Interannual Tropical Pacific Sea Surface Temperatures and Their Relation to Preceding Sea Level Pressures in the NCAR CCSM2

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ABSTRACT

This paper describes aspects of tropical interannual ocean/atmosphere variability in the NCAR Community Climate System Model Version 2.0 (CCSM2). The CCSM2 tropical Pacific Ocean/atmosphere system exhibits much stronger biennial variability than is observed. However, a canonical correlation analysis technique decomposes the simulated boreal winter tropical Pacific sea surface temperature (SST) variability into two modes, both of which are related to atmospheric variability during the preceding boreal winter. The first mode of ocean/atmosphere variability is related to the strong biennial oscillation in which La Niña–related sea level pressure (SLP) conditions precede El Niño–like SST conditions the following winter. The second mode of variability indicates that boreal winter tropical Pacific SST anomalies can also be initiated by SLP anomalies over the subtropical central and eastern North Pacific 12 months earlier. The evolution of both modes is characterized by recharge/discharge within the equatorial subsurface temperature field. For the first mode of variability, this recharge/discharge produces a lag between the basin-average equatorial Pacific isotherm depth anomalies and the isotherm–slope anomalies, equatorial SSTs, and wind stress fields. Significant anomalies are present up to a year before the boreal winter SLP variations and two years prior to the boreal winter ENSO-like events. For the second canonical factor pattern, the recharge/discharge mechanism is induced concurrent with the boreal winter SLP pattern approximately one year prior to the ENSO-like events, when isotherms initially deepen and change their slope across the basin. A rapid deepening of the isotherms in the eastern equatorial Pacific and a warming of the overlying SST anomalies then occurs during the subsequent 12 months.

1. Introduction

It is recognized that El Niño–Southern Oscillation (ENSO) is one of the primary drivers of local, regional, and global climate variations (e.g., Trenberth et al. 1998; Alexander et al. 2002). In addition, much has been written concerning the forcing mechanisms for El Niño–Southern Oscillation (e.g., Neelin et al. 1998) as well as the evolution of the atmospheric and oceanic components of the ENSO system (Rasmusson and Carpenter 1982; van Loon and Shea 1985; Barnett 1985; Philander 1985; van Loon and Shea 1987; Trenberth and Shea 1987; B. Wang et al. 1999; Chan and Xu 2000; Delcroix et al. 2000; Wang 2001; Larkin and Harrison 2002). Although traditionally defined as a coupled mode of variability of the equatorial Pacific ocean/atmosphere system (Philander 1985), additional precursor fields in the extratropics are related to the initiation of ENSO events (Kidson 1975; Trenberth 1976; Reiter 1978; van Loon and Shea 1985; Barnett 1985; van Loon and Shea 1987; Trenberth and Shea 1987; Lysne et al. 1997; Gu and Philander 1997; Li 1997; Barnett et al. 1999; Pierce et al. 2000; Wang 2001; Vimont et al. 2003b; Anderson 2003, 2004). These findings suggest that extratropical processes may play an important role in forcing equatorial ENSO variability.

As suggested by Jin (1997) and others, the coupled equatorial Pacific ocean/atmosphere system can be characterized either as an oscillating system with intermittent disruptions due to stochastic forcing, or as a forced system that evolves as a decaying oscillator. The
oscillations appear to be part of a recharge/discharge process within the subsurface temperature structure (Jin 1997; Li 1997), which is a generalized form of the delayed oscillator mechanism proposed by Battisti (1988). In their theoretical work investigating this recharge/discharge mechanism, Jin (1997) and Li (1997) highlight the importance of both the basin-average subsurface heat content anomalies and the east–west gradients in these anomalies. There is a negative feedback between the two, modulated by the overlying air–sea interactions. During a mature ENSO event, an anomalous east–west slope in the isotherm depths produces positive depth anomalies over the eastern equatorial Pacific. This slope anomaly initiates a discharge of water via Sverdrup transport, which elevates the basin-average thermocline. Eventually, the negative anomalies associated with the discharge overwhelm the positive anomalies in the eastern Pacific associated with the isotherm slope. The shoaling thermocline produces negative SST anomalies in this region, that is, the start of a La Niña event. The SST field then produces a change in the overlying atmospheric circulations, which result in a weakening and eventual reversal of the isotherm slope anomalies. At that point, Sverdrup recharge of the basin-average depths begins.

While the above mechanism emphasizes the importance of dynamic processes in the deep Tropics, previous results suggest that variations in the subtropical trade wind regime can also precondition the equatorial ocean/atmosphere system to a particular phase of ENSO (Reiter 1978; van Loon and Shea 1985; Weisberg and Wang 1997; Li 1997; Barnett et al. 1999; Pierce et al. 2000; Wang 2001; Vimont et al. 2001, 2003a,b; Anderson 2003, 2004). Studies of this coupling using climate models and reanalysis data suggest that Northern Hemisphere wintertime atmospheric variability initiates subtropical/tropical SST anomalies that persist into the summer. These summertime SST anomalies can induce overlying sea level pressure anomalies and zonal wind stress anomalies that are conducive to initiating and maintaining ENSO variability in the tropical Pacific (Vimont et al. 2003a,b). Other observational studies indicate that the influence on the equatorial ocean may also be related to a direct forcing by the wintertime subtropical variability itself (Anderson 2003, 2004).

In this paper, we examine the tropical and extratropical atmospheric patterns related to variability of the equatorial SSTs in the National Center for Atmospheric Research (NCAR) Community Climate System Model 2.0 (CCSM2). The main focus of this paper is to identify precursor modes of sea level pressure variability that precede large-scale changes in the boreal winter sea surface temperature structure, similar to what has been done using observations (Vimont et al. 2003b; Anderson 2003) and other model simulations (Vimont et al. 2003a). We will then describe and characterize the associated evolution of the SST and underlying ocean fields. This characterization will highlight that equatorial Pacific SST variability within the CCSM2 is related both to internal oscillations of the ocean/atmosphere system as well as an initiation and generation by SLP anomalies over the subtropical central and eastern North Pacific. It will also provide evidence that the recharge/discharge paradigm for tropical Pacific variability is evident in the evolution of both. A separate paper will concentrate on the physical mechanisms by which the modes of preceding SLP variability may influence and/or initiate the evolution of the subsurface/surface temperature structure.

The climate model datasets used in this study are described in section 2. The correlation between seasonal sea surface temperature anomalies and antecedent atmospheric anomalies are discussed in section 3. This section also examines the evolution of the underlying SST and ocean structure. Findings are summarized and discussed in section 4.

2. Model and observed data

The NCAR CCSM2

The NCAR Community Climate System Model Version 2.0 is a state-of-the-art climate system model that is composed of four separate model components: ocean, atmosphere, land, and sea ice (Kiehl and Gent 2004). These component models are unified by a flux coupler. A detailed description of CCSM2 physics can be found online at http://www.cccsm.ucar.edu/models/. We analyze 250 yr of model output at monthly resolution following a 350-yr spin-up period with radiative gas profiles fixed at 1990 levels.

Oceanic and atmospheric fields are analyzed at T42 resolution, approximately corresponding to a grid resolution of 2.8° × 2.8°. To characterize the atmospheric structure, we analyze the sea level pressure fields because of their integral nature in subtropical and tropical air–sea interactions. In addition, previous studies indicate that they are likely to contain precursor information about the development of large-scale SST anomalies (Kidson 1975; Trenberth 1976; van Loon 1984; van Loon and Shea 1985; Barnett 1985; van Loon and Shea 1987; Trenberth and Shea 1987; C. Wang et al. 1999; Chan and Xu 2000; Wang 2001; Larkin and Harrison 2002; Vimont et al. 2003a; Anderson 2003). We will also examine the time evolution of the ocean model sea
surface temperature, 20°C isotherm depths, and surface wind stress.

We have removed the climatological seasonal cycle and have detrended all data prior to analysis. The low-frequency (i.e., multidecadal and longer) variance within the various datasets is reduced by removing the trends in the data (e.g., as quantified in Trenberth and Paolino 1981). This step ensures that the algorithms and statistics are not capturing modes of long-term variability and covariability. Instead, we are interested in looking at interannual variations, primarily related to changes in large-scale fields on intradecadal time scales. To arrive at confidence limits for correlation/regression calculations, we estimate degrees of freedom by using twice the $e$-folding time of the autoregression curve for the regressor time series (see below). In all cases, the decorrelation time scale appears to be about one to one and a half years (not shown). We therefore assume each 3-yr period is independent, suggesting approximately 83 effective degrees of freedom. The associated 95% confidence limit applies for correlation values greater than $|r| = 0.20$. As such, the minimum contour for all correlation maps is set to 0.20. However, for clarity, these confidence limits are not shown on the regression maps because they are met by all relevant anomalies presented here.

3. Results

a. Statistically coupled modes of variability

A brief glance at ENSO behavior in the model is given by the spectral characteristics of the Niño-3.4 index, defined as the area-average equatorial SST anomalies from 5°S–5°N, 170°–120°W. Figure 1 shows the CCSM2 and observed Niño-3.4 spectra. Model variance is peaked at higher frequencies (~2 yr) and is narrower than the observed spectral peak. Both model and observed peaks are statistically significant relative to a red noise null hypothesis (not shown). The strong biennial peak in the CCSM2 may indicate that the model ENSO is unstable (D. Vimont 2004, personal communication), as opposed to the more stable and stochastically forced behavior in observations (e.g., Vimont et al. 2003b). A lag-correlation analysis of sea surface height verifies the strong biennial behavior of ENSO in the CCSM2 (not shown). Although the model ENSO appears to be unstable, here we examine the joint role oscillatory behavior and possible stochastic forcing play in generating equatorial Pacific variability within the simulated system.

Both the observed and simulated ENSO system shows seasonal variations in their intensity. The December–February (DJF) period has the strongest simu-

![Fig. 1. Power spectrum of Niño-3.4 variance from observations and from the CCSM2. Variance is displayed as the product of power and frequency, and was calculated using data of monthly resolution. The observed record length was used in the calculation of both spectra.](image-url)
EOFs for each 3-month period separately. The analysis starts with the January–March period two years (23 months) prior to the SST anomalies, and proceeds through the DJF period one year (12 months) after the SST anomalies. The CCA technique attempts to produce an orthogonal set of canonical factor (CF) time series that isolate the highest correlated modes of variability within a subset of the SLP and SST EOF time series (Barnett and Preisendorfer 1987). We limit the subset of EOFs used for the CCA to the first 12 for the SSTs and the first 14 for the sea level pressure fields. Although this number is somewhat arbitrary, it is advisable to include enough EOFs so that a large fraction of the total variability is incorporated into the algorithm (Feddersen et al. 1999). In this case, the subset of 12 SST EOFs captures 62% of the total DJF SST variance while the subset of 14 sea level pressure EOFs explains between 65% and 76% of the respective 3-month mean variance.

The CCA algorithm only identifies linearly related signals within the SLP/SST fields. It is known that the evolution of the observed ENSO system contains asymmetrical patterns associated with El Niño and La Niña events (e.g., Hoerling et al. 1997; Larkin and Harrison 2002). Preliminary investigations suggest the simulated system also has asymmetrical features with regard to the ENSO evolution. Here we only focus on the linear patterns as a first step toward understanding the ENSO system within the CCSM.

Overall, the CCA indicates that the predominant canonical factor sea surface temperature pattern that arises from correlation with preceding SLP anomalies involves SST anomalies in the tropical Pacific (not shown). However, results also indicate that more than one SLP pattern during the same 3-month period may be related to the development of ENSO-like DJF SST variations. Hence, we perform a multivariate regression of the first four canonical factor SLP time series against the simulated DJF Niño-3.4 index for every lead/lag period discussed above. This regression provides an indication of how well the combined SLP patterns correlate with the Niño-3.4 time series at a given lead/lag. We also perform a similar regression for the first four DJF SST time series to determine whether the corresponding SST patterns fully capture the Niño-3.4-related variability. Results are shown in Fig. 3.

During the 24 months prior to the Niño-3.4 index the variance explained by the preceding SLP time series generally increases as the lead time decreases. However, boreal winter SLP fields 11 months before the

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**Fig. 2.** Time series of normalized, seasonal mean Niño-3.4 index for DJF from the CCSM2. Niño-3.4 index calculated as the area-average sea surface temperature anomalies from 5°S–5°N, 120°–170°W.

**Fig. 3.** Multivariate correlation of the first four canonical factor time series for SLP (circles) and corresponding DJF SST (crosses) with the DJF Niño-3.4 index concurrent with the SST fields. The x axis is the first month of the 3-month average of SLP beginning with January, two calendar years prior to the Niño-3.4 index. Solid vertical line represents period concurrent with corresponding DJF Niño-3.4 index and SSTs. Canonical factor time series produced by performing a canonical correlation analysis of the 3-month average SLP–anomaly field with the DJF SST anomaly field concurrent with the Niño-3.4 index. See text for details.
DJF Niño-3.4 [JFM(−1)] are better correlated with subsequent ENSO development than are the following boreal spring values, similar to results obtained from observations (Barnett 1985; Trenberth and Shea 1987; Barnston and Ropelewski 1992; Vimont et al. 2003b). This JFM(−1) period is also when the multivariate SST correlation values first plateau, suggesting that the Niño-3.4 behavior is fully explained by the first four canonical factor patterns of DJF SSTs at a 1-yr lag with antecedent SLP anomalies. These results are the same when including three or more canonical factor patterns in the multivariate regression (not shown).

Figure 4 shows the spatial structures of the first four DJF SST canonical factors related to JFM(−1) SLP variations during the previous year (as identified above). The variance explained by each of these modes is given in Table 1. Both the first mode and second mode of variability [canonical factor 1, 2 (CF1, 2)] have large SST anomalies in the equatorial Pacific. However, CF2 is also related to SST anomalies throughout the subtropical central Pacific in both hemispheres. Neither the third nor the fourth mode shows any large equatorial Pacific SST anomalies. Hence, the high multivariate correlation between the Niño-3.4 index and the CF SST time series as derived from a 1-yr lag with antecedent JFM(−1) SLP anomalies (seen in Fig. 3) arises from the first two modes of variability (e.g., $r = 0.96$ for the multivariate regression of the Niño-3.4 index against the SST CF1–2 time series alone).

Figure 5 shows the corresponding JFM(−1) SLP patterns, again for the first four canonical factors. The variance explained by each of these modes is given in Table 1. The first mode of SLP variability has a significant dipole anomaly in the equatorial/subtropical region and additional anomalies extending across the extratropical regions of both hemispheres. This SLP map is similar to observed maps associated with ENSO variability (see Fig. 1 from Trenberth and Shea 1987) and represents the simulated tropical/extratropical signature of the (positive) Southern Oscillation and its teleconnections. This mode appears to capture a strong biennial ocean/atmosphere shift in which La Niña–related SLP conditions precede El Niño–like SST conditions the following winter. The second canonical factor of JFM(−1) SLP variability contains anomalies over the North and South Pacific, as well as the tropical Indian Ocean and the extratropics of the Northern Hemisphere. Unlike the first mode, no significant anomalies exist over the tropical Pacific. While both the CF1 and CF2 SLP maps precede warm equatorial Pacific SSTs the following year (Figs. 4a,b), the CF1 and CF2 SLP anomalies have opposite signs over the subtropics and midlatitudes of the Pacific and Atlantic basins. In addition, although both the 3rd and 4th canonical factor SLP patterns show significant anomalies across the globe, because the corresponding SST patterns are unrelated to variability in the equatorial Pacific they will not be considered further.

In summary, it appears that the simulated ocean/atmosphere system is characterized by two modes of boreal winter SLP variations that result in the estab-

![Fig. 4](image)

**Table 1.** Fraction of variance explained by the first 4 canonical factors of DJF SSTs as regressed against JFM SLP from the previous year.

<table>
<thead>
<tr>
<th>Canonical factor</th>
<th>SST</th>
<th>SLP</th>
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<tbody>
<tr>
<td>1</td>
<td>0.078</td>
<td>0.118</td>
</tr>
<tr>
<td>2</td>
<td>0.169</td>
<td>0.061</td>
</tr>
<tr>
<td>3</td>
<td>0.032</td>
<td>0.090</td>
</tr>
<tr>
<td>4</td>
<td>0.033</td>
<td>0.064</td>
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lishment of wintertime SST anomalies in the equatorial Pacific basin the following year. One is related to an apparent transition from a negative to positive state within the tropical ocean/atmosphere ENSO system. The other is related to preceding SLP variability in the subtropical Pacific. The frequency spectra of the first and second canonical factor time series of SLP and SST are calculated (Fig. 6). Because we use once-yearly (JFM and DJF averages, respectively) data in the calculation of these spectra, and some spectral smoothing, these spectra only give a broad comparison of the temporal characteristics between the two modes. Overall, the first canonical factor of simulated SLP and SST variability are weighted toward biennial frequencies, consistent with the behavior of Figs. 4a, 5a (because a period of two years corresponds to the Nyquist frequency in this spectrum, the two-year peak cannot be resolved). In contrast, the CF2 spectrum generally has a broader distribution of variance with greater power at lower frequencies.

Figure 7 shows the time series for the SST canonical factor maps. The prominent biennial oscillations seen in the full Niño-3.4 index (Fig. 2) are represented by CF1, which has extended periods of oscillatory behavior. In comparison, CF2 is less periodic with longer intervals between events. It too contains occasional systematic oscillations between events, however. The correlations of the two time series with the “parent” Niño-3.4 index are similar ($r = 0.62$ and 0.74). In addition, as mentioned the multivariate correlation of the two with the Niño-3.4 index is $r = 0.96$, indicating together they almost completely explain model ENSO-like variability.

b. Evolution of the sea surface temperature fields

To investigate how the modes of preceding JFM SLP variability relate to the establishment of DJF SST anomalies in the equatorial Pacific, monthly anomalies of simulated SSTs are regressed against the JFM CF1 SLP time series (Fig. 8). Panels show results for alternating months beginning with January 12 months prior to the CF1 SLP time series. Figure 9 shows the SST anomalies as regressed against the CF2 SLP time series. For the first canonical factor, SSTs during boreal winter preceding the SLP anomalies are positive ($>0.4$ K) across the equatorial Pacific. These anomalies are replaced by negative equatorial Pacific SST anomalies during late boreal summer. The negative anomalies persist into boreal winter concurrent with the SLP time series and are indicative of a La Niña–like SST state as discussed earlier. The negative anomalies again weaken and are replaced in late boreal summer with positive anomalies over the eastern and central Pacific. These positive anomalies persist into boreal winter concurrent with the SLP time series and are indicative of a La Niña–like SST state as discussed earlier. The negative anomalies again weaken and are replaced in late boreal summer with positive anomalies over the eastern and central Pacific. These positive anomalies persist through the end of the year, resulting in the positive SST anomalies found in the canonical factor map of DJF SSTs (Fig. 4a). This SST oscillation confirms the strong biennial frequency of the SST time series seen in Figs. 6, 7.

In contrast, the CF2-related SSTs do not exhibit any large-scale anomalies during the 12 months preceding the SLP anomalies (Fig. 9). In addition, no significant equatorial warm water anomalies are present during
the concurrent January period, although the central and eastern subtropical North Pacific begins to warm during this time. In February and March, the subtropical warming intensifies while cooling occurs over the extratropical central North Pacific. The onset of warm anomalies in the equatorial waveguide occurs during April (not shown). This warming intensifies through the boreal summer [July(−1)] and weakens slightly in boreal fall. It then intensifies and expands in late fall and into boreal winter, resulting in the positive SST anomalies found in the canonical factor map of DJF SSTs (Fig. 4b). This evolution suggests that the boreal fall/winter ENSO events following the JFM CF2 SLP variability are not simply due to the internal variability of the atmosphere–ocean system, but may instead be initiated by subtropical forcing (Vimont et al. 2001, 2003a; Anderson 2003).

c. Evolution of the subsurface temperature fields

The underlying ocean structure plays a key role in the evolution and onset of the equatorial SST fields. Figure 10 shows the monthly anomalies of simulated 20°C isotherm depths (as diagnosed from the ocean temperature profiles) regressed against the CF1 time series of JFM SLP. During boreal winter prior to the JFM SLP time period, a weak Pacific dipole contains negative depth anomalies centered on Indonesia and positive depth anomalies over the eastern equatorial Pacific. There are also positive off-equatorial depth anomalies north of the equator. With time, the negative anomalies extend across the Pacific while the positive off-equatorial anomalies shift westward. During the following boreal winter (concurrent with the JFM SLP time series) the initial dipole between Indonesia and the equatorial eastern Pacific is reversed. As during the previous year, the anomalies centered on Indonesia extend eastward across the Pacific while the off-equatorial anomalies over the eastern Pacific shift westward. Consistent with the overlying SST signature, it appears the subsurface temperature signature associated with CF1 is strongly oscillatory, and is related to an equatorially trapped adjustment of the surface/subsurface temperature structure.

Figure 11 shows the 20°C isotherm depths regressed against the CF2 time series of JFM SLP. These maps suggest that while there is no surface manifestation of warm water anomalies preceding or concurrent with the overlying atmospheric variability (Figs. 9a–f), positive isotherm depth anomalies (indicative of warmer subsurface temperatures) surround Indonesia during September(−2). In January concurrent with the SLP time series, these depth anomalies begin to spread throughout the equatorial Pacific. The corresponding surface signature however does not appear until around April–May (Fig. 9i). By late boreal summer and boreal fall following the JFM SLP variations, the 20°C isotherm depths show a dipole structure similar to that in the CF1-related maps, consistent with the similarity in the overlying SST patterns.

These results suggest that the development of mature ENSO events associated with antecedent boreal winter SLP variability may be predicated upon the establishment of subsurface temperature anomalies across the equatorial Pacific basin, in agreement with previous findings (Jin 1997; Li 1997; Meinen and McPhaden 2000, 2001; McPhaden 2003). These subsurface anoma-
lies may in turn be related to the recharge/discharge paradigm discussed in the introduction. To investigate this hypothesis, we calculate the basin-average 20°C isotherm depths (averaged over 4°S–4°N, 140°E–120°W). We also calculate the basin-scale slope of the equatorial 20°C isotherm depths by first computing the slope of the best-fit line for the isotherm depth anomalies from 140°E–120°W. We then take the difference between the height of the slope anomaly at each end point and divide by 2 to arrive at an effective eastern equatorial Pacific depth change associated with the slope anomaly.

The evolutions of the slope and basin-average depth anomalies are regressed against the first two canonical factor time series of JFM SLP (Fig. 12). The y axis is oriented such that positive anomalies represent a deepening of the isotherms in the eastern Pacific. The CF1 pattern is related to significant preceding anomalies in both the mean depths and basin slopes. These oscillate with approximately a 24-month period, which is shorter than that from observations or theoretical considerations (e.g., Li 1997; Meinen and McPhaden 2001). However, the lag between negative (positive) basin depth anomalies during April(–2) [April(–1)] and negative [positive] slope anomalies during January(–1) [January (0)] is 8–10 months, in agreement with the linear recharge/discharge model (Li 1997) and observations (Meinen and McPhaden 2000).

The basin-average depths and slopes related to CF2 do not contain significant precursor anomalies prior to the JFM SLP time period. Basin-average depth anomalies initially increase during January while the slope of the isotherms across the basin decrease (equivalent to deeper eastern equatorial Pacific isotherms). The magnitude of the initial isotherm slope anomalies continues to increase through the next January. However the basin-average depth anomalies start to decrease around April(–1)/May(–1) and continue to decrease through the following February/March, consistent with Sverdrup discharge. As with the CF1 pattern, the phase lag between the maximum basin-average depth anomaly and the maximum slope anomaly is approximately 8–10 months [April(–1) to January(0)].

A strong similarity exists between the evolution of the CF1-related subsurface anomalies during the period from January(–2) to January(0) and the evolution of the CF2-related subsurface anomalies (of the opposite sign) during the period from January(–1) to January(+1). This similarity suggests that the subsurface/surface ocean anomalies following CF2-related SLP
variability may continue to oscillate and project onto the CF1 pattern. Indeed, the one-year lead correlation between the CF2 SLP time series with the subsequent CF1 SLP time series is $r = 0.44$. However, the one-year lead correlation between the CF1 SLP time series and the subsequent CF2 SLP time series is 0.08, indicating as before that the CF2 SLP time series is not simply part of a sustained oscillation within the simulated ocean/atmosphere system.

To see how the subsurface evolution relates to the

Fig. 8. Regression of monthly mean sea surface temperature anomalies against the first canonical factor time series for JFM SLP. Contour interval is 0.15 K; minimum contour is -0.15 K. Only alternating months are shown (a) starting from January 12 months before the JFM SLP time series, and (l) continuing through November 11 months after the JFM SLP time series; numbers correspond to calendar year before DJF Niño-3.4 SST anomalies. Positive values are shaded.

Fig. 9. As in Fig. 8 except for the second canonical factor.
overlying sea surface temperature and zonal wind stress anomalies, we follow the methods of Meinen and McPhaden (2000) and Li (1997). Shown are the time evolution of the simulated monthly mean, basin-average 20°C isotherm depths anomalies (averaged over 4°S–4°N and 140°E–120°W), central/eastern equatorial SST anomalies in the Niño-3.4 region (averaged from 4°S–4°N and 170°–120°W), and basin-average 3-month mean equatorial wind stress anomalies (averaged from 2.5°S–2.5°N and 140°E–120°W). All fields are regressed against the JFM CF1 and CF2 time series (Fig. 13). Anomalies have been normalized by the stan-

Fig. 10. Regression of monthly mean 20°C isotherm–depth anomalies against first canonical factor time series for JFM SLP. Only alternating months are shown, (a) starting from January 12 months before the JFM SLP time series and (l) continuing through November 11 months after the JFM SLP time series; numbers correspond to calendar year before DJF Niño-3.4 SST anomalies. Contour interval is 200 cm; minimum contour is ±200 cm. Positive values are shaded.

Fig. 11. As in Fig. 10 except for the second canonical factor.
dard deviation of the 36-month evolution for both the CF1 and CF2 time series. In addition, the zonal mean depths are plotted with positive values oriented in the same direction as positive anomalies in the other two fields.

As seen in observations (Li 1997; Meinen and McPhaden 2000), the SST anomalies lag the zonal-wind stress anomalies by only about 1–2 months. For CF1, the maxima in the basin-average depth anomalies lead the SST-anomaly maxima by 8–10 months, as seen in observations (Meinen and McPhaden 2000) and theoretical considerations (Li 1997). However the SST anomalies also show a relatively fast transition period during boreal summer following the March–April maxima in the isotherm depths [approximately 5–6 months; February(−1) to August(−1)]. For the CF2 time series, there are no significant basin-average depth, equatorial SST, or wind stress anomalies prior to the JFM(−1) period. In JFM(−1), there is an initial increase in the overall isotherm depths (as well as an in the isotherm slope anomalies seen in Fig. 12), followed approximately two to three months later [March/April(−1)] by positive wind stress and SST anomalies. These SST and wind stress anomalies intensify through July(−1), reaching their plateau value approximately 1–2 months prior to those associated with the CF1 time series, and persist until the following boreal winter. As with the CF1 pattern, both the SST and wind stress anomalies subside approximately 8–10 months after the weakening and reversal of the basin-average depth anomalies [April(−1) to December(−1)/January(0)].

4. Conclusions and discussion

a. Conclusions

The relationship between changes in antecedent sea level pressure patterns and interannual variability in sea surface temperature anomalies is investigated using 250 yr of data from the CCSM2 coupled ocean–atmosphere model. Our interest here is in identifying the relationship between these sea level pressure patterns and the characteristics and evolution of the simulated ENSO behavior in the model. A mechanistic study of the role the SLP and associated wind fields
play in initiating and/or sustaining ENSO will be presented in a subsequent paper.

The simulated ENSO system has a much stronger biennial frequency than the observed system. However, like the observed system, it shows strongest amplitude in boreal winter (DJF). Performing a canonical correlation analysis between the DJF SST fields and preceding global-scale SLP anomalies indicates the overall variance in the equatorial Pacific SST variability increases as the lead-length decreases. However, it is found that the preceding boreal winter (DJF/JFM) SLP fields explain relatively more variance in the subsequent boreal winter SSTs than do the following boreal spring SLP fields, as observed.

Investigation of the sea level pressure anomalies during this preceding JFM period indicates that two of the modes of SLP variability are related to equatorial Pacific SST variability the following year. The first mode of ocean/atmosphere variability [canonical factor 1 (CF1)] is associated with a strong biennial oscillation in the tropical Pacific Ocean in which La Niña-like conditions during boreal winter evolve into El Niño-like conditions the following year. In comparison, the second mode of variability (CF2) is related to sea level pressure anomalies in the subtropical North and South Pacific that occur 12 months earlier, similar to anomalous patterns in the observed data (Vimont et al. 2003b; Anderson 2003). As opposed to the biennial oscillatory behavior in the first mode, the CF2 of SLP variability has a broader frequency spectrum and is not associated with preceding SST, subsurface temperature, or wind stress anomalies in the tropical Pacific. However, the CF2 SLP field does appear to be related to subsurface temperatures surrounding Indonesia 4–5 months earlier.

![Fig. 13. (a) Regression of simulated monthly mean basin-average 20° isotherm depths (solid line), eastern Pacific SST anomalies (dashed line), and 3-month mean basin-average zonal wind stress anomalies (dash-dot line) taken against the first canonical factor time series for JFM SLP. Twenty-degree isotherm depths averaged from 4°S–4°N and 140°E–120°W; SST anomalies averaged from 4°S–4°N and 170°–120°W; zonal-wind stress anomalies averaged from 2.5°S–2.5°N and 140°E–120°W. All values normalized by standard deviation of both the CF1 and CF2 36-month regression values in order to compare relative anomalies for the two patterns; normalization values are 410 cm, 0.37 K, and 0.002 N m⁻² for the depths, SSTs, and wind stress values, respectively. Thick vertical line represents period concurrent with JFM SLP time series; dashed vertical line represents period concurrent with corresponding DJF SST time series. (b) Same as (a) except for regression against the second canonical factor time series.](image-url)
We also show here that the recharge/discharge mechanism plays an important role in the SST evolution related to the two modes of SLP variability. For the CF1, this recharge/discharge behavior is present up to two years prior to mature ENSO events and is captured by a lag between the basin-average equatorial Pacific isotherm depth anomalies and the isotherm–slope, SST and wind stress anomalies. The simulated phase lag between the basin-average depth anomalies and these latter fields (approximately 8–10 months) is similar to observed, although the period is far too short. This high-frequency oscillation appears to arise not because of the lag between the basin-average depth and slope anomalies, but rather the rate at which they transition from maxima to minima. The simulated transition period for CF1 is approximately 12 months while in observations and linear models the transition period is closer to 18 months.

Similar to the CF1, the surface and subsurface temperature structures for the CF2 show a characteristic lag between the basin-average equatorial Pacific isotherm depth anomalies and the isotherm–slope, SST, and wind stress anomalies. However, for the CF2, this recharge/discharge mechanism is apparently induced concurrent with the JFM SLP pattern itself and involves an initial deepening of the basin-average isotherm depths and a change in the isotherm slopes across the basin. These produce a rapid deepening of the isotherms in the eastern equatorial Pacific and warming of the overlying SST anomalies. The basin-average deepening reverses fairly quickly due to Sverdrup transport from the equatorial region, while both the slope and SST anomalies continue to increase. It is important to note that the sign of the initial isotherm slope anomalies are in agreement with the wind stress patterns associated with the overlying SLP anomalies. However, the initial basin-average isotherm deepening is opposite to that associated with Sverdrup discharge, which starts the following boreal spring. Hence, the initiation of basin-average isotherm depth anomalies by the CF2 SLP pattern may be related to a different process, possibly Ekman pumping.

b. Discussion

Here we briefly discuss the relation of the two modes of boreal winter equatorial Pacific sea surface temperature variability in the CCSM with those found in the observations; this discussion is based on preliminary research that will be presented in a subsequent paper but is relevant to the results presented above. Overall, it appears the first mode of ocean/atmosphere ENSO variability in the CCSM is related to a strong biennial oscillation in equatorial SSTs, a shorter time scale than observed. Li (1997) has performed sensitivity studies relating the period of oscillations to the base state of the equatorial Pacific climate system. He found that the period of oscillations generally decreases with increasing climatological vertical velocities and temperature gradients across the thermocline. The period is also sensitive to the climatological basin-average thermocline depth. Increasing the mean depth reduces the period of oscillations (Jin 1997). The high-frequency oscillations within the simulated system could arise from improper simulation of the ocean base state in the model.

Sensitivity tests with the National Oceanic and Atmospheric Administration/Geophysical Fluid Dynamics Laboratory (NOAA/GFDL) model indicate that biennial oscillations in the coupled tropical ocean/atmosphere system also arise from the improper specification of atmospheric convective transport of horizontal momentum. Neglecting convective transport of horizontal momentum results in wind stress anomalies that are too strong and too confined to the equator (A. T. Wittenberg 2004, personal communication). A realistic parameterization for momentum transport can spread momentum meridionally and decrease its magnitude, resulting in weaker ocean forcing and a longer oscillation period. Figure 14 shows the three-month mean JFM wind stress anomalies correlated with the
first canonical factor time series of JFM SLP (which is representative of a strong La Niña–like state). Note the very narrow structure associated with both the equatorial and off-equatorial wind stress anomalies. As such, the CCSM2 may produce high-frequency oscillatory behavior because it does not explicitly transport momentum through convective processes (Zhang and McFarlane 1995).

In contrast to the CF1, the CF2 better captures the spatial and temporal characteristics found in the observed ENSO system. For instance, a similar analysis using the National Centers for Environmental Prediction (NCEP)/NCAR reanalysis data has been described in detail in Anderson (2003). Results indicate the CF1 pattern from observations, reflective of a subtropical influence upon the equatorial waveguide (e.g., Anderson 2003, Vimont et al. 2003b), most closely matches the CF2 pattern from the model (Fig. 5b). The simulated CCSM2 CF2 SLP patterns over the Pacific are also in general agreement with modes of SLP variability in the Commonwealth Scientific and Industrial Research Organisation (CSIRO) model (both with and without SST coupling) that lead the tropical ENSO system by a year (Vimont et al. 2003a). Previous investigations of these model and observational results suggest that the subtropical SLP variability can influence the evolution of the ENSO system by establishing underlying subtropical SST anomalies. These ocean anomalies persist into the following summer and initiate low-level subtropical atmospheric circulations that force the subsurface Kelvin wave structure in the equatorial Pacific (Vimont et al. 2003a, b). Results presented here suggest that boreal winter SLP anomalies may also be related to changes in the underlying equatorial ocean structure during the concurrent wintertime period, as well as to changes during the following summer. Preliminary investigations of the observed system during this preceding boreal winter period indicate that SLP-related changes in the slope of the isotherms, in combination with preexisting depth anomalies, produce a deepening of the isotherms and a subsequent warming of SSTs over the eastern equatorial Pacific, as seen in the simulation. As with the simulated system, observed basin-average anomalies show evolution consistent with Sverdrup-driven discharge, which again is offset by enhanced steepening of the isotherm slope anomaly through the course of the year. A rapid return of the isotherm slopes and SST anomalies to climatological values follows the boreal winter peak SST anomalies.

At the same time, the evolution of the simulated basin-average depths, slope anomalies, and eastern equatorial Pacific SSTs also show significant modulation through boreal summer and into early fall. These variations suggest additional modification of the ENSO evolution by the seasonal footprinting mechanism as described by Vimont et al. (2001). For instance, the evolution of the subsurface structure, and overlying ENSO system, may be enhanced if it is sustained by wind forcing during the boreal summertime period. If this forcing is absent, the evolution may be hampered, resulting in a weak or neutral ENSO period despite the presence of subtropical forcing during the preceding boreal wintertime. The interaction of these two mechanisms and whether either is necessary and/or sufficient for the onset and maintenance of ENSO variability remains an open question.

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