A Diagnostic Study of the Indian Ocean Dipole Mode in El Niño and Non–El Niño Years

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ABSTRACT

The Indian Ocean dipole mode (IODM) is examined by comparing the characteristics of oceanic and atmospheric circulations, heat budgets, and possible mechanisms of IODM between El Niño and non–El Niño years. Forty-year ECMWF Re-Analysis (ERA-40) data, Reynolds SST data, and ocean assimilation data from the Modular Ocean Model are used to form composites of the IODM that occur during El Niño (1972, 1982, and 1997) and non–El Niño (1961, 1967, and 1994) years. In El Niño years, two off-equatorial, anticyclonic circulations develop, associated with the increased pressure over the eastern Indian Ocean. The anticyclonic circulation over the Northern Hemisphere enhances the easterly component of the winds in the northwestern Indian Ocean. This enhanced easterly component increases the mixed layer temperature by inducing an anomalous westward ocean current that advects the warm mean mixed layer from the central to the western Indian Ocean. Meanwhile, the anticyclonic circulation over the southeastern Indian Ocean strengthens southeasterlies, thereby causing oceanic meridional and vertical advection of the cold mean temperature. Consequently, the IODM in El Niño years is characterized by the warming in the northwestern and the cooling in the southeastern Indian Ocean. In non–El Niño years, a monsoonlike wind flow increases the westerly and southeasterly components of the wind over the northwestern and southeastern Indian Ocean, respectively. Oceanic currents induced by these winds result in anomalous cold advection in both of these regions. In addition, the monsoonlike wind flow over the southeastern Indian Ocean enhances the anomalous latent and sensible heat fluxes in non–El Niño years. Hence, the cooling of the eastern tropical Indian Ocean, rather than the warming of the western Indian Ocean, becomes the major feature of the IODM during non–El Niño years.

1. Introduction

On interannual time scales, the Indian Ocean dipole mode (IODM) is one of the dominant modes in the tropical Indian Ocean. The spatial structure of the IODM can be characterized by negative sea surface temperature anomalies (SSTAs) in the southeastern tropical Indian Ocean (ETIO), and positive SSTAs in the western tropical Indian Ocean (WTIO). The possible impacts of this spatial structure may cause anomalous precipitation over East Africa, the tropical Indo-Pacific region (Black et al. 2003), and the Indian summer monsoon region (Terray et al. 2003). In particular, the presence of the IODM during El Niño years may reduce the influence of an El Niño on the Indian summer rainfall (Ashok et al. 2004). In addition, Saji and Yamagata (2003) suggested that the impact of the IODM reaches several remote regions away from the Indian Ocean. They found a strong correlation between the IODM, warm land surface temperatures, and reduced rainfall over Europe, northeast Asia, North and South America, and South Africa. Therefore, an understanding of the IODM is critical for the prediction of the Indian summer monsoon system and possibly other regions on a global scale.

As the role of the IODM in climate variability has gained attention in recent years, many efforts have been made to explain its formation. Gualdi et al. (2003) analyzed the IODM in a coupled model, and suggested a mechanism for the formation of the IODM in El Niño years. During the developing phase of an El Niño, positive sea level pressure anomalies are created in the southeastern part of the tropical Indian Ocean. Associated with this anomalous sea level pressure are en-
hanced southeasterly anomalies, which set up the favorable conditions for the IODM (Gualdi et al. 2003).

The role of southeasterlies anomalies on the formation of the IODM is further investigated by Li et al. (2003). According to their theory, the presence of both anomalous and mean southeasterlies near the coast of Sumatra in summer enhances the evaporative cooling in this area.

Assume initially there is a modest cold SSTA off Sumatra. Since the southeastern Indian Ocean is a region of intense convection, the cold SSTA implies the decrease of atmospheric convective heating or an atmospheric heat sink. According to Gill’s (1980) solution, the heat sink will induce a descending Rossby wave response to its west, resulting in an anomalous low-level anticyclonic flow. In the northern summer, the mean flow is southeasterly. Thus, the anomalous wind enhances the total wind speed and lowers the SST further through enhanced surface evaporation, vertical mixing, and coastal upwelling.

This positive feedback in air–sea interaction is one of the mechanisms that can enhance the IODM. Another sustaining mechanism of the IODM is proposed by Annamalai et al. (2003). That is, when southeasterlies develop off of Sumatra, they induce alongshore upwelling and trigger the IODM, which grows in summer by a Bjerknes-type feedback process.

In spite of these general agreements on the role of southeasterlies, there is a lack of a conclusive theory on the variability of the IODM. For example, theories on the IODM range from viewing the IODM as a self-sustained independent mode (Saji et al. 1999; Webster et al. 1999) to connecting the IODM with El Niño (e.g., Lau and Nath 2003; Annamalai et al. 2003; Li et al. 2003; Loschnigg et al. 2003; Shinoda et al. 2004a). The theory for the independent IODM is supported by statistical evidence (Saji et al. 1999; Yamagata et al. 2002; Behera et al. 2003) and by coupled general circulation model simulations that can simulate the IODM without El Niño (Iizuka et al. 2000; Fischer et al. 2005).

Annamalai et al. (2003) supported the importance of El Niño by suggesting that the natural mode of the coupled variability of the eastern equatorial Indian Ocean is weak on its own but intensifies in spring/early summer, usually when El Niño–like conditions exist in the western Pacific. Lau and Nath (2003) and Shinoda et al. (2004a) also showed that SST variations in the central and eastern Pacific are capable of producing realistic El Niño–related zonal wind variations and surface heat flux anomalies over the eastern tropical Indian Ocean, through one-dimensional mixed layer processes. These surface heat flux anomalies capture the cooling of the eastern Indian Ocean during boreal summer and fall, followed by a rapid warming in boreal winter, resulting in the development of basin-wide warming that peaks 2–3 months after an El Niño event. However, surface heat flux anomalies fail to explain the variability in the western Indian Ocean. They hypothesized that the exclusion of ocean dynamics in the one-dimensional mixed layer model is the source of the discrepancy in the western Indian Ocean.

While these studies were successful in describing the impact of El Niño on the eastern part of the IODM, the variability in the western Indian Ocean deserves more elaboration. Connecting the variability of the western Indian Ocean to El Niño requires not only the air–sea interaction, but also dynamical oceanic processes (Shinoda et al. 2004a). From this point of view, oceanic processes and air–sea interaction in both the western and eastern regions of the Indian Ocean need to be considered when the development of the IODM during El Niño and non–El Niño years is compared.

In this study, we extend the work of Shinoda et al. (2004a) by providing a comprehensive comparison between the IODM in association with or without an El Niño based on the compilation of the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analyses (ERA-40) data, Reynolds SST data, and ocean analysis outputs. The comparison between the IODM during El Niño and non–El Niño years ranges from atmospheric circulations prior to the IODM to the oceanic and atmospheric heat budgets, which describe the possible mechanisms of each case of the IODM. Since the ocean analysis data provide the state variables in three dimensions, the examination of the three-dimensional oceanic mixed layer processes is possible in this study. Furthermore, in order to simplify the diagnostic analysis, a linear estimation is applied so that the interaction between the climatological annual cycle (mean) and the interannual variability (anomalies) is explicitly identified. Although it is not our interest to overly simplify or emphasis the role of linear processes, this diagnostic analysis can be used to understand the linear mechanisms, thereby supplementing existing studies of the IODM (Iizuka et al. 2000; Li et al. 2003; Vinayachandran et al. 2002; Annamalai et al. 2003; Loschnigg et al. 2003; Shinoda et al. 2004a,b).

In the next section, a description of the data and models is given, followed by an explanation of the composite method (section 3). The formation of the IODM during the El Niño years is examined in section 4. The formation of the IODM during non–El Niño years is presented in section 5. Finally, the main results are summarized in section 6.
2. Data and models

The data used for the composite of the IODM are the ERA-40 data, the Reynolds SST data (Reynolds and Smith 1994), and an ocean analysis from the Modular Ocean Model (MOM) with the assimilation of the temperature profile of the World Ocean Dataset 1998 (WOD98; Conkright et al. 1998, Masina et al. 2004). The version of the MOM used in this study is the eddy permitting version (Cox 1984; Rosati and Miyakoda 1988) with a longitudinal resolution of 0.5, and a meridional resolution varying from a minimum of 1/3 between 10°S and 10°N to a maximum 0.5° at the northern boundary. The 31 vertical levels are unevenly spaced, with the first 14 levels confined to the upper 450 m. The model is initialized with the ocean at rest, and the climatology of the temperature and salinity taken from winter WOD98. The cloud cover used in the MOM is derived from the climatology cloud cover of the Comprehensive Ocean–Atmosphere Data Set (COADS). The atmospheric forcing variables, such as the air and dewpoint temperatures at 2 m, the mean sea level pressure, and winds at 10 m, are taken from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis project (Kalnay et al. 1996) in order to compute the momentum and heat fluxes interactively with the velocity and sea surface temperature (Rosati and Miyakoda 1988).

The assimilation scheme, used in conjuncture with MOM, consists of the univariate variational optimal interpolation scheme with some changes to the parameters in the original usage in Masina et al. (2001). In the present formulation, based on the scheme developed by Derber and Rosati (1989), the global temperature profile of WOD98 is assimilated into MOM down to the depth of 250 m by applying a correction to the forecast temperature field at every model step. In this variational approach (Lorenc 1986), the correction is made by a weighted average between the model field and the observation. In other words, observational data with a time window of 15 day (7 day to either side) are weight averaged with the model field at every time step. The weight given to the observation increases as the time difference between the observation and model simulation approaches zero. The benefit of this algorithm is that whenever/wherever the observational data are not available, the model provides the solution. Thus, the preservation of the model physics in the assimilation scheme prevents the missing data from affecting the result, and enables us to generate monthly mean fields with minimal model adjustment perturbations. This is a significant difference from the other assimilation schemes, which do not take into account the model physics. Detailed descriptions of MOM and the assimilation procedure can be found in Masina et al. (2001, 2004).

3. Methods

a. Composite of the IODM

The monthly means of the ERA-40 and ocean analysis data are averaged from 1959 to 1999 to produce the climatological annual cycle. The anomaly field is then calculated by subtracting the climatological annual cycle from the monthly means. The years of 1961, 1967, 1972, 1982, 1994, and 1997 are selected to represent the major IODM events (Saji et al. 1999). In those years, the maximum of the normalized IODM (IODM/standard deviation of IODM) exceeds two (Figs. 1a and 1c). In the years 1972, 1982, and 1997, the IODM accompanies the major El Niño events in which the maximum of the normalized Niño-3 SSTA equals (or exceeds) two (Fig. 1b). In the years 1961, 1967, and 1994, the Niño-3 SSTA does not show the increasing trends throughout the year, and none of their maximum values reaches two standard deviations (Fig. 1d). Hence, the IODMs of 1972, 1982, and 1997 are averaged to make the composite of the IODM during El Niño. For the composite of the IODM during non–El Niño years, the years 1961, 1967, and 1994 are averaged. The sample size is small as there are only a few major IODMs during this time period. Thus, we advise the reader to use discretion in interpreting the results from this composite analysis.

b. Mixed layer equation

To understand the variability of the mixed layer temperature, the physical processes that control the mixed layer temperature, such as the horizontal advection of mixed layer temperature, the supply of net heat fluxes to the mixed layer, and the entrainment of the lower water into the mixed layer, need to be examined. Specifically, in order to describe the entrainment process, the relative vertical velocity with respect to the varying mixed layer depth has to be estimated. One way to capture this relative vertical velocity is to transform the variables in z coordinates into a new coordinate system where the bottom of the varying mixed layer becomes a constant reference level (Wang et al. 1995). By assuming the consistency of the horizontal velocity and temperature within the mixed layer, and neglecting the shortwave radiation at the base of the mixed layer and the effect of diffusion, the mixed layer equation can be written as (Wang et al. 1995)
\[
\frac{\partial h_{\text{ML}}}{\partial t} + \nabla \cdot (h_{\text{ML}} \mathbf{V}_{\text{ML}}) = W_e 
\]

\[
\frac{\partial T_{\text{ML}}}{\partial t} = -\mathbf{V}_{\text{ML}} \cdot \nabla T_{\text{ML}} - \frac{W_e}{h_{\text{ML}}} H(W_e)(T_{\text{ML}} - T_e) + \frac{Q_o}{\rho_o C_w h_{\text{ML}}}. 
\]  

where the subscripts ML and e indicate the mixed layer and entrainment, so that \(T_{\text{ML}}\) and \(\mathbf{V}_{\text{ML}}\) denote the temperature and horizontal current, vertically averaged over the mixed layer depth, \(h_{\text{ML}}\). \(W_e\) and \(T_e\) are the entrainment velocity at the mixed layer base and the temperature of the entrained water, respectively (Fig. 2); and \(H(W_e)\) is a Heaviside function of the entrainment velocity. In addition, \(Q_o\) is the net downward heat flux at the ocean surface, \(\rho_o = 10^3 \text{ kg m}^{-3}\) is the density of the water, and \(C_w = 4.2 \times 10^7 \text{ J g}^{-1} \text{ K}^{-1}\) is the heat capacity of the water.

c. Linearized mixed layer equation

The mixed layer derived from Eqs. (1a) and (1b) can be linearized, since the variability of the mixed layer anomalies (\(h'_{\text{ML}}\)) is smaller than that of \(h_{\text{ML}}\) (figure not shown). The linearized equations are

\[
W'_e = \frac{\partial h'_{\text{ML}}}{\partial t} + \nabla \cdot (h'_{\text{ML}} \mathbf{V}_{\text{ML}}) + \nabla \cdot (h_{\text{ML}} \mathbf{V}_{\text{ML}}) 
\]

\[
\frac{\partial T'_{\text{ML}}}{\partial t} = -\mathbf{V}_{\text{ML}} \cdot \nabla T_{\text{ML}} - \mathbf{V}_{\text{ML}} \cdot \nabla T'_{\text{ML}} - \frac{W'_e}{h'_{\text{ML}}} H(W_e)(T_{\text{ML}} - T_e) + \frac{Q'_o}{\rho_o C_w h_{\text{ML}}}.
\]

Fig. 1. Normalized IODM index of (a) the El Niño year composite and (c) the non-El Niño year composite. Normalized Niño-3 SSTA index used to categorize (b) the El Niño year composite and (d) the non-El Niño year composite.
To estimate all terms in (2a) and the first four terms on the rhs of (2b), the data from the ocean analysis are used. The net heat flux anomalies \( Q'_o \) in the fifth term of (2b), however, is estimated not from the ocean analysis but from the ERA-40 data, since the net heat flux \( Q'_o \) in the ocean analysis is inaccurate. Most importantly, the use of climatological cloud cover in the ocean simulation eliminates the interannual variability of the surface solar radiation. Therefore, the heat fluxes derived from the ERA-40 data are used to describe the heat flux budget of the IODM. In the next section, the analysis on the linear approximation of the net heat flux [fifth term in (2b)] is followed by the linear estimation of the ocean processes [the first four terms on the rhs in (2b)] during the El Niño years.

\[
\frac{\partial}{\partial t} \text{IODM}_{\text{NETFL}} = \left( \frac{\text{LHF}' + \text{SHF}' + \text{SSR}' + \text{STR}'}{C_w \rho_o \bar{h}_{ML}} \right)_{\text{WTIO}} - \left( \frac{\text{LHF}' + \text{SHF}' + \text{SSR}' + \text{STR}'}{C_w \rho_o \bar{h}_{ML}} \right)_{\text{ETIO}}
\]

(4c)

In the above, the anomalies LHF’, SHF’, SSR’, and STR’ represent the variabilities of the latent heat flux, sensible heat flux, surface solar radiation, and surface thermal radiation; \( C_w \) and \( \rho_o \) are the heat capacity and the density of the water; the subscripts WTIO and ETIO indicate the area average over the western tropical Indian Ocean (10°S–10°N, 50°–70°E) and the eastern tropical Indian Ocean (10°S–0°, 90°–110°E), re-

4. Formation of the IODM in the El Niño years

a. Estimation of heat flux budget from the ERA-40 and Reynolds SST data

According to Saji et al. (1999), the Indian Ocean dipole mode (IODM) is defined as the difference in the SST anomalies between the tropical western Indian Ocean (10°S–10°N, 50°–70°E) and the tropical southeastern Indian Ocean (10°S–0°, 90°–110°E). That is, IODM = SST*(10°S–10°N, 50°–70°E)

\[-\text{SST}*(10°S–0°, 90°–110°E).
\] (3)

The composite of the IODM from the Reynolds SST data for El Niño years shows a maximum from October to November associated with a warming (cooling) of the western (eastern) Indian Ocean (Fig. 3a). The tendency of the IODM (\( \partial \text{IODM}/\partial t \); Fig. 3b) indicates that a persistent forcing of this positive IODM exists from January to October. By using the heat fluxes from the ERA-40 data, the contribution from the surface latent heat fluxes, and sensible heat fluxes, and the net surface solar and thermal radiation, on the tendency of the IODM (\( \partial \text{IODM}/\partial t \)) is estimated. That is, Effect of latent and sensible heat fluxes (positive, into ocean) on the formation of dipole mode:

\[
\frac{\partial}{\partial t} \text{IODM}_{\text{LHF-SHF}} = \left( \frac{\text{LHF}' + \text{SHF}'}{C_w \rho_o \bar{h}_{ML}} \right)_{\text{WTIO}}
\]

(4a)

Effect of surface solar radiation and surface thermal radiation (positive, into ocean):

\[
\frac{\partial}{\partial t} \text{IODM}_{\text{SSR-STR}} = \left( \frac{\text{SSR}' + \text{STR}'}{C_w \rho_o \bar{h}_{ML}} \right)_{\text{WTIO}}
\]

(4b)

Effect of net heat flux (positive, into ocean) on the formation of dipole mode:

\[
\frac{\partial}{\partial t} \text{IODM}_{\text{NETFL}} = \left( \frac{\text{LHF}' + \text{SHF}' + \text{SSR}' + \text{STR}'}{C_w \rho_o \bar{h}_{ML}} \right)_{\text{WTIO}} - \left( \frac{\text{LHF}' + \text{SHF}' + \text{SSR}' + \text{STR}'}{C_w \rho_o \bar{h}_{ML}} \right)_{\text{ETIO}}
\]

(4c)
spectively. The term $h_{ML}$ denotes the climatology of the monthly mean mixed layer depth derived from the ocean analysis (see the next section).

The positive forcing of latent and sensible heat fluxes on the development of the IODM is found in spring and fall (Fig. 3b), while the negative forcing of surface solar and thermal radiation (Fig. 3b) hampers the positive effect of the latent and sensible heat fluxes. The positive feedback of the latent and sensible heat fluxes, and the damping effect of the surface solar and thermal radiations are consistent with the results of previous studies (Lau and Nath 2003; Annamalai et al. 2003; Li et al. 2003; Shinoda et al. 2004a,b). However, the relatively large damping effect of the surface solar and thermal radiations is noted in this study. Considering that the interannual variability of the high cloud in most parts of the world in the ERA-40 data is not reliable before January 1979 (Chevallier et al. 2005), the large damping effect of the surface solar and thermal radiation in this study requires further verification with more robust data in the future.

b. Estimation of mixed layer heat budget from the ocean analysis

The mixed layer depth ($h_{ML}$) is defined as the depth whose density difference from the surface is closest to 0.01 kg m$^{-3}$ (Jackett and McDougall 1997). The linearized oceanic processes in (2b), such as the horizontal advection ($-V_{ML} \cdot \nabla T_{ML} - \nabla_{ML} \cdot \nabla T_{ML}$) and entrainment [$-(W_e (T_{ML} - T_e)/h_{ML})$, $- (W_e (T_{ML} - T_e)/h_{ML})$], are calculated from the ocean analysis. The sum of these linearized terms is then compared with the tendency of the IODM from the time series of $T_{ML}$ in order to assess the relative importance of the linearized ocean processes on the formation of the IODM (Fig. 4a). Furthermore, the anomalous ocean processes, which include both linear and nonlinear processes, are

![Diagram](image-url)
estimated by 1) calculating each term of the ocean processes (such as horizontal advection and entrainment) from the monthly mean of the ocean analysis, 2) obtaining a climatological annual cycle of each term by making an average of each month over 41 yr, and 3) subtracting the climatological annual cycle of each term from its monthly mean value.

Two local maxima in the forcing of the IODM, calculated from the time series of $T_{ML}$, are found in March and September (Fig. 4a). In those months, the oceanic forcings of the IODM also become maxima and comparable to the tendency of the IODM. This implies that the linearized ocean processes contribute to the formation of the IODM in early spring and late summer. From April to July, however, the forcing of the IODM cannot be explained in terms of oceanic processes. In those months, the nonlinear interaction in the coupled ocean–atmosphere dynamics may become important. Nevertheless, the role of the linearized oceanic processes during the formation of the IODM is discernible and can be explained by oceanic advection and entrainment in (2b).

Shown in Fig. 4b is the net effect of the linearized oceanic process, as well as the individual contributions to the IODM tendency. The tendency of the IODM, induced by the linear approximation, exhibits maxima in both March and September. While the entrainment plays an important role in early spring and late summer, the effect of the meridional advection increases throughout the summer and becomes as large as that of the entrainment by September (Fig. 4b).

Although the effect of zonal advection seems to be marginal on the formation of the IODM (Fig. 4b), its spatial structure is important in understanding the temperature variability of the western Indian Ocean. For example, in August of El Niño years, the anomalous westward current in the northwestern Indian Ocean advects the warm climatological mixed layer temperature of the central tropical Indian Ocean (Fig. 5a). Thus, the advection of mean mixed layer temperature by the
anomalous zonal current \([-u'(\partial T_{ML}/\partial x) > 0]\) contributes to the increase of \(T_{ML}\) in the WTIO (Fig. 5b). Since this positive zonal advection is found not only in the WTIO but also in the ETIO, its contribution to the IODM \([-u'(\partial T_{ML}/\partial x)]_{WTIO} - [-u'(\partial T_{ML}/\partial x)]_{ETIO}\) becomes trivial.

The structure of the anomalous meridional advection \([-v'(\partial T_{ML}/\partial y) > 0]\) is also examined in Fig. 5. When a Rossby wave responds to the cooling of the ETIO in September (Fig. 5c), the anomalous northward (southward) current in the eastern (central) part of the southern tropical Indian Ocean (10°–2°S, 70°–100°E) advects the cold (warm) climatological mixed layer temperature from the south (north) (Fig. 5c). As a result, the cold (warm) advection becomes dominant in the southwestern (central) Indian Ocean (Fig. 5d). The anticyclonic circulation in the northern tropical Indian Ocean (2°–10°N, 75°–95°E; Fig. 4c) also produces the cold (warm) advection in the northeastern (central) Indian Ocean (Fig. 5d).

It was shown in Fig. 4b that the contribution of the entrainment on the IODM is strongest in March and September. In March, the anomalous entrainment of the mean vertical temperature gradient \([-[W_e(T_{ML} - \bar{T})/h_{ML}] < 0]\) is negative in the southeastern Indian Ocean, from 10°–4°S to 92°–110°E (Fig. 6a). The average of this anomalous entrainment over the reference area of the ETIO (10°S–0°, 90°–110°E) is about \(-0.07 \, ^\circ \text{C month}^{-1}\) in March. Since the contribution of the entrainment to \(\partial \text{ODM}/\partial t\) is about \(0.15 \, ^\circ \text{C month}^{-1}\) in March (Fig. 4b), the anomalous entrainment cooling in the ETIO contributes half of this value. The other half (0.08 \, ^\circ \text{C month}^{-1}) comes from the warming of the WTIO by anomalous detrainment (Fig. 6a). The contribution of this anomalous detrainment is the area-averaged value of \([-[W_e(T_{ML} - \bar{T})/h_{ML}]]\) between
10°S–10°N and 50°–70°E. Thus, it does not represent the complicated spatial structure of the detrainment in the western Indian Ocean. The local convergence (divergence) of the anomalous current, especially in the meridional direction, seems to be responsible for the local downwelling (upwelling) and, eventually, the detrainment (entrainment) in March.

In September, the spatial structure of the anomalous entrainment and detrainment is rather simple (Fig. 6b). Most importantly, there is cooling along the coastline of Sumatra in the ETIO and a general warming of the WTIO, except for some small-scale areas. The resultant contribution of the anomalous entrainment (detrainment) in the ETIO (WTIO) is about 0.12°C month⁻¹ (0.06°C month⁻¹) in September.

It should be noted that the temperature change associated with the detrainment (entrainment) along the coastline of Somalia (Sumatra) is smaller than that expected by coastal downwelling (upwelling) \[ h_{ML} (\nabla \cdot \vec{V}_{ML}) + \vec{h}_{ML} (\nabla \cdot \vec{V}_{ML}) \]. The discrepancy between the detrainment (entrainment) and downwelling (upwelling) in this region might be caused by the exclusion of the nonlinear process, or simply by the fact that the surge of cold water, induced by the upwelling, does not always penetrate into the mixed layer and affect the mixed layer temperature [see Eq. (2a)]. Since, the entrainment velocity in (2a) is the relative velocity with respect to the mixed layer bottom, which also changes in time \( \partial h_{ML}/\partial t \), the effect of the entrainment and detrainment in the linear approximation seems often smaller than that of the upwelling and downwelling.

c. Characteristics of the IODM in El Niño years

Based on the previous analysis of the ocean and atmosphere, the characteristics of the IODM during El Niño years are identified. The first feature, which induces the warming along the eastern coast of Somalia–Arabian Peninsula, is the easterlies and northeasterlies in the region between the northern Arabian Sea and the equator (Figs. 7b and 7c). These easterlies and northeasterlies induce a westward component in the ocean current, which advects the climatologically warm SST of the central Indian Ocean to the western Indian Ocean (Figs. 5a and 5b). As a result, the first sign of the warming in the western Indian Ocean is found at latitudes between 5°S and 15°N. The second feature of the IODM is the development of the southeasterlies and southerlies in the southeastern Indian Ocean from late spring to fall (Figs. 7b and 7c). The joined force of these anomalies and mean winds along the coast of Sumatra enhances the latent and sensible heat fluxes, cold advection \( -v_{ML} (\partial T_{ML}/\partial y) < 0 \), and entrainment in the southeastern Indian Ocean.

The major atmospheric circulation of the IODM, such as easterlies and northeasterlies (southerlies and southeasterlies) in the northwestern (southeastern) Indian Ocean, seems to be related to the anticyclonic circulation, which is located in the northern (southern) Indian Ocean (Figs. 7f and 7g). As the local maximum geopotential anomaly develops poleward of 10°N (10°S) from 50° to 90°E (from 80° to 130°E), the accompanying two anticyclonic circulations result in the northeasterlies and easterlies over the tropical and the northwestern Indian Ocean, and southeasterlies and southerlies in the southeastern Indian Ocean (Figs. 7f and 7g). This suggests that the development of two anticyclonic circulations has an effect on the basic characteristics of the IODM during El Niño years.

It turns out that these two anticyclonic circulations form one of the signatures of El Niño, itself. When the

![Fig. 6. Temperature forcing (°C month⁻¹) induced by the anomalous entrainment velocity (cm s⁻¹) acting on the mean vertical temperature gradient \(- \left[ W(T_{ML} - T_{e})/\partial ML \right]\) for (a) March and (b) September of El Niño years.](image_url)
lagged cross correlations of the 850-hPa wind and SST anomalies are calculated with respect to the Niño-3 SSTA (Fig. 8), the development of the anticyclonic circulation in the northern Indian Ocean (5°–25°N, 50°–100°E) and southern Indian Ocean (0°–30°S, 50°–100°E) is detected as early as 6 months before the mature phase of an El Niño (Fig. 8b). Associated with these two anticyclonic circulations are the easterlies and northeasterlies over the western Tropical Indian Ocean (10°S–10°N, 50°–70°E), and the southerlies and southeasterlies over the southeastern Indian Ocean (Fig. 8c). The presence of the IODM-like features in the lagged cross-correlation map implies that the development of IODM is indeed related to El Niño events.

5. Formation of the IODM in the non–El Niño years

It is suggested in the previous section that the development of the easterlies and northeasterlies (southerlies and southeasterlies) over the western (eastern) In-
The characteristics of the IODM in non–El Niño years are different from those in El Niño years in three respects. First, the dominant easterlies and northeasters, often found in the northwestern Indian Ocean during the El Niño years, are absent or weak (Fig. 9). Instead, the westerlies and southwesterlies become dominant in the northwestern Arabian Sea in summer (Fig. 9c). Consequently, the northwestern Indian Ocean no longer experiences warming (Figs. 9c and 9d), as was the case during El Niño years (Figs. 7c and 7d). This leads to the second difference of the IODM in non–El Niño years. That is, the cooling of the ETIO becomes a much more dominant component than the warming of the WTIO in the IODM during non–El Niño years. This result is similar to that found in the Geophysical Fluid Dynamics Laboratory (GFDL) coupled climate model (Song et al. 2007), in the sense that the WTIO is warmer during El Niño years than in non–El Niño years.

The third difference can be found in the geopotential field in spring (Fig. 9f) and summer (Fig. 9g). While the existence of maximum geopotential anomalies in the southeastern Indian Ocean remains unchanged from El Niño years, the geopotential anomalies along 20°N change from a local maximum in El Niño years (Figs. 7f and 7g) to a local minimum in non–El Niño years (Figs. 9f and 9g). Consequently, the northwestern and southeastern Indian Ocean regions experience monsoonlike wind anomalies (Fig. 9g), enhancing the southerlies and southeasterlies over the ETIO.

b. Estimation of heat flux budget from the ERA-40 and Reynolds SST data

In the previous section, the differences in the spatial structure of the IODM between El Niño and non–El Niño years are described in terms of winds, geopotential, and SST anomalies. These differences affect the atmospheric heat budget, so that the net atmospheric heat flux forcing is now positive through June (Fig. 10b). During this part of non–El Niño years, the enhanced latent heat flux overcomes the damping effect of the surface solar and thermal radiations and contributes to the early development of the IODM (Fig. 10b).

To illustrate the spatial structure of this feature, the latent heat flux anomalies are averaged from February
to June (FMAMJ mean), and presented for El Niño (Fig. 11a) and non–El Niño (Fig. 11b) years. The averaged effect of the latent heat flux on the IODM from February to June is much stronger during non–El Niño (Fig. 11b) than El Niño (Fig. 11a) years. The difference between the El Niño and non–El Niño composites (El Niño minus non–El Niño years) for the latent heat flux (Fig. 11c) and the net heat flux (Fig. 11d) further confirms that the initial effect of net heat flux on the formation of the IODM is stronger in non–El Niño years. As the year progresses, however, the damping effect of the surface solar and thermal radiations cancels out the positive feedback of the sensible and latent heat fluxes in both El Niño (Fig. 3b) and non–El Niño years (Fig. 10b). Consequently, an analysis of the oceanic processes is required in order to examine the rest of the development of the IODM through summer and fall.

c. Estimation of mixed layer heat budget from ocean analysis

The comparison between the tendency of the IODM and the effect of the linearized ocean processes indicates that the linearized ocean processes before (after) July are marginal (dominant) in the forcing of the
IODM. One of the possible causes of this discrepancy is the increased involvement of the atmospheric net heat flux on the formation of the IODM until June (Fig. 10b). The impact of the linearized ocean processes, on the other hand, is maximum in September (Fig. 12a).

Since the atmospheric winds in non–El Niño and El Niño years are different, the formation of the IODM relies on different oceanic processes. The first difference is the negative effect of the entrainment in the early months of the year (Fig. 12b). The northwesterlies along the coast of Sumatra in those months (Fig. 9a) induce the downwelling in the ETIO. The other difference is the increased effect of the meridional advection compared to other terms in September (Fig. 12b). This is due to the monsoonalike atmospheric circulation, which enhances the meridional component of the ocean currents, thereby increasing the meridional advection. The spatial comparison of the temperature anomalies between El Niño and non–El Niño years (El Niño minus non–El Niño) during this season is shown (Fig. 13a) with the oceanic processes, such as zonal (Fig. 13b) and meridional (Fig. 13c) advections, and the entrainment (Fig. 13d). In summer and fall (JAS mean), the maximum difference in the mixed layer temperature is found in the northwestern and the southeastern Indian Ocean regions (Fig. 13a). The positive maximum in the northwestern Indian Ocean is due to the stronger warming during El Niño years (Fig. 7c), while the other positive maximum found in the southeastern Indian Ocean (Fig. 9c) is caused by the stronger cooling during non–El Niño years. The warmer region in the northwestern Indian Ocean in El Niño years is caused by the warmer zonal advection (Fig. 13b), while the colder region of the southeastern Indian Ocean during non–El Niño years is induced by the colder meridional advection (Fig. 13c) and the entrainment (Fig. 13d).
The relative importance between the surface heat flux anomalies and the dynamical oceanic response during El Niño and non–El Niño years has been previously studied by Shinoda et al. (2004a,b). They suggested that surface heat flux anomalies play an important part in the eastern Indian Ocean, while the dynamical oceanic response seems to govern the variability in the western Indian Ocean (Shinoda et al. 2004a) in El Niño years. During non–El Niño years, however, the dynamical oceanic response in the eastern Indian Ocean becomes crucial during the formation of the IODM, while the involvement of the oceanic processes decreases in the western Indian Ocean (Shinoda et al. 2004b). Our study extends the work of Shinoda et al. by providing the reason for these changes. Based on the quantitative analysis of the oceanic mixed layer heat budget, changes in the major components during El Niño and non–El Niño years are explained as follows.

In El Niño years, two off-equatorial, anticyclonic circulations develop, associated with the increased pressure over the eastern Indian Ocean. The anticyclonic circulation in the Northern Hemisphere enhances the easterly component of the winds in the northwestern Indian Ocean. This enhanced easterly component increases the mixed layer temperature in the northwestern Indian Ocean by inducing an anomalous westward ocean current that advects the warm mean mixed layer from the central to the western Indian Ocean. Meanwhile, the anticyclonic circulation over the southeastern Indian Ocean strengthens the southeasterlies, thereby causing oceanic meridional and vertical advection of the cold mean temperature.

While Shinoda et al. (2004b) suggested that the ENSO-induced surface dipole is primary controlled by surface heat fluxes, we found that the oceanic processes, such as meridional advection and entrainment, are as important as the surface heat fluxes. They also suggested that when the IODM is independent of ENSO, subsurface variations play an important role, especially in the eastern Indian Ocean where the strong

Fig. 11. The contribution of the latent heat flux on the tendency of the mixed layer temperature anomalies in (a) El Niño and (b) non–El Niño years. The difference in the contribution of (c) the latent heat flux and (d) the net heat flux between El Niño and non–El Niño years. Dotted line indicates negative value. (Units: K month$^{-1}$.)
surface cooling in late summer is generated by upwelling and horizontal heat advection in response to basin-wide surface easterlies. In addition to the enhanced contribution of the meridional advection and entrainment, as well as the evaporative cooling due to strengthened southeasterlies in the southeastern Indian Ocean, we also found that the reduction of the easterlies over the northwestern Indian Ocean during non–El Niño year hampers the contribution of the anomalous zonal advection to the warming of the northwestern Indian Ocean. This explains why the dynamical oceanic response decreases (increases) in the northwestern (southeastern) Indian Ocean from El Niño to non–El Niño years.

6. Summary and discussion

The Indian Ocean dipole mode (IODM) is examined in a series of composites using data from the ERA-40, the Reynolds SST dataset, and an ocean analysis using the Modular Ocean Model (MOM). The main purpose of this study is to not only help understand the formation of the IODM, but to also compare the differences in the IODM during El Niño and non–El Niño years. The differences in the IODM during El Niño and non–El Niño years are found in the spatial structure of the wind and geopotential anomalies. In El Niño years, the development of two off-equatorial, anticyclonic circulations is associated with easterlies and northeasterlies (southerlies and southeasterlies) over the northwestern (southeastern) Indian Ocean. One of the contributing factors of the warm WTIO in El Niño years is the anomalous easterlies and northeasterlies, which induce ocean currents that advect the warm mean mixed layer of the central Indian Ocean toward the western Indian Ocean. In addition, this westward current may decrease the Somalia–Arabian Peninsula upwelling.
thereby further enhancing the warming in the northwestern Indian Ocean. Meanwhile, the cooling of the southeastern Indian Ocean from spring to late summer is caused by the atmospheric southerlies and southeast-erlies, which increase the latent heat flux, the entrainment, and the meridional advection in the ETIO.

Unlike the case in El Niño years, the geopotential field in non–El Niño years is antisymmetric with respect to the equator. Thus, the resultant wind anomalies in the northwestern and southeastern Indian Ocean regions are similar to a monsoonlike circulation. In other words, the westerlies and southwesterlies (southerlies and southeasterlies) are intensified over the northwestern (southeastern) Indian Ocean. Thus, the cold zonal (cold meridional and vertical) advection is enhanced in the northwestern (southeastern) Indian Ocean. In addition, the anomalous winds over the southeastern Indian Ocean are the same sign as the climatological monthly mean winds. Therefore, the anomalous latent and sensible heat fluxes further contribute to the cooling of the eastern Indian Ocean, especially during the early part of non–El Niño year. Consequently, the cooling of the ETIO, rather than the warming of the WTIO, dominates the IODM in non–El Niño years.

The significance of this study is that the IODM can be induced by not only El Niño–related winds, but also by the locally enhanced monsoon-type circulation. However, the IODM under these two types of wind anomalies may evolve into different spatial structures. Particularly, the most distinct difference can be found in the SST anomalies in the western Indian Ocean. Since the western Indian Ocean is an important moisture source for the Indian summer monsoon, the different SST anomalies in this region may have varying degrees of influence on the Indian monsoon.

According to Loschnigg et al. (2003), the warm SST anomalies of the IODM, which are developed as a part of the ENSO–monsoon system and are enhanced through the anomalous heat transport, can persist through the winter season and contribute to the development of a stronger than normal monsoon during the following summer. Within this context, the weak SST anomalies of the western Indian Ocean during non–El Niño year may have less chance of survival and con-
tributing to the following monsoon year. For example, the anomalous summer precipitation (JJA mean) in the Indian monsoon region (10°–25°N, 70°–100°E) is positive in 1973, 1983, and 1998, following the positive IODM of El Niño years (figure not shown). In the years 1962 and 1968, however, the precipitation anomalies are negative in spite of the IODM during non–El Niño years (figure not shown). Since the connection between the warm SST in the Indian Ocean and the strong monsoon variability of the following year is an essential part of the tropical biennial oscillation (Li et al. 2003; Loschnigg et al. 2003), the IODM during non–El Niño years may not grow into the self-sustained mode of the tropical biennial oscillation.

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