Diagnostics performed with twentieth-century (1861–2000) ensemble integrations of the Geophysical Fluid Dynamics Laboratory Climate Model, version 2.1 (CM2.1) suggest that, during the developing phase, El Niño events that co-occur with the Indian Ocean Dipole Zonal Mode (IODZM; class 1) are stronger than those without (class 2). Also, during class 1 events coherent sea surface temperature (SST) anomalies develop in the Indonesian seas that closely follow the life cycle of IODZM. This study investigates the effect of these regional SST anomalies (equatorial Indian Ocean and Indonesian seas) on the amplitude of the developing El Niño.

An examination of class 1 minus class 2 composites suggests two conditions that could lead to a strong El Niño in class 1 events: (i) during January, ocean–atmosphere conditions internal to the equatorial Pacific are favorable for the development of a stronger El Niño and (ii) during May–June, coinciding with the development of regional SST anomalies, an abrupt increase in westerly wind anomalies is noticeable over the equatorial western Pacific with a subsequent increase in thermocline and SST anomalies over the eastern equatorial Pacific. This paper posits the hypothesis that, under favorable conditions in the equatorial Pacific, regional SST anomalies may enable the development of a stronger El Niño.

Owing to a wealth of feedbacks in CM2.1, solutions from a linear atmosphere model forced with May–June anomalous precipitation and anomalous SST from selected areas over the equatorial Indo-Pacific are examined. Consistent with our earlier study, the net Kelvin wave response to contrasting tropical Indian Ocean heating anomalies cancels over the equatorial western Pacific. In contrast, Indonesian seas SST anomalies account for about 60%–80% of the westerly wind anomalies over the equatorial western Pacific and also induce anomalous precipitation over the equatorial central Pacific. It is argued that the feedback between the precipitation and circulation anomalies results in an abrupt increase in zonal wind anomalies over the equatorial western Pacific.

Encouraged by these results, the authors further examined the processes that cause cold SST anomalies over the Indonesian seas using an ocean model. Sensitivity experiments suggest that local wind anomalies, through stronger surface heat loss and evaporation, and subsurface upwelling are the primary causes. The present results imply that in coupled models, a proper representation of regional air–sea interactions over the equatorial Indo-Pacific warm pool may be important to understand and predict the amplitude of El Niño.
1. Introduction

a. Background

Theoretical and modeling studies have demonstrated that the El Niño–Southern Oscillation (ENSO) owes its existence to the unstable ocean–atmosphere interactions in the tropical Pacific (Battisti 1988; Neelin et al. 1998). It is well recognized that ENSO influences global climate anomalies and its successful prediction has important implications for society worldwide. Based on the delayed oscillator theory, ENSO is a low-frequency basinwide mode of oscillation that is perfectly predictable (Latif et al. 1998). Many factors, however, limit its predictability, including nonlinear interactions with the annual cycle (Jin et al. 1994), mean state changes at decadal time scales (Kleeman et al. 1996), and stochastic subseasonal forcing (Kessler 2002).

Modeling studies demonstrate that, during strong El Niño years, tropical and extratropical circulation anomalies are highly predictable (Shukla 1998; Kumar et al. 2005). However, the present-day state-of-the-art coupled models are still poor in forecasting the amplitude of El Niño (Landsea and Knaff 2000; McPhaden 2008). Two necessary but not sufficient conditions for the growth of strong El Niño events are (i) wave-induced dynamical processes (e.g., Battisti 1988) and (ii) mean conditions such as strong upwelling and a shallow thermocline during May through September over the eastern equatorial Pacific that favor instability growth (Battisti 1988; Neelin et al. 1998; Wittenberg et al. 2006). Regarding the first condition, a large-scale coherent oceanic response requires sustained (30–90 days) zonal wind anomalies with a long fetch over the equatorial western Pacific (Kessler et al. 1995; Lengaigne et al. 2004). Therefore, sources of sustained zonal wind anomalies with sufficient fetch to allow the necessary conditions to act in concert must be identified and properly represented in coupled models in order to achieve better prediction of strong El Niño events.

Such ENSO characteristics as frequency and amplitude are known to vary at decadal and longer time scales (Wittenberg et al. 2006; Neale et al. 2008). Of particular interest here is the amplitude of El Niño. Based on the observed sea surface temperature (SST), El Niños that occurred during 1976–2000 were stronger than those that occurred during 1950–75 (Wallace et al. 1998). Annamalai et al. (2005, hereafter AXMM05) hypothesized that the contrasting heating patterns over the TIO simulated by the model and the sample size from observations was too small to attest statistical significance. Moreover, the focus was only on the effect of boreal fall (September–November) TIO SST anomalies on El Niño. To overcome these shortcomings, we will use in the present diagnostic research multicentury ensemble simulations conducted with the latest version of the coupled model developed at the Geophysical Fluid Dynamics Laboratory (GFDL Climate Model, version 2.1, hereafter CM2.1). CM2.1 captures the salient features of the tropical Pacific climate and ENSO (Wittenberg et al. 2006) and also the TIO climate and IODZM (Song et al. 2007). Therefore, long integrations with CM2.1 allow for a detailed and comprehensive analysis of the IODZM–ENSO linkages.

Figure 1 shows the temporal evolution of SST, zonal wind at 850 hPa, and precipitation averaged over regions relevant to ENSO and IODZM. The results are based on composites constructed from CM2.1 integrations (section 2 provides details of CM2.1 integrations and our data processing method). Throughout the manuscript, class 1 events refer to those instances when both strong (>1.0 standard deviation in SST anomalies) El Niño and IODZM occurred, and class 2 events refer to years when only a strong El Niño occurred. Year (0) refers to the
developing phase and year (+1) refers to the decaying phase of El Niño. By any measure, Figs. 1a–c suggest that class 1 El Niños are stronger than those in class 2. The fact that the temporal evolution characteristics of the variables follow an in-phase trajectory during year (0) for both classes allows us to construct difference maps (class 1 minus class 2 composites, Figs. 2 and 3). Three noticeable points in Figs. 1–3 are: (i) in class 1, an abrupt increase in westerly anomalies over the equatorial western Pacific occurs during May–June of year (0) (Fig. 1b). These changes coincide with the development of cold SST anomalies over the eastern equatorial Indian Ocean (EEIO, Fig. 1d). (ii) Subsequently, the thermocline deepens over the eastern equatorial Pacific (Fig. 3a) and Niño-3 SST (Fig. 1a) rises. (iii) During IODZM years, cold SST anomalies also develop over the Indonesian seas (10°S–0°, 110°–140°E) (Fig. 2a). These three points motivate us to study the effect of the SST anomalies over the near-equatorial Indian Ocean–Indonesian seas (hereafter referred to as regional SST anomalies) in generating the increase in zonal wind anomalies over the equatorial western Pacific in class 1.

The boreal fall of 2006 witnessed a strong IODZM but, contrary to the modeling results of AXMM05, the observed El Niño was only of moderate intensity (McPhaden 2008). Does this mean that the use of regional SST anomalies for predicting a stronger El Niño can lead to a false alarm? During IODZM years, the life cycle of cold SST anomalies over the Indonesian seas closely follows those over the EEIO (Figs. 2a,d and 3a). What are the main processes causing the SST anomalies over the Indonesian seas?

To answer the questions raised above, we performed diagnostics with CM2.1 integrations and observed conditions in 2006 and carried out a suite of experiments with a linear atmospheric model to identify the region that is responsible for generating the zonal wind anomalies. Sensitivity experiments were also conducted with an ocean model to identify the processes that may be responsible for the observed Indonesian seas SST anomalies in 2006 and in the CM2.1 simulations. We should stress here that we are not interested in studying whether El Niño triggers IODZM or vice versa.

Section 2 presents a brief description of observations, CM2.1 simulations, and the models used. Section 3 presents diagnostics from CM2.1 and results from the linear atmosphere model. Section 4 describes the observed characteristics for 2006 together with atmospheric and oceanic model results. Section 5 summarizes the conclusions and highlights the implications.

2. Data and models

2a. Data and method

We have examined the five-member ensemble twentieth-century (20c3m) integrations performed with CM2.1. The runs spanning the period 1861–2000 were conducted at GFDL as part of the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4). Model details including numerical and physical packages employed are described in Delworth et al. (2006); the
forcing fields used for the 20c3m runs are discussed in Knutson et al. (2006).

In the model runs, monthly anomalies were created with respect to 30-yr climatology (1971–2000). The secular trends in the SST time series were removed by linear regression analysis. Two SST indices—one averaged over the EEIO (10°S–0°, 90°–110°E) during June–November and another averaged over Niño-3 region (5°S–5°N, 90°–150°W) during December–February (DJF)—were constructed to represent IODZM and ENSO, respectively. Based on the threshold cutoff of 1.0 standard deviation on these indices, three classes are identified as (i) class 1: IODZM that co-occur with El Niño (36 events), (ii) class 2: El Niño only (65 events), and (iii) class 3:
FIG. 3. Monthly temporal evolution of (class 1 minus class 2 El Niño events) variables averaged over the equatorial region (6°S–6°N) shown for year (−1), year (0), and year (+1) of El Niño: (a) depth of the 20°C isotherm (m), (b) SST (°C), (c) zonal wind at 850 hPa (m s⁻¹), and (d) precipitation (mm day⁻¹). On the ordinate axis, year (0) represents the developing phase of El Niño (months 13–24) and year (+1) the decaying phase of El Niño (months 25–36); year (−1) refers to calendar months (1–12) prior to developing phase of El Niño.
models. To study the linkage between strong IODZM (1.0 standard deviation) and moderate El Niño (>0.5 but <1.0 standard deviation), class 4 (20 events) is also considered. Composites for selected variables are constructed and a t test is applied to assess their statistical significance.

CM2.1 climatology is compared to observations and reanalysis products for the period 1971–2000, except for precipitation whose climatology is based on the period 1979–2008. Variables include SST (Reynolds and Smith 1994), precipitation (Xie and Arkin 1996), winds from National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996), and thermocline depth from the Simple Ocean Data Assimilation product (Carton et al. 2000). In addition, the Hadley Centre interpolated SST (HadISST) dataset of Rayner et al. (2003) for the period 1871–2000 is also used to compare the model ENSO and IODZM statistics to observations.

Satellite-based observations are used to study the conditions in 2006. Monthly values of precipitation from the Tropical Rainfall Measuring Mission (TRMM) version 3B43, SST from TRMM Microwave Imager (TMI), surface wind from QuikSCAT, and three-dimensional convective–stratiform heating (CSH) profiles estimated from TRMM precipitation radar (Tao et al. 2006) are employed. For satellite products, monthly anomalies were created with respect to 10-yr climatology (1998–2007).

b. Models

1) LINEAR BAROCLINIC MODEL

A linear atmosphere model described in Watanabe and Jin (2003) is used. This model is a global, time-dependent, primitive equation model; for the present study, we linearized it around CM2.1 (NCEP) climatology when forcing fields are taken from CM2.1 composites (observations). The horizontal resolution is T42 and there are 20 vertical levels. Two versions are used: the “dry” version has prescribed heating, (i.e., convective heating functions are derived from precipitation) and the “moist” version is forced by SST in order to incorporate the feedback between convection and dynamics. Surface heat fluxes generated by SST forcing in the moist version are parameterized as in Betts and Miller (1986). A linearized moisture equation for perturbation specific humidity is used, and heat and moisture sources associated with cumulus convection are also parameterized (Watanabe and Jin 2003). In summary, the moist version generates its own heating functions. While all equations in the linear baroclinic model (LBM) are linear with respect to perturbation, advection of moisture and temperature are allowed, and the relationship between evaporation and temperature depends on the basic state. The LBM has been extensively used to study the processes causing tropical precipitation and circulation anomalies during ENSO (Watanabe and Jin 2003) and over the monsoon region (Annamalai and Sperber 2005; Annamalai 2010).

2) REGIONAL OCEAN MODEL

The Regional Ocean Modeling System (ROMS) (Shchepetkin and McWilliams 2005) is used for the ocean model. The k-profile parameterization scheme of Large et al. (1994) is used for the subgrid-scale vertical viscosity and diffusivity. A bulk-flux formula (Fairall et al. 1996) is used for atmospheric forcing, with incoming shortwave and longwave radiation, atmospheric temperature, specific humidity, precipitation, and pressure fields derived from the corrected normal year forcing product of Common Ocean–Ice Reference Experiments (Large and Yeager 2004). A spatial average of 2-min gridded elevations/bathymetry for the world (ETOPO2; data available online at http://www.ngdc.noaa.gov/mgg/fliers/06mgg01.html) is used as bottom topography.

Two sets of ocean models are used: one, the PacInd model (Antonov et al. 2006; Locarnini et al. 2006), has $1/8^\circ$ horizontal resolution and covers both the tropical Indian and Pacific Oceans (50°S–30°N, 30°E–70°W); the other, the IndArc model, has $1/8^\circ$ horizontal resolution and focuses only on the Indonesian seas (14°–1°S, 110°–142°E). In the PacInd model, the Indonesian Archipelago is modified to better represent the width of the straits that connect the Indonesian seas to the Pacific and Indian Oceans. Lombok Strait, however, is closed to avoid the large amount of the throughflow that passes over this strait contrary to observations (Gordon 2005). The IndArc model, used in Kida and Richards (2009), enables us to examine the processes within the Indonesian seas and allows us to test the robustness of our results to horizontal resolution.

3. Regional SST anomaly impact on zonal wind anomalies

In this section, the CM2.1 basic state and simulated statistics in ENSO and IODZM are briefly first described (section 3a). We then present our working hypothesis (section 3b). The hypothesis will be tested utilizing the LBM and the key region that is responsible for generating the increase in zonal wind anomalies along the equatorial Pacific during May–June of year (0) in class 1 events is elucidated (section 3c). We close the section by highlighting the point that regional SST anomalies and El Niño are selectively interactive systems (section 3d).
a. CM2.1 annual-mean basic state

Figure 4 shows the annual-mean climatology for CM2.1 (Figs. 4a–d) and observations (Figs. 4e–h) over the tropical Indo-Pacific region. Compared to observations, the Indo-Pacific warm pool in CM2.1 is confined to the equatorial region, resulting in excess rainfall there. In CM2.1, the strengthened equatorial easterlies combined with a shallow thermocline in the eastern equatorial Pacific favor an intense ENSO (Wittenberg et al. 2006; AchutaRao and Sperber 2006; Joseph and Nigam 2006). Over the TIO, weaker equatorial westerlies and the associated diffused thermocline over the EEIO may favor frequent IODZM events. The model SST standard deviations of Niño-3 and IODZM indices are 1.8° and 0.6°C, respectively, indicating that both ENSO and IODZM are overactive in CM2.1. In addition, the class 1 events tend to have a biennial tendency (Figs. 1 and 3b). To assess the gross measure of the simulated ENSO and IODZM, we show in Fig. 5 the probability distribution of Niño-3 (Fig. 5a) and EEIO (Fig. 5b) SST anomalies for CM2.1 and observations. In CM2.1, the probability distribution was estimated for each model run, and then a grand ensemble mean was constructed and shown. In agreement with observations, CM2.1 captures the skewed distributions of both ENSO and IODZM indices, but simulates stronger (greater than two standard deviations) El Niño and IODZM events more frequently. Nevertheless, the composites suggest that the spatial patterns of anomalous SST, thermocline depth, precipitation, and zonal wind along the equatorial Pacific are similar in the two classes except that class 1 events are stronger (Figs. 1–3). Thus, the model systematic errors do not necessarily influence the main results of our study.

b. Working hypothesis

Coupled models that are confined only to the tropical Pacific simulate a strong El Niño after few moderate events (Battisti 1988), a feature attributed to nonlinearity in the tropical Pacific (Timmermann and Jin 2002). On the other hand, coupled models that incorporated other tropical oceans found that variations in convection over the Maritime Continent–EEIO influence El Niño amplitude (Anderson and McCreary 1985; Kessler et al. 1995). Based on these results, our assumption is that in CM2.1 two possible, but not mutually exclusive, scenarios for stronger El Niño in class 1 may be expected: (i) they are entirely due to unstable air–sea interactions confined to the tropical Pacific and/or (ii), during the time that favors instability growth, wind anomalies forced by regional SST anomalies enhance the growth of El Niño. We will test each of these scenarios next.

1) Scenario One

Starting from November of year (−1) to April of year (0), positive precipitation anomalies over the equatorial western Pacific (130°–160°E) are sandwiched between negative precipitation anomalies (Fig. 3d). The zonal gradient in heating anomalies between the Maritime Continent and the western Pacific generates westerly wind anomalies over the equatorial western Pacific (Fig. 3c). Concomitantly, positive thermocline anomalies, suggestive of warm water accumulation, are noted around the date line (Fig. 3a). Under these preconditions downwelling Kelvin waves excited by the westerly wind anomalies (Fig. 3c) advect positive thermocline anomalies eastward. Our interpretation is that during the early months of year (0) indications are ripe for the development of stronger El Niños in class 1, (i.e., well before the initiation of IODZM). However, similar features are not noticeable in class 2 but are seen in class 3 composites [not shown here, but see Fig. 13 in Song et al. (2007)]. A further examination (not shown) suggests that the intensity of westerly wind anomalies is stronger for El Niño events of higher intensity (Niño-3 DJF SST index >1.5 and 2.0 standard deviations). The presence of equatorial westerly anomalies from November of year (−1) implies the role of stochastic forcing prior to the development of strong El Niño events (Kessler et al. 1995).

2) Scenario Two

During May–June of year (0), in both difference maps (Fig. 2 left panels and Fig. 3) and class 3 composites (Fig. 2 right panels), prominent signals are spatially coherent cold SST anomalies and suppressed rainfall over the EEIO–Indonesian seas (10°S–0°, 80°–140°E). Consistent with the known tropical circulation response to diabatic heating, easterlies prevail to the west and westerlies prevail to the east of the negative precipitation anomalies (Figs. 2c,f). The coherent signals in both SST and rainfall imply that during the development stages of stronger El Niño, air–sea interactions are also underway over the EEIO–Indonesian seas. Coinciding with IODZM development, a near-basinwide increase in wind anomalies is noticeable over the equatorial Pacific (Figs. 2c,f and 3b). This timing, May–June of year (0), coincides when mean conditions favor instability growth in the eastern equatorial Pacific (Battisti 1988; Wittenberg et al. 2006). Combining both scenarios, our working hypothesis is that under favorable conditions in the Pacific, regional SST and heating anomalies are capable of enabling the development of a stronger El Niño in class 1. Since many feedbacks in the fully coupled CM2.1 model make diagnosis of the processes impossible, we turn to LBM to test our hypothesis.
To check that the main focus of this study is not swayed by our methodology, we have also made composites based only on the ENSO index, separately for strong (>1.0 but <1.5 standard deviation) and very strong (>1.5 standard deviation) El Niño events (not shown). This new composite suggests no appreciable development of SST cooling over the EEIO (i.e., no IODZM development) during strong El Niño events. In contrast, IODZM does develop during very strong El Niño events. These composites, while based only on the ENSO index, are consistent with our earlier composites based on both ENSO and IODZM indices (class 1 and class 2). An alternate interpretation is that very strong El Niño is associated with the development of IODZM, an issue already discussed in many previous articles (e.g., Annamalai et al. 2003). As mentioned in section 1, our focus here is not to study whether or not El Niño triggers IODZM or vice versa.

FIG. 4. Annual mean climatology from (left) CM2.1 and (right) observations: (a),(e) SST (°C), (b),(f) precipitation (mm day⁻¹), (c),(g) winds at 850 hPa (m s⁻¹), and (d),(h) depth of the 20°C isotherm (m). Unit vector (15 m s⁻¹) in (c) and (g) is also shown.
c. Hypothesis demonstration

To identify the key region responsible for generating the zonal wind anomalies over the western equatorial Pacific during May–June of year (0), both dry and moist versions of the LBM are employed. In the dry version, the LBM is forced with an anomalous diabatic heating rate proportional to precipitation amplitude (boxed areas in Fig. 2b) and the vertical heating profile has a maximum at 400 hPa, as in Rodwell and Hoskins (1996). In the moist version, the LBM is forced with anomalous SST (boxed areas in Fig. 2a). Table 1 lists the main components of the experiments; unless otherwise mentioned, the LBM is integrated for 30 days with fixed forcing. With the dissipation terms adopted in the model, the tropics approaches steady state in about 15 days (Annamalai 2010) and the solutions averaged for days 20–25 are used for detailed analysis.

1) PRECIPITATION FORCING

Figures 6a–d show the response of the zonal wind anomalies for precipitation forcing over different regions during May–June of year (0). The response to total heating (Fig. 6d) closely resembles the composite results (Fig. 2c). In accordance with linear theory, easterly (westerly) anomalies to the east (west) of positive rainfall anomalies and westerly (easterly) anomalies to the east (west) of negative rainfall anomalies are apparent. For the wind anomalies over the equatorial Pacific, the net response due to TIO heating anomalies is near zero and the precipitation anomalies over the Indonesian seas contribute about 25%–30%, while the rest is due to equatorial central Pacific precipitation anomalies. However, the SST anomalies over the equatorial central Pacific are indeed small (Fig. 2a) to generate any large-scale precipitation anomalies (Fig. 2b). This issue is discussed next.

2) SST FORCING

Figure 7 shows the moist-version-generated zonal wind anomalies, and the corresponding precipitation anomalies are shown in Fig. 8. For the equatorial Pacific case (MOIST_EPAC), the simulated in situ precipitation anomalies are not strong enough to generate coherent zonal wind anomalies. For the EEIO forcing (MOIST_EEIO), dipolelike precipitation anomalies and easterly wind anomalies are captured over the equatorial Indian Ocean. The concentrated westerly anomalies over the southern Bay of Bengal and a similar, but lesser intensity, patch over the southeastern Indian Ocean are reminiscent of a Rossby wave response. Theoretical studies have shown that the basic state in boreal summer favors amplification of Northern Hemisphere Rossby waves (Wang and Xie 1997) and promotes positive precipitation anomalies along the monsoon trough (Annamalai 2010). In this experiment, the absence of zonal wind anomalies over the equatorial western Pacific can be understood as follows: the north–south heating anomalies over the TIO and monsoon region force opposite-signed Kelvin waves into the equatorial western Pacific; thus the net response cancels out. Similar arguments can be made for the east–west heating anomalies over the equatorial Indian Ocean (Figs. 6c and 2b), an interpretation consistent with AXMM05. In summary, moist LBM solutions indicate that neither the EEIO nor the equatorial Pacific SST anomalies are responsible for inducing the equatorial Pacific wind anomalies.

Figure 7b (Fig. 8b) shows zonal wind (precipitation) anomalies from the experiment forced with cold SST

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**Table 1. Details of various LBM experiments including the forcing region and forcing function derived from CM2.1 composites.**

<table>
<thead>
<tr>
<th>Expt</th>
<th>Forcing region</th>
<th>Forcing specifics</th>
</tr>
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<tbody>
<tr>
<td>DRY_EIO</td>
<td>(10°S–5°N, 50°–110°E)</td>
<td>Precipitation anomalies</td>
</tr>
<tr>
<td>DRY_IS</td>
<td>(10°S–0°, 120°–160°E)</td>
<td>Precipitation anomalies</td>
</tr>
<tr>
<td>DRY_EPAC</td>
<td>(10°S–10°N, 165°–280°E)</td>
<td>Precipitation anomalies</td>
</tr>
<tr>
<td>MOIST_EIO</td>
<td>(12°S–5°N, 65°–110°E)</td>
<td>SST anomalies</td>
</tr>
<tr>
<td>MOIST_IS</td>
<td>(10°S–5°N, 120°–150°E)</td>
<td>SST anomalies</td>
</tr>
<tr>
<td>MOIST_EPAC</td>
<td>(7°S–7°N, 180°–280°E)</td>
<td>SST anomalies</td>
</tr>
</tbody>
</table>
anomalies over the Indonesian seas (MOIST_IS). The off-equatorial westerly anomalies over the monsoon region and easterly anomalies to the west of the negative heating anomalies reflect a Rossby wave response. In this experiment, the generation of basinwide westerly anomalies along the equatorial Pacific is interpreted as follows. First, the presence of cold SST anomalies over the Indonesian sea results in a subdued east–west climatological SST gradient that subsequently favors westerly anomalies. Second, negative SST anomalies force in situ reduced precipitation whose Kelvin wave response results in anomalous westerlies. Third, net low-level convergence induced by anomalous westerlies favors the creation of positive heating anomalies in the equatorial central Pacific. Finally, due to interaction between convection and dynamics, the zonal gradient in heating anomalies between the Maritime Continent and the date line generates stronger westerly anomalies.

In summary, moist solutions suggest that about 60%–80% of the total equatorial western Pacific wind anomalies are due to Indonesian seas SST anomalies. While dry version solutions suggest the importance of equatorial Pacific precipitation anomalies in generating the wind anomalies, the moist version suggests that those precipitation anomalies are largely induced by Indonesian seas SST anomalies.

d. Regional SST anomalies and El Niño–selectively interactive systems

To understand if all IODZM events are associated with stronger El Niños, Figs. 9c–h show May–June and September–November composites of selected variables from class 4 events (moderate El Niño and strong IODZM). Based on a threshold of 0.5°C for the onset of El Niño and −0.5°C for IODZM onset, one can note from Fig. 9b that the initiation of El Niño occurs around May–June whereas IODZM starts in July–August (Fig. 9a). Compared to classes 1 and 2, the composite El Niño in class 4 has substantial differences. Apart from moderate intensity, the warm SST anomalies are largely confined around the date line (Figs. 9c,f), and the positive precipitation (Figs. 9d,g) and westerly wind anomalies (Figs. 9e,h)
are prominent only between 140°E and 180°. In class 4, despite the occurrence of a strong IODZM, the effect of regional forcing will be limited because of the presence of both positive and negative precipitation anomalies over the Indonesian seas region that will result in a net cancellation of Kelvin wave response over the equatorial western Pacific. It is thus clear that not all strong IODZM events are indicators for subsequent development of strong El Niño events. Similarly, not all El Niño events (e.g., class 2) are associated with IODZM development. These results suggest that the ENSO and IODZM are selectively interactive systems, provided that the ocean–atmosphere systems in the respective ocean basins respond to external factors [see also the discussions in Song et al. (2007)].

4. Observed characteristics in 2006

To further understand the significance of the Indonesian seas, we examine the observed IODZM and El Niño conditions during boreal fall 2006 (section 4a). We use the LBM but now forced with TRMM-derived CSH profiles and TMI SST anomalies (Table 2; section 4b). An ocean model is then used to identify the processes that caused the cold SST anomalies over the Indonesian seas in 2006 (section 4c).

a. IODZM and a moderate El Niño in 2006

Figure 10a shows time series of TMI SST anomalies averaged over the EEIO (solid), Indonesian seas (dotted), and Niño-3 (dashed) during 2006. The timing of Niño-3 (EEIO) SST anomalies reaching 0.5°C (−0.5°C) is conventionally regarded as the onset of El Niño (IODZM). The following points are inferred from Fig. 10a: (i) the IODZM is short lived with a late initiation in August, but peaks sharply in September, and demises in December; (ii) the evolution of Indonesian seas SST anomalies follows closely the life cycle of IODZM; and (iii) the El Niño onset occurs around August, peaks in December, and ends abruptly thereafter. The spatial patterns during
boreal fall (Figs. 10b–e) confirm the occurrence of a strong IODZM. Similar to class 4 events in CM2.1, the observed precipitation and westerly wind anomalies in 2006 are concentrated around the date line in the equatorial Pacific, suggestive of a moderate El Niño. Our interest is to isolate the contribution of Indonesian seas SST anomalies to the wind anomalies in 2006 and to identify the processes that caused those SST variations.

b. Impact of Indonesian seas on wind anomalies

Figures 11a–d show solutions obtained by forcing the LBM with observed three-dimensional CSH anomalies (Fig. 11e) and TMI SST anomalies (Figs. 11f–h). The horizontal and vertical structures of the CSH profiles are preserved with only a 1–2–1 spatial smoothing applied to the data. Both CSH- and SST-forced solutions suggest a negligible contribution from the TIO to the zonal wind anomalies over the equatorial western Pacific (Figs. 11a,b and 11f), whereas that of the Indonesian seas and equatorial Pacific is substantial (33%–50%). Interestingly, moist solutions indicate that Indonesian seas SST anomalies account for about 25% of the total precipitation anomalies around the date line (Figs. 10f–h). One notable difference is that the simulated zonal wind anomalies are stronger when the LBM is forced with SST (Fig. 11h) rather than with CSH (Fig. 11d) anomalies. One explanation is that the LBM generates only convective heating while CSH contains contributions from stratiform precipitation as well. The implication is that the transpacific gradient in stratiform rainfall becomes stronger during El Niño, causing changes in the vertical heating gradient (Schumacher and Houze 2003) and hence in low-level wind intensity.

The LBM solutions suggest the significance of the Indonesian seas SST on the observed conditions in 2006, but the observed El Niño was of moderate intensity. While there could be many reasons, we articulate the following: (i) the timing of the development of the regional cold SST anomalies in August was too late to favor instability growth in the eastern Pacific and (ii) the amount
FIG. 9. Class 4 composites of temporal evolution of (a) SST anomalies over the EEIO and (b) Niño-3 and Niño-3.4 SST indices for two years. (c)–(e) May–June composites of anomalous (c) SST, (d) precipitation, and (e) zonal wind. (f)–(h) September–November composites of anomalous (f) SST, (g) precipitation, and (h) zonal wind. In (c)–(h), positive values are shaded progressively and negative values are shown as contours.
of warm water accumulated over the equatorial western Pacific was indeed low (McPhaden 2008), a feature probably internal to the tropical Pacific.

c. Cause of cold SST anomalies over the Indonesian seas

We have suggested the importance of SST over the Indonesian seas throughout the manuscript. But how does SST anomaly over the Indonesian seas develop during IODZM years? Past studies (e.g., Sprintall et al. 2000) have suggested some role by the equatorial Indian Ocean wind anomalies (remote), while recent model experiments by Kida and Richards (2009) suggest that variations in local surface winds are the primary factor in determining the annual cycle of SST over the Indonesian seas. We therefore examine whether the interannual SST variations are caused by local or remote wind anomalies.

1) REMOTE EFFECTS

The PacInd model is first used to investigate the impact of remote processes on Indonesian seas SST variations. Monthly climatological wind derived from QuikSCAT is used to spinup and the model is integrated for 20 years to obtain realistic climatological SST and near-surface flow fields. In particular, the model simulates the basic seasonal variability including the meridional migration of the warm pool, although the simulated SST is higher attributed to lack of synoptic variance in the monthly forcing fields (not shown). In model year 21 from January onward the wind forcing over the equatorial Indian Ocean (5°S–5°N, 60°–95°E) gradually becomes that of the 2006 observations. We will refer to this experiment as OCN_IND, and the year 21 solution subtracted from the 5-yr model climatology (years 16–20) is termed the simulated 2006 anomalous conditions.

During boreal fall 2006, OCN_IND, forced with zonal wind anomalies only over the equatorial Indian Ocean, simulates an east–west contrast in SST (Fig. 12a) and thermocline depth (Fig. 12b) anomalies, representative of the IODZM. Compared to observations (Fig. 10b), the simulated SST anomalies are slightly weaker (−0.5°C) over the western, but stronger (−1.0°C) over the eastern, Indian Ocean. Various ocean models forced with equatorial wind anomalies have reproduced the east–west contrast in SST and thermocline depth anomalies during IODZM years (Murtugudde et al. 2000). The present results are consistent with earlier ones. However, the OCN_IND experiment shows no appreciable SST anomaly development over the Indonesian seas (Fig. 12a). Possible reasons are (i) the upwelling oceanic Kelvin wave forced by the easterly wind anomalies of the equatorial Indian Ocean did not penetrate into the Indonesian seas with significant magnitude to affect the SST; (ii) by construction, PacInd model does not resolve Lombok Strait where the Kelvin wave is observed to penetrate into the Indonesian seas (Sprintall et al. 2000); and (iii) the Kelvin wave dissipates by the time it reaches Timor Strait where the Indonesian seas and the Indian Ocean meet (Fig. 12b) such a scenario is supported from observations (Potemra et al. 2002). While our model experiment cannot determine which reason is likely to be the cause, it does suggest that SST anomalies in the Indonesian seas are not affected strongly by remote winds. This finding is supported by a sensitivity experiment in which the Paclnd model is forced with an IODZM composite of CM2.1 surface wind fields (not shown).

2) LOCAL EFFECTS

To investigate local effects, winds over the Indonesian seas (13°–3°S, 115°–140°E) are gradually shifted from climatology to that observed in 2006 in model year 21. We will refer to this experiment as OCN_IS. The anomalies are calculated in the same manner as in OCN_IND. Both experiments are identical except that in model year 21 the forcing regions are different. OCN_IS simulates a cold SST anomaly of −2°C in the Indonesian seas analogous to observation (Fig. 10b), thus supporting our premise that the local wind is likely the cause for inducing SST anomaly over the Indonesian seas. To further examine whether lateral advection of the Indonesian Throughflow has an effect, the IndArc Model is used. As mentioned in section 2, this model configuration examines the impact of the local winds in isolation from the lateral advection from adjacent seas. Climatological wind from QuikSCAT is used to force the model for 5 years of spinup, then in year 6 the wind field gradually becomes that observed during 2006, and the integration is continued until April 2007. We will refer to this experiment as OCN_IS_HIGH and the SST anomaly is estimated as the difference between year 6 solutions and the 3-yr model climatology (years 3–5).

OCN_IS_HIGH simulates a strong cold SST anomaly of −2°C over the Indonesian seas (Fig. 13a), similar to OCN_IS, which suggests that the cold SST anomaly during fall 2006 is indeed caused by interannual variability

<table>
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<tr>
<th>Expt</th>
<th>Forcing region</th>
<th>Forcing specifics</th>
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<tr>
<td>DRY_WIO</td>
<td>(10°S–10°N, 50°–75°E)</td>
<td>CSH anomalies</td>
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<tr>
<td>DRY_EIO</td>
<td>(10°S–10°N, 75°–110°E)</td>
<td>CSH anomalies</td>
</tr>
<tr>
<td>DRY_IS</td>
<td>(10°S–10°N, 120°–160°E)</td>
<td>CSH anomalies</td>
</tr>
<tr>
<td>DRY_EPAC</td>
<td>(0°–10°N, 165°–280°E)</td>
<td>CSH anomalies</td>
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<tr>
<td>MOIST_EIO</td>
<td>(12°S–5°N, 65°–110°E)</td>
<td>SST anomalies</td>
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<tr>
<td>MOIST_IS</td>
<td>(10°S–5°N, 120°–150°E)</td>
<td>SST anomalies</td>
</tr>
<tr>
<td>MOIST_EPAC</td>
<td>(7°S–7°N, 170°–280°E)</td>
<td>SST anomalies</td>
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Fig. 10. (a) Monthly SST anomaly (°C) indices from January 2006 to April 2007; 2006 boreal fall (b) anomalous SST (°C), (c) 10-m zonal winds (m s\(^{-1}\)), (d) 10-m meridional winds (m s\(^{-1}\)), and (e) precipitation (mm day\(^{-1}\)). The LBM-generated precipitation anomalies (mm day\(^{-1}\)) for (f) equatorial Indian Ocean SST forcing, (g) Indonesian seas SST forcing, and (h) equatorial Pacific SST forcing. In (b)-(h), positive values are shaded progressively and negative values as contours.
FIG. 11. Steady-state response of zonal wind anomalies at 850 hPa (m s\(^{-1}\)) for CSH forcing for boreal fall 2006: (a) western equatorial Indian Ocean, (b) eastern equatorial Indian Ocean, (c) Indonesian seas, and (d) equatorial Pacific. (e) CSH heating anomalies (K day\(^{-1}\)).

Steady-state response of zonal wind anomalies at 850 hPa for TMI SST forcing for boreal fall 2006 over the (f) equatorial Indian Ocean, (g) Indonesian seas, and (h) equatorial Pacific. In all panels, positive (negative) values are shaded progressively (shown as contours). Solid boxes in (e) correspond to forcing regions in the LBM experiments (Table 2).
in the local winds over the Indonesian seas. A mixed layer heat budget over the whole model domain also shows that this cold SST anomaly is produced because of stronger surface heat loss due to stronger latent heat flux release and subsurface upwelling (Fig. 13b). In summary, the three numerical ocean model experiments conducted in this section strongly suggest that the cold SST anomaly during fall 2006 is a result of a stronger and probably prolonged monsoon season.

5. Summary and discussion

a. Summary

In this study, we examine the possible role of variations in SST and precipitation over the equatorial eastern Indian Ocean and Indonesian seas on the amplitude of a developing El Niño in multicentury integrations of the CM2.1 coupled model. Composite evolution of selected variables (Figs. 1–3) suggests that El Niño is much stronger when the El Niño and IODZM co-occur (class 1) than when only El Niño occurs (class 2). Based on this inference, two possible but not mutually exclusive reasons for stronger events are proposed. The first is that stronger El Niño events are due to air–sea interactions confined to the tropical Pacific. The difference maps (class 1 minus class 2 composites, Fig. 3) suggest that during the early months of year (0) indications are ripe for the development of stronger El Niños in class 1. Thus, the possible role of stochastic forcing for the development of strong El Niños is not ruled out (Kessler et al. 1995). The second possibility is that during May–June of year (0), coinciding with IODZM development, an increase in

![Fig. 12. OCN_IND simulated anomalous conditions during boreal fall (September–November) 2006: (a) SST (°C) and (b) 22°C isotherm depth (m), a measure of model thermocline depth. (c) OCN_IS simulated anomalous SST (°C) during boreal fall 2006.](image-url)
equatorial zonal wind anomalies over the western Pacific occurs in class 1. The LBM solutions forced with anomalous precipitation and SST (Fig. 2) over different regions of the tropical Indo-Pacific reveal that SST anomalies over the Indonesian seas account for about 35%–65% of those wind anomalies (Figs. 6–8) and also determine anomalous precipitation over the equatorial Pacific (Fig. 8). These are new findings of the present study.

Our interpretation of the role of regional SSTs is as follows. First, the presence of cold SST anomalies over the Indonesian sea results in a suppressed east–west climatological SST gradient that favors westerly anomalies. Second, negative SST anomalies force a reduction in local precipitation whose Kelvin wave response results in anomalous westerlies. Third, low-level convergence induced by anomalous westerlies favors creation of positive heating anomalies in the equatorial central Pacific. Finally, owing to the interaction between convection and dynamics, the zonal gradient in heating anomalies between the Maritime Continent and the date line generates stronger westerly anomalies. The possible effect of the wind anomalies is to force stronger downwelling oceanic Kelvin waves that lead to excess thermocline anomalies in the eastern equatorial Pacific (~10 m, Fig. 3) during August–September of year (0). Our conclusion is that regional SST and heating anomalies are not the primary cause but likely enable the development of stronger El Niños in class 1. This conclusion, however, needs to be verified through performing prediction experiments with fully coupled models.

Experiments using the ocean models suggest that oceanic waves triggered by the equatorial Indian Ocean wind anomalies are not the primary cause for Indonesian seas SST anomalies but that local wind anomalies determine them. Mixed layer heat budget analysis suggests that local buoyancy based entrainment is one of the principal factors that cause SST cooling.

b. Discussion

Over the TIO, it is well recognized that ENSO influences both the basinwide SST variations (Venzke et al. 2000; Xie et al. 2002) as well as IODZM development (Annamalai et al. 2003). For instance, ENSO-induced wind anomalies over the equatorial Indian Ocean force downwelling oceanic Rossby waves that then influence the thermocline and SST anomalies over the southwest Indian Ocean (Xie et al. 2002). During their growth both IODZM and basinwide SST anomalies modify local precipitation and circulation anomalies. These changes in the TIO are then suggested to influence either the growth (AXMM05) or phase transition of ENSO (Kug and Kang 2006). In this view, years when both strong El Niño and IODZM co-occur there is a possibility for regional SST anomalies to feed back onto El Niño.

Identifying the factors responsible for the occurrence of stronger El Niño events is essential for a successful prediction of global climate anomalies. Our earlier study (AXMM05) and the present one suggest that ocean–atmosphere processes over the EEIO and Indonesian seas may also feed back to developing El Niño. In particular, the identification of the Indonesian seas SST anomalies that follow closely the life cycle of the IODZM deserves attention. Annamalai et al. (2003) conjectured that the IODZM is a part of the equatorial Indo-Pacific warm pool variability, and the present results support that view. It should be mentioned here that either in CM2.1 or in observations (see Fig. 4a in AXMM05) cold SST anomalies do not develop over the Indonesian seas during years when only El Niño occurred.

Our results also suggest that not all strong IODZM years are associated with stronger El Niño events (classes 3 and 4), implying that the occurrence of IODZM is not always a presage for a stronger El Niño development. Similarly, appearance of a deeper than normal thermocline around the date line in the beginning of a calendar year [year (−1) in Fig. 3a] is not always associated with stronger El Niño development. While either is a necessary condition, their simultaneous existence appears necessary for stronger El Niño development.

For realistic simulation of ENSO, Wittenberg et al. (2006) highlighted the need for better representation of regional SST, particularly over the Indonesian seas, for proper simulation of westerly wind associated with
intraseasonal variability. Jochum and Potemra (2008) indicated the role of tidal mixing parameterization in coupled models for accurate simulation of Indonesian seas SST anomalies. By analyzing long integrations from the Community Climate System Model, version 3 (CCSM3), Potemra and Schneider (2007) note the dominant effect of local winds on upper-ocean Indonesian Throughflow transport and also find that such conditions occur when cold SST anomalies prevail in both the Indonesian seas and EEIO. In the latest version of CCSM3, Neale et al. (2008) note a significant improvement in ENSO simulation but their diagnostics also suggest the presence of coherent SST and wind anomalies over the Indonesian seas and equatorial western Pacific well before El Niño initiation (their Fig. 5). Thus, for a variety of reasons a proper representation of regional SST anomalies in coupled models and recognition of their impact on equatorial wind anomalies over the western Pacific are important to understand and predict El Niño. While the conceptual model of Jansen et al. (2009) does not find improvement in ENSO forecast skill by including the TIO, sensitivity experiments with a fully coupled model need to be performed to understand the implications of the present study. We note here that the atmospheric model indicated the role of SST, while the ocean model indicated the role of the wind. These results then indicate that coupled dynamics are important to understand the underlying dynamics occurring over the Indonesian Seas, which would be the next step of this work. On those lines, our future study will focus on the air–sea interaction over the Indonesian seas in more detail, with a particular aim at understanding the origin of local wind anomalies and on performing idealized coupled models experiments to understand the role of regional SST anomalies with regard to ENSO.

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