modulate the “background” state of the HCM tropical Pacific Ocean SST by up to ~0.4°C (almost one HCM SD). An SSTA of almost one SD in the equatorial Pacific is not enough to be classified as an ENSO event according to the Australian Bureau of Meteorology. However, the SSTA contribution from OffEqF could very easily constructively interfere with a sub-one SD equatorially forced SSTA of the same sign to produce a significant equatorial region ENSO anomaly. Thus, the relatively small OffEqF contribution could prove to be critical for thresholds of ENSO. As such, we believe that a comprehensive understanding and more accurate seasonal-to-interannual prediction could be gained by taking this OffEqF variability into account.

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