Dynamic and Thermodynamic Air–Sea Coupling Associated with the Indian Ocean Dipole Diagnosed from 23 WCRP CMIP3 Models*

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ABSTRACT

The performance of 23 World Climate Research Programme (WCRP) Coupled Model Intercomparison Project, phase 3 (CMIP3) models in the simulation of the Indian Ocean dipole (IOD) is evaluated, and the results show large diversity in the simulated IOD intensity. A detailed diagnosis is carried out to understand the role of the Bjerknes dynamic air–sea feedback and the thermodynamic air–sea coupling in shaping the different model behaviors. The Bjerknes feedback processes include the equatorial zonal wind response to SST, the thermocline response to the equatorial zonal wind, and the ocean subsurface temperature response to the thermocline variation. The thermodynamic feedback examined includes the wind–evaporation–SST and cloud–radiation–SST feedbacks. A combined Bjerknes and thermodynamic feedback intensity index is introduced. This index well reflects the simulated IOD strength contrast among the strong, moderate, and weak model groups. It gives a quantitative measure of the relative contribution of the dynamic and thermodynamic feedback processes.

The distinctive features in the dynamic and thermodynamic coupling strength are closely related to the mean state difference in the coupled models. A shallower (deeper) equatorial mean thermocline, a stronger (weaker) background vertical temperature gradient, and a greater (smaller) mean vertical upwelling velocity are found in the strong (weak) IOD simulation group. Thus, the mean state biases greatly affect the air–sea coupling strength on the interannual time scale. A number of models failed to simulate the observed positive wind–evaporation–SST feedback during the IOD developing phase. Analysis indicates that the bias arises from a greater contribution to the surface latent heat flux anomaly by the sea–air specific humidity difference than by the wind speed anomaly.

1. Introduction

The Indian Ocean dipole (IOD) is a basin-scale ocean–atmosphere coupled mode, characterized by a zonal contrast of a positive and a negative sea surface temperature anomaly (SSTA) along the equatorial Indian Ocean (IO) and a zonal wind anomaly over the central equatorial IO (Saji et al. 1999; Webster et al. 1999; Li et al. 2002, 2003). During a positive IOD event, the SST is anomalously cool in the southeastern IO (SEIO) off Java–Sumatra and warm in the western IO (WIO), accompanied with a pronounced anomalous southeasterly along the coast of Sumatra and anomalous easterlies over the central equatorial IO (CEIO). While the IOD rapidly develops in boreal summer, it reaches a mature phase in boreal fall. A number of studies showed that the convection associated with IOD exerted great impacts on the climate variability in Africa, South Asia, East Asia, and other remote regions (e.g., Saji and Yamagata 2003a,b; Ashok et al. 2004; Behera et al. 2005; Matthew et al. 2006).

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It was suggested that Bjerknes feedback (Bjerknes 1966, 1969) may operate in the equatorial IO (Webster et al. 1999; Li et al. 2003). This dynamic feedback involves interactions among the zonal SST gradient, the low-level zonal wind in the CEIO, and the east–west thermocline displacement. For example, a negative SSTA off Sumatra would induce anomalous low-level easterlies in the CEIO, which depress (lift) the thermocline to the west (east). The lifted thermocline to the east may enhance the SST cooling through the upwelling of anomalous cold subsurface water. The enhanced surface cooling further amplifies the easterly anomaly. Through this positive dynamic feedback, the IOD develops. Because of the strong seasonal change of the monsoon winds, conditions are favorable for the Bjerknes feedback over boreal summer and early fall season in the IO, sometimes leading to the development of an IOD event (Schott et al. 2009).

In addition to the dynamical coupling, the thermodynamic air–sea feedback also played a role during IOD development. Different from the Pacific El Niño, during which there is a spatial phase shift between convection and SST anomalies, SST and cloud anomalies are, in general, in phase in the tropical IO (Li et al. 2003). This implies a stronger cloud–radiation–SST negative feedback in the IO. Another notable feature during the IOD development is a season-dependent wind–evaporation–SST feedback (Li et al. 2003). A pronounced southeasterly mean flow in boreal summer leads to a positive wind–evaporation–SST feedback; that is, a cold SST in the SEIO forces a low-level anticyclonic flow to its west (Gill 1980), and an anomalous southeasterly along the coast of Java–Sumatra associated with this anticyclone strengthens the surface evaporation and cools the SST further. The seasonality of the IOD is also somehow regulated by the maximum upwelling off Java and Sumatra during boreal summer and fall. Li et al. (2003) pointed out that it is the season-dependent air–sea feedback that is responsible for the occurrence of the maximum SSTA in the SEIO in boreal fall.

The strength of the dynamic and thermodynamic feedbacks mentioned above depends greatly on the mean state of the tropical IO. For example, the mean thermocline depth and the background upwelling velocity and vertical temperature gradient are critical in determining the effect of the dynamic feedback. A deeper mean thermocline, a smaller mean upwelling velocity, and a weaker upper-ocean vertical temperature gradient may reduce heat exchange between the subsurface and surface (Li 1997a,b; Neelin et al. 1998). As discussed above, the wind–evaporation–SST feedback also depends on the background monsoon flow.

Given the great impact of the IOD on circulation and rainfall variations in various regions, it is crucial to simulate and predict the IOD evolution and amplitude as well as its related climate impact in current state-of-art climate models (Luo et al. 2007, 2008). Various efforts have been made in evaluating the performance of IOD simulations in coupled general circulation models (CGCMs) (e.g., Yamagata et al. 2004; Cai et al. 2005; Zhong et al. 2005; Luo et al. 2007; Song et al. 2007; Cai et al. 2009). The objective of the present study is to conduct a detailed diagnosis of the dynamic and thermodynamic feedbacks associated with the IOD from 23 coupled models that participated in the Coupled Model Intercomparison Project phase 3 (CMIP3). CMIP3 was organized by the World Climate Research Programme (WCRP) Climate Variability and Predictability (CLIVAR) Working Group on Coupled Models (WGCM) Climate Simulation Panel. It collected global coupled atmosphere–ocean general circulation model experiments targeted at twentieth-and twenty-first-century climate simulations, as well as climate change experiments for assessment in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4). CMIP3 represents the largest and most comprehensive international global coupled climate model experiment and multimodel analysis effort ever attempted (Meehl et al. 2007). Saji et al. (2006) compared IOD simulations from 17 CMIP3 models and found a great diversity among the models in IOD intensity. The cause of the difference among these models, however, was not addressed. In this study we intend to conduct a detailed analysis of 23 CMIP3 models, with a particular emphasis on the diagnosis of the dynamic and thermodynamic coupling strength in each model. As the coupling strength may depend on the basic state, the simulated background states of the 23 models will also be examined.

The rest of the paper is organized as follows: Section 2 briefly describes the datasets used. Section 3 examines the IOD simulations from the 23 coupled models. Section 4 further reveals the dynamic and thermodynamic coupling strength of each model. The relationship of the air–sea coupling strength to the background mean state is examined in section 5. Section 6 discusses the possible cause of misrepresentation of the wind–evaporation–SST feedback and the cloud–radiation–SST feedback in some of the CMIP3 models. Finally, a summary is given in section 7.

2. Data

The primary datasets used for this study are the outputs of 23 models from the WCRP CMIP3 datasets. The model variables used in the diagnosis include 3D ocean temperature, current, surface wind, cloud fraction, specific humidity, and surface heat flux fields. Table 1 lists information for all 23 models used in this study (further detailed information can be obtained by visiting...
Most of the models are labeled by the name of the institution that performed the run and supplied the data. The particular datasets analyzed here are from the so-called Twentieth-Century Climate in Coupled Model (20C3M). The forcing agents of this experiment include greenhouse gases \([\text{CO}_2, \text{CH}_4, \text{N}_2\text{O}, \text{and chlorofluorocarbons (CFCs)}]\), sulfate aerosol direct effects, volcanoes, and solar forcing. The results reported in this analysis were only for the period between January 1950 and December 1999. The mean climatological annual cycle was defined based on this 50-yr period. The interannual anomalies were then obtained by subtracting the monthly-mean variables from their respective mean climatological annual cycles.

For comparison of the model simulations with the observations, various observed and reanalyzed (assimilated) atmospheric and oceanic datasets were used. Three-dimensional atmospheric wind fields, the surface specific humidity, the surface latent heat flux, cloud cover, and the net surface shortwave radiation (SWR) were obtained from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP-NCAR) reanalysis (Kistler et al. 2001) and 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40) products, and three-dimensional ocean temperatures were obtained from the Simple Ocean Data Assimilation (SODA) reanalysis (Carton et al. 2000). In addition, we used both the Met Office Hadley Centre Sea Ice and Sea Surface Temperature

Table 1. List of the 23 CMIP3 models and corresponding Indian Ocean dipole indices as well as the observed indices: (second column) the ISI value for each model, (third column) the dipole mode index (DMI), the SSTA difference between western Indian Ocean \(10^\circ\text{S}-10^\circ\text{N}, 50^\circ-70^\circ\text{E}\) and eastern Indian Ocean \(10^\circ\text{S}-0^\circ, 90-110^\circ\text{E}\). (fourth column) the Southeast Indian Ocean Dipole Mode Index (SEDMI) defined as the SSTA averaged over \(10^\circ\text{S}-0^\circ, 90-110^\circ\text{E}\). The fifth and sixth columns list the short and official names of each model.

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<th>Number</th>
<th>ISI</th>
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<th>SEDMI</th>
<th>Model short name</th>
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3. IOD simulations from 23 CMIP3 models

Regardless of the difference in the simulated IOD intensity, the 23 CMIP3 models all exhibit a common character; that is, the maximum standard deviation of the zonal SSTA gradient appears in boreal fall. The result is consistent with Saji et al. (2006). An examination of the zonal wind standard deviation field shows a similar phase-locking feature; that is, the maximum zonal wind variation in the CEIO appears in boreal fall.

Figure 1 shows the composite SSTA and 850-hPa zonal wind anomaly fields from the observation and each of the 23 CMIP3 model outputs. The composites are made based on the dipole mode index (DMI) defined by Saji.
et al. (1999), with the index exceeding one standard deviation during September–November (SON). It is obvious that some models generate a strong IOD variability, whereas others show a relatively weak IOD variability. Most of the models reproduce the maximum SSTA variability in the SEIO off Sumatra and the maximum zonal wind variability in the CEIO.

As shown in Fig. 1, the 23 models exhibit a great diversity in the strength of simulated IODs. To objectively distinguish the difference of IOD strength among these models, we introduce an IOD strength index (ISI). Different from previous indices that measure only the ocean temperature anomaly strength [e.g., the zonal SST gradient between the western and eastern Indian Ocean (Saji et al. 1999) or the Sumatra cooling strength (Xie et al. 2002; Meyers et al. 2007; Luo et al. 2010)], the current index depends on the product of the zonal SSTA gradient and the zonal wind anomaly in the CEIO and, thus, it better reflects the air–sea coupling nature of the IOD events.

The ISI is defined as a product of the standard deviation of the zonal difference between the SSTA averaged in the SEIO (10°S–0°, 90°–110°E) and the WIO (10°S–10°N, 50°–70°E) and the standard deviation of the zonal wind anomaly averaged in the CEIO (5°S–5°N, 70°–90°E) during September–November. Based on this definition, we calculate the IOD strength for each model. Table 1 lists the values of the ISI calculated for all 23 models. For comparison, Table 1 also lists two other dipole model indices: the original dipole mode index (DMI) defined by Saji et al. (1999), the SSTA gradient between (10°S–10°N, 50°–70°E) and (10°S–0°, 90°–110°E) and the southeast Indian Ocean dipole mode index (SEDMI), defined as the SSTA averaged over (10°S–0°, 90°–110°E) (Luo et al. 2010). As one can see, the three indices (DMI, SEDMI, and ISI) give qualitatively consistent results.

Based on the ISI values, the 23 models are classified into three groups: a strong IOD simulation group (with ISI > 2.9), a moderate IOD simulation group (1.1 < ISI < 2.9), and a weak IOD simulation group (ISI < 1.1). In Table 1, the 23 models are sorted based on the ISI values (from largest to smallest). Based on the above classification, the strong IOD simulation group includes the following six models: csiro.mk35, iap, ncar.pcm1, gfdl1, mpi, and cnrm.cm3. The weak IOD simulation group consists of the following six models: cccma, cccma.t47, ncar.ccsm3, ipsl, giss1, and giss.aom.

A composite analysis is conducted to reveal robust contrasting features between the strong and weak groups.
different models simulate the IOD strength differently and which air–sea coupling processes contribute to the difference? In this section, through the diagnosis of dynamic and thermodynamic coupling strength of each model during the developing phase in July–September (JAS), we will reveal the fundamental causes of the IOD simulation diversity among the models.

a. Bjerknes feedback intensity

We examine the Bjerknes feedback strength in all models. This dynamic air–sea coupling consists of three processes. The first is how the atmospheric low-level wind responds to the SSTA forcing in the SEIO. It is well known that during the IOD development period (JAS), the most significant feature of SSTA is the dipole pattern. Such a zonal SSTA gradient causes a zonal wind response in the CEIO through the SST-gradient-induced pressure gradient in the atmospheric planetary boundary layer (Lindzen and Nigam 1987) or the midtropospheric heating anomaly (Gill 1980).

To quantitatively measure the strength of the zonal wind response to the SSTA, we plot a scatter diagram for each model. Figure 4 shows the relationship between the SSTA in the SEIO (10°S–0°, 90°–110°E) and the zonal wind anomaly in the CEIO (5°S–5°N, 70°–90°E) from the 23 models. For comparison, the observed SSTA–wind relation is also plotted in the top left corner. Consistent with the observed relationship, all of the models exhibit a positive correlation between the zonal wind and SST anomalies; that is, a negative SSTA in the SEIO [which corresponds to a positive IOD event according to Saji et al. (1999)] is accompanied by an easterly anomaly in the CEIO.

We hereby denote a SSTA–wind coupling coefficient, \( R(u, T) \), to represent how strong the low-level wind responds to a unit SSTA forcing at each model. Mathematically, it equals the linear slope at each of scatter diagrams in Fig. 4 (the linear slope in all scattering diagrams was calculated based on the least squares fitting method). It is interesting to note that the averaged slope in the strong composite is about 2.2 m s\(^{-1}\) K\(^{-1}\), which is a little larger than the averaged slope (1.7 m s\(^{-1}\) K\(^{-1}\)) in the weak composite. Compared to the observed \( R(u, T) \), the SSTA–wind coupling coefficient in the strong group appears more realistic.

The second process involves how the ocean thermocline responds to the equatorial surface wind forcing. The zonal wind anomaly in the CEIO, in general, forces the ocean thermocline response through the propagation of induced oceanic Rossby and Kelvin waves (Li 1997b; Li et al. 2002, 2003; Yu et al. 2005).

Figure 5 shows the scatter diagrams between the zonal wind anomaly in the CEIO and the thermocline depth

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**Figure 3.** Time evolution of the composite SSTA (solid line) in the SEIO (0°–10°S, 90°–110°E) and 1000-hPa zonal wind anomaly (dashed line) in the CEIO (5°N–5°S, 70°–90°E) averaged for six strong (blue) and six weak (red) models. The composite was made based on DMI exceeding one standard deviation during SON.
anomaly in the SEIO for each model. The observed feedback relation between the wind and SEIO thermocline depth anomaly is positive because an easterly wind anomaly in the CEIO may lift (suppress) the thermocline in the eastern (western) part of the basin. Such a positive relationship is captured by all of the coupled models.

We hereby denote a wind–thermocline coupling coefficient $R(D, u)$ to measure the strength of thermocline depth change for a given unit zonal wind forcing. It can be represented by the slope in Fig. 5. The averaged $R(D, u)$ for the strong composite is 4.2 m (m s$^{-1}$)$^{-1}$, while it is about 2.9 m (m s$^{-1}$)$^{-1}$ for the weak composite. The observed $R(D, u)$ is somewhere between the values of the strong and weak composites.

The third process involves how the ocean subsurface temperature responds to the ocean thermocline variation. In the SEIO, a shoaling (deepening) of the thermocline leads to a negative (positive) temperature anomaly at a fixed subsurface level. The change of the subsurface temperature may further affect SST through the anomalous vertical temperature advection by the mean upwelling. In the region of deep climatological mean thermocline, the subsurface temperature variation is small and so is the SST variability.

Figure 6 shows the observed and the model-simulated relationship between the thermocline depth anomaly and the temperature anomaly at 70 m [where the subsurface temperature variability is largest, Hong and Li (2010)] in the SEIO. A positive correlation appears between the observed thermocline depth and subsurface temperature. We hereby denote $R(T_e, D)$ as a therocline–subsurface temperature coupling coefficient, which can be measured by the slope in Fig. 6. The observed slope is 0.1 K m$^{-1}$, implying that a 1-m thermocline change would lead to a subsurface temperature change of 0.1 K. Most of the CMIP3 models reproduce such a positive relationship, even though the slope is markedly different. Three models—giss3, giss.aom, and giss1—reproduce a weak slope. The comparison of the strong versus the weak composite shows that the thermocline–subsurface temperature feedback slope in the former is about 20% greater than that in the latter.

The overall Bjerknes dynamic feedback strength should be determined by the combined effect of the three above processes. To quantitatively measure the Bjerknes feedback intensity and to compare it with the thermodynamic feedback intensity, we introduce a simplified SST tendency equation:
where $T$ and $T'$ denote the surface and subsurface ocean temperature anomalies, respectively; $\bar{w}$ denotes the climatological mean vertical velocity at the base of the ocean mixed layer; $Q'$ is the net surface heat flux anomaly; $\rho$ and $C_w$ are the seawater density and specific heat, respectively; and $h$ is the ocean mixed layer depth. In Eq. (1), we only show the thermocline feedback and heat flux terms and have neglected other advection terms for simplicity. Assuming $T' = \delta T e^{\sigma t}$, Eq. (1) may be rewritten as

$$\frac{\partial T'}{\partial t} = \frac{T'}{\bar{w} h} + \frac{Q'}{\rho C_w h},$$

where $T'$ and $T=e^{\sigma t}$ denote the surface and subsurface ocean temperature anomalies, respectively; $\bar{w}$ denotes the climatological mean vertical velocity at the base of the ocean mixed layer; $Q'$ is the net surface heat flux anomaly; $\rho$ and $C_w$ are the seawater density and specific heat, respectively; and $h$ is the ocean mixed layer depth.

In Eq. (1), we only show the thermocline feedback and heat flux terms and have neglected other advection terms for simplicity. Assuming $T' = \delta T e^{\sigma t}$, Eq. (1) may be rewritten as

$$\frac{\rho C_w h \sigma}{\rho C_w h} = \frac{T'}{\bar{w} h} + \frac{Q'}{T'},$$

Here the left-hand side of Eq. (2) is proportional to the growth rate ($\sigma$) of the SST$A$. The second term in the rhs of Eq. (2) represents how strong the surface heat flux anomaly is in response to a unit SST$A$ change. Thus, it reflects the strength of the thermodynamic air–sea feedback. The first term in rhs of Eq. (2) represents the vertical advection of anomalous subsurface temperature by the mean upwelling velocity, thus reflecting the strength of the Bjerknes dynamic air–sea feedback. We hereby define the first term on the rhs of Eq. (2) as the Bjerknes feedback intensity index (BFI) and the second term as the thermodynamic feedback intensity index (TFI). The BFI may be written as

$$\text{BFI} = \rho C_w \bar{w} R(u, T) R(D, u) R(T_e, D).$$

Equation (3) states that the BFI depends on the mean vertical velocity and a product of the SST–wind, wind–thermocline, and thermocline–subsurface temperature coupling coefficients during the IOD developing phase. It measures the overall strength of the Bjerknes feedback in each model. Figure 7 shows the averaged values of the SST–wind, wind–thermocline, and thermocline–subsurface temperature coupling coefficients and BFI for the strong, moderate, and weak composites. For comparison, the observed coupling coefficients and BFI are also shown in the figure. Note that the BFI is consistent with the overall strength of the IOD simulations, with the greatest (smallest) value occurring in the strong (weak) composite. A Student’s $t$ test was used to examine the statistical significance. Our calculation shows that the BFI difference between the strong and weak group is significant at the 95% confidence level.

The above analysis points out the important role of the dynamic air–sea coupling in determining the strength
of the model IODs. Saji et al. (2006) showed the relationship between the standard deviation of DMI and the mean thermocline depth off Java (see their Fig. 13a) and found that, although there is a weak tendency for a larger DMI variance to occur in a model with a shallower mean thermocline off Java, the scatter is quite large. This implies that the intensity of the IOD in a model is not simply determined by the Bjerknes feedback but that other processes may also play a role. Given that the IOD is also affected by the thermodynamic air–sea coupling, in the following we will examine the wind–evaporation–SST feedback and the cloud–radiation–SST feedback in the SEIO from the 23 models.

b. Thermodynamic air–sea feedback strength

As demonstrated by Li et al. (2002, 2003), two types of thermodynamic air–sea feedback processes were involved during the IOD developing phase. One is a positive feedback among the SST, surface wind, and evaporation. This positive feedback is attributed to the background southeasterly in the SEIO, under which an anomalous southeasterly induced by a cold SSTA may further enhance the cold SSTA through increased surface evaporation. Another is a negative feedback among the SST, cloud, and shortwave radiation anomalies; that is, a cold SSTA leads to the decrease of clouds, which further increases the downward shortwave radiation and decreases the cold SSTA.

How well do the CMIP3 models simulate such a positive wind–evaporation–SST feedback? To quantitatively measure the feedback strength, we plot the scatter diagram (Fig. 8) to illustrate the relationship between
the surface latent heat flux (LHF) anomaly and the SSTA in the SEIO in JAS. For the comparison, the observed counterpart is also plotted.

We hereby denote \( R(LHF, T) \) to represent the SST–evaporation feedback coefficient, which can be measured by the slope in each panel of Fig. 8. The observed feedback coefficient is 6.02 \( \text{W m}^{-2} \text{K}^{-1} \), which implies that, given 1 K SST cooling, the resulting latent heat flux anomaly is 6.02 \( \text{W m}^{-2} \). This amount of anomalous heat flux could be used to further enhance the local SST cooling. Surprisingly, many CMIP3 models (including those in the strong simulation group) fail to reproduce such a positive feedback process. The cause of this failure will be analyzed in section 6. The large bias in the wind–evaporation–SST feedback poses a great challenge to many state-of-art coupled GCMs.

To illustrate how different the cloud–radiation–SST feedback is among the 23 models, we show in Fig. 9 the simultaneous relation between the shortwave radiation and SST anomalies averaged in JAS over the SEIO. Different from the positive wind–evaporation–SST feedback, a negative feedback between the observed SST and shortwave radiation anomalies exists in the SEIO. While most of the coupled models reproduce such a negative feedback process, two models (csiro, giss1), surprisingly, exhibit a weak positive feedback.

We hereby denote \( R(SWR, T) \) to measure the strength of the negative cloud–radiation–SST feedback. The averaged slope for the strong composite is \(-14 \text{ W m}^{-2} \text{K}^{-1}\), which is about twice as large as that \((-7 \text{ W m}^{-2} \text{K}^{-1})\) in the weak composite.

The overall thermodynamic feedback intensity (TFI) may be measured by the sum of the wind–evaporation–SST feedback and the cloud–radiation–SST feedback, that is,

\[
\text{TFI} = R(LHF, T) + R(SWR, T).
\]

Figure 10 shows the diagrams of \( R(LHF, T) \), \( R(SWR, T) \), and TFI for the strong, moderate, and weak composites and for the observation. The major bias appears in the latent heat flux–SST relationship. Consequently, the thermodynamic damping in strong and moderate groups is overestimated, compared to the observation. The strongest (weakest) thermodynamic damping appears in the strong (weak) group. The difference between the strong and weak group is significant at the 95% confidence level for TFI.

c. Combined Bjerknes and thermodynamic feedback intensity

The diagnosis above reveals that greater (weaker) Bjerknes positive and thermodynamic negative feedbacks
coexist in the strong (weak) group. Thus, it is necessary to examine the combined dynamical and thermodynamic feedback processes. A combined dynamic and thermodynamic feedback intensity index (CFI) is defined as

$$\text{CFI} = \text{BFI} + \text{TFI}. \quad (5)$$

Figure 11 illustrates the averaged CFI for the strong, moderate, and weak groups. For comparison, the observed CFI is also shown. The CFI values in the three groups reflect well the simulated IOD strength; that is, the greater the CFI, the stronger the overall dynamic and thermodynamic coupling strength, and the greater the model IOD intensity. The difference between the strong and weak group is significant at the 90% confidence level for CFI.

5. Differences in the coupled model mean state

Why do some CMIP3 models produce greater dynamic and thermodynamic coupling on the interannual time scale while others do not? As the interannual anomalies evolve under the background mean state, it is necessary to examine the difference in the mean climate state among the coupled models.

From the point of view of the ocean dynamics, a deeper background thermocline implies a weaker Bjerknes feedback strength. This partly accounts for the fact that the SSTAs associated with El Niño grow much faster in the eastern equatorial Pacific than in the western equatorial Pacific. Figure 12 shows the scatter diagram of the mean thermocline depth–ISI relationship for each of the coupled models. Note that the averaged annual mean thermocline depth in the strong composite is 110 m, with
a small annual variation (figure not shown), while the
mean thermocline depth is 200 m for the weak com-
posite. This mean thermocline depth difference may
indirectly contribute to the difference in the Bjerknes
feedback intensity between the strong and weak groups.
The difference of the mean thermocline depth between
the strong and weak groups is significant at the 95%
confidence level.

The difference in the background ocean vertical ve-
locity is another possible factor that contributes to the
distinctive Bjerknes dynamic air–sea coupling strength
in the CMIP3 models. Figure 13 illustrates the climato-
logical annual cycle of the vertical velocity at the base
of the mixed layer in the SEIO from both strong and weak
composites. As shown in Eq. (3), the BFI depends on
both the mean vertical upwelling velocity and the ratio
of the subsurface and surface temperature anomalies.
The physical argument behind this is that, given the
same thermocline depth anomaly or the same subsur-
face temperature anomaly, the anomalous vertical ad-
vection by a stronger (weaker) mean upwelling velocity
would lead to a greater (smaller) SSTA tendency (Li
1997b). The changed SSTA would further alter the wind
and thus the thermocline depth, magnifying the SSTA
change. Figure 13 illustrates that the mean vertical up-
welling during the IOD developing phase is greater
(smaller) in the strong (weak) composite. Thus, through
a positive thermocline–SST feedback, the SSTA grows
at a greater rate in a model with a larger background
mean upwelling velocity.

In addition to the anomalous temperature advection
by the mean vertical velocity, the advection of the mean
vertical temperature gradient by anomalous vertical
velocity may also contribute to the SSTA growth rate.

Considering an idealized case in which a wind pertur-
bation induces the identical anomalous vertical velocity,
the difference in the background vertical temperature
gradient would lead to the different SSTA tendency
through the anomalous vertical advection of the mean
temperature gradient, that is, \( w' \partial \bar{T} / \partial z \). This term was
shown to be important in determining both annual and
interannual SST variations in the eastern equatorial

\[ \text{FIG. 11. The combined Bjerknes–thermodynamic feedback in-
tensity index (CFI) (unit: W m}^{-2} \text{K}^{-1}) \text{ during the IOD developing}
\text{phase (JAS) for S, M, W, and O (as in Fig. 7).} \]

\[ \text{FIG. 12. Scatter diagram illustrating the ISI-mean thermocline}
\text{depth represented by the depth of the 20°C isothermal relation-
ship from the 23 CMIP3 models. The star, diamond, and cross}
\text{represent the strong, moderate, and weak groups, respectively,}
\text{with black (colored) characters denoting individual (composite)
model results.} \]

\[ \text{FIG. 13. Climatological annual cycles of the vertical velocity}
\text{(m s}^{-1} \text{)} \text{ at the base of the mixed layer in the SEIO for the strong}
\text{(solid) and weak (dashed) composites. The ‘‘x’’ and ‘‘o’’ represent}
\text{the standard deviation among the six strong and six weak models,}
\text{respectively.} \]
Pacific (Li and Philander 1996; Li 1997a) and the SSTA in the SEIO during the IOD (Hong et al. 2008a,b). The generated SSTA may further amplify the difference through the change of the anomalous wind and upwelling velocity fields. Figure 14 illustrates the climatological annual cycle of the background vertical temperature gradient in the upper ocean for the strong (solid) and weak (dashed) composites. The "x" and "o" as in Fig. 13.

To sum up, the diagnosis of the model mean climate state shows that the strength of dynamic and thermodynamic air–sea coupling on the interannual time scale appears closely related to the model performance in the mean state simulation. Marked differences appear in the mean thermocline depth, the upper-ocean vertical temperature gradient, and the upwelling velocity between the strong and weak groups. Therefore, the mean state differences contribute to the difference in the IOD simulations between the strong and weak groups.

The examination of the summer mean surface wind field and the vertical cross section of the mean ocean temperature along the equatorial IO shows that the difference of the mean upwelling velocity fields between the strong and weak groups is closely related to the wind difference (figure not shown). Strengthened (weakened) summer northward cross-equatorial flows correspond well with strengthened (weakened) upwelling velocity off Sumatra. Meanwhile, the strengthened (weakened) alongshore wind helps lift (deepen) the thermocline, leading to the decrease (increase) of ocean subsurface temperature and thus the increase (decrease) of upper-ocean vertical temperature gradients.

6. Discussion

Although most of the WCRP CMIP3 models could reproduce the IOD-like events and relevant dynamic and thermodynamic feedback processes in the tropical IO, unreasonable simulations exist among these models. The most significant bias is the wind–evaporation–SST feedback during the IOD developing phase. Among 23 coupled models, 11 of them show a reversed relationship between the surface LHF and SST anomaly in the SEIO compared with the observation (Fig. 8). What causes such a bias?

Because the surface LHF in the coupled models depends primarily on the surface wind speed and sea–air specific humidity difference, we take a further analysis of these quantities to disclose the possible causes of the incorrect SST–LHF relationship in the coupled models.

Figure 15 shows the composite maps of the anomalous SST, surface wind, surface wind speed, sea − air specific humidity, and LHF fields for \( R(LHF, T) \) and \( R(LHF, T) < 0 \) models. For comparison, we also plot the observed counterparts. The observational composite shows a negative SSTA in the SEIO and pronounced southeasterly anomalies off the Sumatra coast. The anomalous wind leads the enhanced surface wind speed, as the mean wind is also southeasterly (second row in Fig. 15). Because the sea − air specific humidity difference anomaly is, in general, proportional to the SSTA on seasonal and longer time scales (Li and Wang 1994), the \( (q_s - q_a) \) anomaly is negative in the SEIO (third row in Fig. 15). With the two opposing effects competing, a negative surface LHF anomaly results, indicating that the wind speed anomaly dominates the LHF in the region. This favors a positive wind–evaporation–SST feedback, as discussed in Li et al. (2003).

The positive \( R(LHF, T) \) model composite shows a relatively weak SSTA in the SEIO compared to the negative \( R(LHF, T) \) model composites. However, the anomalous southeasterly in the SEIO is stronger in the former than in the latter. As a result, the surface wind speed anomaly is greater (smaller) in the positive (negative) \( R(LHF, T) \) model composite. A marked difference appears in the \( q_s - q_a \) field. Whereas a great negative value appears in the SEIO in the negative \( R(LHF, T) \) composite, a much weaker averaged sea − air specific humidity difference anomaly appears in the positive \( R(LHF, T) \) composite. As a result, the LHF anomaly
is primarily controlled by the surface wind speed anomaly in the $R_{(LHF, T)} > 0$ composite, whereas it is mainly determined by the $(q_s - q_a)$ anomaly in the $R_{(LHF, T)} < 0$ composite. As a consequence, the LHF anomaly is negative (positive) in the former (latter) composite. The contrast of the opposite sign of the LHF anomaly between the middle and right columns of Fig. 15 implies distinctive wind–evaporation–SST feedback mechanisms operated in the SEIO between the two model groups.

The composite results shown in Fig. 15 include all CMIP3 models, regardless of their IOD intensity. We further examine the wind speed, $q_s - q_a$, and LHF characteristics for the strong, moderate, and weak groups. Figure 16 gives the averaged values of the wind speed anomaly, $q_s - q_a$ anomaly, and LHF anomaly in the SEIO for the three groups and the observation. For each group, we separate the models into two subgroups according to $R_{(LHF, T)} > 0$ and $R_{(LHF, T)} < 0$. It turns out that, no matter whether the models are in a strong, moderate, or weak group, a major difference between the two subgroups appears in the relative strength of the surface wind speed and sea–air specific humidity difference anomalies. A correct (incorrect) SST–LHF relationship is achieved only when the wind speed (the humidity difference) anomaly dominates the LHF anomaly. Such a relationship is clearly reflected by all cases in the left (right) panel of Fig. 16.

The distinctive characteristic of the wind–evaporation–SST feedback for each model may be further revealed by comparing the correlations of time series of the LHF anomaly in the SEIO with the local wind speed anomaly and sea–air specific humidity difference anomaly. Table 2 shows the correlation coefficients for each model group. The common feature for all positive $R_{(LHF, T)}$ groups is a greater LHF correlation with the wind speed anomaly,
consistent with the observation. The negative LHF–wind correlation implies a positive wind–evaporation–SST feedback scenario. In contrast, in all negative $R(LHF, T)$ groups, anomalous LHF depends more on the sea — air specific humidity difference field than on the surface wind speed field.

The analyses above indicate that the unrealistic simulations of the anomalous latent heat flux in about a half of the 23 CMIP3 models are primarily attributed to the unrealistic simulations of the near-surface air specific humidity field. The high dependence of the LHF on the sea — air specific humidity difference leads to a negative, rather than a positive, wind–evaporation–SST feedback in the SEIO during the IOD developing phase. In contrast, the observed latent heat flux anomaly is more closely related to the wind speed anomaly, indicating that the wind speed anomaly is a primary factor in determining the surface LHF anomaly. This observed feature is well captured by the other half of the models. The humidity bias may give model developers a clue as how to improve the representation of the LHF-related thermodynamic air–sea feedback. The unrealistic LHF–SST relationship in the SEIO may be attributed to the difference in the atmospheric static stability. Zheng et al. (2010) noticed that the increased static variability under global warming could weaken the zonal wind response to SSTA forcing during the IOD development, which would lead to an incorrect SST–LHF relation. In addition to the wind speed bias, the incorrect representation of the sea — air specific humidity difference is another important error source for the incorrect LHF–SST relation in some of the coupled models.

Table 2. Correlation coefficients between the time series of anomalous LHF and surface wind speed [C(LHF, WS)] and sea–air specific humidity difference [C(LHF, $q_s - q_a$)] in the SEIO. O denotes the observation; SP and SN denote the positive and negative $R(LHF, T)$ in the strong group respectively, MP and MN denote the positive and negative $R(LHF, T)$ in the moderate group, and WP and WN denote the positive and negative $R(LHF, T)$ in the weak group.

<table>
<thead>
<tr>
<th></th>
<th>O</th>
<th>SP</th>
<th>SN</th>
<th>MP</th>
<th>MN</th>
<th>WP</th>
<th>WN</th>
</tr>
</thead>
<tbody>
<tr>
<td>$C(LHF, WS)$</td>
<td>−0.81</td>
<td>−0.67</td>
<td>−0.18</td>
<td>−0.62</td>
<td>−0.05</td>
<td>−0.62</td>
<td>0.18</td>
</tr>
<tr>
<td>$C(LHF, q_s - q_a)$</td>
<td>0.28</td>
<td>0.53</td>
<td>−0.39</td>
<td>−0.05</td>
<td>−0.31</td>
<td>0.03</td>
<td>−0.76</td>
</tr>
</tbody>
</table>

In addition to the wind–evaporation–SST feedback, the shortwave radiation–SST feedback is another important thermodynamic air–sea coupling process that could impact the IOD strength. While most of the CMIP3 models captured the observed negative relation between the shortwave radiation and SST anomaly, two models—csiro and giss1—failed to reproduce the observed feature. To understand the cause of the unrealistic simulation of the shortwave radiation during the IOD developing phase, we plot in Fig. 17 the composite maps of the SST and cloud cover anomalies. Here the composite is made based on $R(SWR, T) > 0$ and $R(SWR, T) < 0$ model groups. For comparison we also show the observed counterpart. In the observation, in response to a negative SSTA in the SEIO, the local cloud cover anomaly is negative. This implies that local convection is suppressed and more shortwave radiation may reach the ocean surface to damp the cold SSTA. Similar to the observation, negative $R(SWR, T)$ models capture well a negative cloud cover anomaly over the SEIO. It is interesting to note that the positive $R(SWR, T)$ model composite shows a positive cloud cover anomaly right above the cold SSTA in the SEIO and a negative cloud cover anomaly appears to be shifted to its west. This unrealistic anomalous cloud pattern implies a positive rather than a negative shortwave radiation–SST feedback. The analysis demonstrates that apparently there is a bias in cloud fraction calculation in the csiro and giss1 models, which might lead to the unrealistic cloud–SST feedback over the region.

As shown in Eq. (2), the mixed layer depth is also important in determining the growth rate of the IOD. Figure 18 shows the composite mixed layer depth in the SEIO for the strong, moderate, and weak groups as well as from the observation. As one can see, the difference of the averaged mixed layer depth among the three groups is relatively small (less than 10 m), with the strong (weak) group exhibiting a shallower (deeper) mixed layer depth. The mixed layer depth effect on the growth rate is illustrated in the lower panels of Fig. 18.
which show the modified BFI, TFI, and CFI values. As one can see, the modified CFI seems better at representing the intensity difference among the strong, moderate, and weak groups. The observed feedback index is somewhere between the strong and moderate groups.

The results presented in this section point out possible biases in the thermodynamic air–sea coupling in some of the current state-of-art coupled GCMs. The comparison of overall performance of the IOD simulations among the 23 models suggests that the most serious problem lies in the thermodynamic air–sea coupling. As most of these GCMs are also used for future climate projection, a caution is needed in interpreting the model-generated global and regional SST changes.

7. Summary and conclusions

The Indian Ocean dipole is one of dominant modes in the tropical IO on the interannual time scale. One issue with important scientific and social relevance is how well the state-of-art coupled atmosphere–ocean GCMs are capable of simulating IOD events. In this study we evaluate the performance of 23 WCRP CMIP3 models on the aspect of the IOD simulation, by examining in great detail the dynamical and thermodynamical air–sea coupling processes.

An IOD strength index (ISI) is introduced to measure the IOD intensity in the 23 models. This index takes into account both the zonal SST gradient and surface wind anomaly and thus it better reflects the air–sea interaction nature of the IOD. Based on the ISI, the 23 coupled models are classified into strong, moderate, and weak IOD simulation groups. The composite analysis is further conducted to reveal common features and differences among the strong and weak groups.

To reveal the cause of the diversity in the model IOD intensity, we first examine the Bjerknes feedback strength in all 23 models. This dynamic ocean–atmosphere feedback consists of the following three key processes: 1) how strong the atmospheric low-level wind responds to a one unit SST forcing, 2) how strong the ocean thermocline depth responds to a one unit surface wind forcing, and 3) how strong the ocean subsurface temperature responds to a one unit thermocline depth variation. These three dynamical feedback processes are examined and the respective coupling coefficients are estimated in all 23 models. The overall strength of the Bjerknes dynamic feedback is determined by the product of the three coupling coefficients and the mean upwelling velocity. The comparison of the strong and weak composites shows that the former attains a much greater Bjerknes feedback intensity than the latter.

Next, we examine the thermodynamic air–sea coupling strength for all 23 models. Two thermodynamic air–sea feedback processes, the wind–evaporation–SST feedback and the cloud–radiation–SST feedback, are specifically examined. While the observation shows a positive feedback among the wind, evaporation (or surface latent heat flux), and SST during the IOD developing phase (Li et al. 2003), about a half of the CMIP3 models failed to capture such a positive thermodynamic air–sea

![Fig. 17. Composite maps of the SST (shading, unit: K) and cloud cover fraction (contours) anomalies during the IOD developing phase (JAS) from (a) the observation, (b) the negative $R(S_{WR}, T)$ model composite, and (c) the positive $R(S_{WR}, T)$ model composite. For each model, the composite was made based on the four most significant negative/positive $R(S_{WR}, T)$ events [(b),(c)] over a negative SSTA in the SEIO ($10^\circ S$–$0^\circ$, $90^\circ E$–$110^\circ E$).](image)

![Fig. 18. The (a) composite mixed layer depth $h$ (m) and modified feedback indices—i.e., (b) BFI, (c) TFI, and (d) CFI—divided by $h$ (unit: $W \cdot m^{-3} \cdot K^{-1}$) during JAS for S, M, W, and O (as in Fig. 7).](image)
feedback. The cause of this latent heat flux bias lies in the errors in both the surface wind speed and the sea–air specific humidity difference fields in these models. The composite result shows that a weak positive wind–evaporation–SST feedback occurs in both strong and weak groups, while a negative feedback appears in the moderate group.

As a strong negative feedback process, the cloud–radiation–SST feedback may slow down the IOD development. Most of the CMIP3 models successfully simulated this negative feedback process, even though the feedback intensity varies among the models. The averaged negative feedback coefficient is greater (smaller) in the strong (weak) composite, implying a stronger (weaker) thermodynamic damping.

Given the offset effect in the dynamic and the thermodynamic air–sea feedback, a combined Bjerknes and thermodynamic feedback intensity index (CFI) is introduced. The CFI well reflects the simulated IOD strength among the strong, moderate, and weak groups. It gives a quantitative measure of the integrated contribution of the dynamic and thermodynamic feedback processes.

The distinctive features in the dynamic and thermodynamic coupling between the strong and weak composite are closely related to the difference in the coupled model mean state. Comparison of the strong and weak composites shows that there are remarkable differences in the mean thermocline depth, vertical temperature gradient, and upwelling velocity in the equatorial IO. The models with a deeper mean thermocline, a weaker mean vertical temperature gradient, and a weaker mean upwelling are often associated with weaker dynamic coupling strength and a weak IOD signal. A further examination of the mean thermocline depth simulations shows that the large bias in the weak group arises from two individual models. The long-term mean 20°C isothermal depth fields in the two models have a large positive bias over the entire Indian Ocean, implying that the bias may result from interbasin water exchange. The exact cause of the bias is unclear and deserves further observational analyses and modeling studies. The differences of the mean vertical temperature gradient and the mean vertical velocity between the strong and weak groups are not statistically significant (slightly below the 90% confidence level). This implies that the BFI difference between the two groups is primarily attributed to air–sea coupling coefficients \( [i.e., R(u, T) \times R(D, u) \times R(T_{\text{w}}, D)] \), not to the mean upwelling velocity.

In this study we examine the detailed dynamic and thermodynamic air–sea feedback processes in 23 CMIP3 models and reveal their possible roles in the diversity of IOD simulations and the connection to the model mean state. The methodology applied here may be readily extended to other climate phenomena, such as ENSO, and may provide general guidance for coupled model assessment and improvement. For instance, a comparison of scatter diagrams of the SST–wind, wind–thermocline, thermocline–subsurface temperature, SST–latent heat flux, and SST–shortwave radiation relations with the observations may provide model developers about what parts of air–sea feedback processes or model physics might be inaccurate and need to be further improved. The unusually deep mean thermocline in some of the weak IOD simulation models needs to be fixed so as to reproduce a realistic oceanic response to atmospheric forcing and a realistic air–sea coupling strength.

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REFERENCES


