Interdecadal Indian Ocean Basin Mode Driven by Interdecadal Pacific Oscillation: A Season-Dependent Growth Mechanism

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ABSTRACT

The interdecadal variability of basinwide sea surface temperature anomalies (SSTAs) in the tropical Indian Ocean (TIO), referred to as the interdecadal Indian Ocean basin mode (ID-IOBM), is caused by remote forcing of the interdecadal Pacific oscillation (IPO), as demonstrated by the observational datasets and tropical Pacific pacemaker experiments of the Community Earth System Model (CESM). It is noted that the growth of the ID-IOBM shows a season-dependent characteristic, with a maximum tendency of mixed layer heat anomalies occurring in early boreal winter. Three factors contribute to this maximum tendency. In response to the positive IPO forcing, the eastern TIO is covered by the descending branch of the anomalous Walker circulation. Thus, the convection over the southeastern TIO is suppressed, which increases local downward shortwave radiative fluxes. Meanwhile, the equatorial easterly anomalies to the west of the suppressed convection weaken the background mean westerly and thus decrease the upward latent heat fluxes over the equatorial Indian Ocean. Third, anomalous westward Ekman currents driven by the equatorial easterly anomalies advect climatological warm water westward and thus warm the western TIO. In summer, the TIO is out of the control of the positive IPO remote forcing. The ID-IOBM gradually decays due to the Newtonian damping effect.

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1. Introduction

The Indian Ocean basin mode (IOBM) is one of two dominant modes of sea surface temperature anomalies (SSTAs) in the tropical Indian Ocean (TIO) (Klein et al. 1999; Huang and Kinter 2002; Krishnamurthy and Kirtman 2003; Yamagata et al. 2004; Schott et al. 2009), and the other is the Indian dipole mode (IOD; Saji et al. 1999, Webster et al. 1999). Conventionally, the IOBM refers to a phenomenon on an interannual time scale (hereafter IA-IOBM), which is primarily driven by the remote forcing from El Niño–Southern Oscillation (ENSO) (Latif and Barnett 1995; Klein et al. 1999; Xie et al. 2002, 2009; Alexander et al. 2002; Chiang and Sobel 2002; Saji and Yamagata 2003; Lau and Nath 2004; Liu and Alexander 2007; Lee et al. 2015).

The evolution of the IA-IOBM is both season-dependent (Du et al. 2009; Roxy et al. 2011) and associated with the phase of ENSO. Specifically, the IA-IOBM tends to form during the mature El Niño winter. The remote forcing from the equatorial central–eastern Pacific drives the IA-IOBM by modulating the surface shortwave radiative fluxes and latent heat fluxes (Klein et al. 1999; Wu and Kinter 2010). The IA-IOBM can maintain itself in the following spring through a positive wind–evaporation–SST feedback as a response to an antisymmetric TIO heating pattern along the equator (Du et al. 2009). In a decaying El Niño summer, although warm SSTAs in the equatorial central–eastern Pacific evolve to neutral states or even cold SSTAs, the IA-IOBM is still maintained [Du et al. 2009; Wu et al. 2010; also see Xie et al. (2016) for a review] and plays a key role in modulating East Asian–western North Pacific monsoons [Xie et al. 2009; Wu and Zhou 2008; Wu et al. 2009, 2010, 2012; also see Li et al. (2017) for a review].

It was found that the IOBM also existed on the interdecadal time scale, with a spatial pattern that was similar to that of the IA-IOBM (hereafter ID-IOBM; Han et al. 2014a; Dong and McPhaden 2017b; Tozuka et al. 2007; Zhang et al. 2018). The ID-IOBM is the leading mode of interdecadal variability of the SSTAs in the TIO (variance contribution >50%) based on the HadISST1.1 dataset (Han et al. 2014b). These features were also seen in three other observational SST datasets (Dong and McPhaden 2017b; Zhang 2016) and an atmosphere–ocean coupled model (Tozuka et al. 2007). The ID-IOBM was regarded as a passive response to remote forcing from the interdecadal Pacific oscillation (IPO; Mantua et al. 1997; Power et al. 1999; Newman et al. 2016; Henley et al. 2017). The IPO drives the ID-IOBM through an atmospheric bridge, which modulates the surface heat flux and thermocline depth in the TIO (Dong et al. 2016). However, the detailed ocean dynamic effects on the formation of the ID-IOBM remain inconclusive.

Though the ID-IOBM is correlated with the IPO, their relationship does not seem stationary. It was reported that the ID-IOBM index was positively (negatively) correlated with the IPO index before (after) 1985 (Han et al. 2014a,b). Several possibilities have been proposed to explain the negative correlation relationship after 1985. These include the following: 1) The rapid warming of the TIO SST due to the external forcing since the mid-1990s cannot be eliminated completely through the conventional method of removing the global mean SST. The enhanced effect of external forcing overwhelmed the effect of the IPO on the evolution of the ID-IOBM after 1985 (Dong and McPhaden 2017b; Zhang et al. 2018). 2) The TIO SSTAs can be influenced by the Atlantic multidecadal oscillation (AMO) (Li et al. 2016; Kucharski et al. 2016). 3) The warming in the TIO has been stronger than that in the Pacific and Atlantic since 1950 (Arora et al. 2016; Du and Xie 2008; Dong and Zhou 2014; Dong and McPhaden 2016, 2017a), implying that the role of the Indian Ocean changed from passive to active in the Indo-Pacific coupled system (Luo et al. 2012).

However, there are limited studies focusing on the mechanisms responsible for the formation of the ID-IOBM, largely due to the scarcity of reliable long-term observational datasets in the TIO. In this study, the pacemaker experiments are conducted by using the Community Earth System Model, in which the SSTAs in the equatorial central–eastern Pacific are restored to the observations. The model results provide us with reliable long-term data to investigate the mechanisms responsible for the formation of the ID-IOBM driven by the IPO. We find that the amplitude of the ID-IOBM shows a typical seasonally evolving feature; that is, its intensity in boreal early spring is stronger than other seasons. This feature is caused by a season-dependent growth mechanism of the ID-IOBM.

The rest of the paper is organized as follows. Datasets, methods, and model experiments used in this study are described in section 2. Section 3 analyzes the annual-mean features of the ID-IOBM and its relationship with the IPO. Section 4 investigates the seasonal evolutions of the ID-IOBM intensity and related mechanisms. Finally, section 5 summarizes the key findings of this study and provide a discussion.

2. Datasets, model experiments, and methods

a. Observational datasets

The following four gridded observational SST datasets are used: 1) the Hadley Centre Sea Ice and Sea Surface Temperature dataset, version 1.1 (HadISST1.1; $1^\circ \times 1^\circ$; Rayner et al. 2003); 2) the National Oceanic and

FIG. 1. Spatial distributions of the ID-IOBM derived from three observational datasets, (a) HadISST1.1, (b) ERSST5, and (c) Kaplan2, and (d) CESM results. They are calculated through regressing preprocessed SSTAs onto their corresponding ID-IOBM indices. (e) The annual-mean ID-IOBM indices derived from the ensemble mean of three observational datasets (blue solid line), CESM results (black line), and the individual observational datasets (blue dashed lines). The ID-IOBM index is defined as area-averaged SSTAs in the TIO (25°N–30°S, 40°–110°E). The IPO index derived from the HadISST1.1 dataset is also shown (red line).

Atmospheric Administration Extended Reconstructed SST, version 5 (ERSST5; $2^\circ \times 2^\circ$; Huang et al. 2017); 3) the Kaplan Extended SST, version 2 (Kaplan2; $5^\circ \times 5^\circ$; Kaplan et al. 1998); and 4) the Hadley Centre SST, version 3 (HadSST3; $5^\circ \times 5^\circ$; Kennedy et al. 2011a,b). All the datasets cover the period 1900–2012.

b. Model and experimental designs

We used the coupled general circulation model (CGCM) Community Earth System Model (CESM), version 1.20, in this study. The CESM was developed by the National

Center for Atmospheric Research (NCAR) (Hurrell et al. 2013). Its atmospheric component is the Community Atmosphere Model (CAM), version 4, with 26 vertical levels and a horizontal resolution of 1.25° longitude $\times$ 0.94° latitude (Neale et al. 2013; Lamarque et al. 2012). Its oceanic component is an extension of the Parallel Ocean Program (POP), version 2, with 60 vertical layers and an irregular horizontal resolution (approximately $0.5^\circ \times 0.5^\circ$) (Smith et al. 2010).

Two sets of numerical experiments were conducted. The first is a historical climate simulation (hereafter

FIG. 2. (a) Spatial distribution of the ID-IOBM-related annual-mean SSTAs (K) over the TIO from the CESM results. Also shown are the contributions of (b) ocean heat transport effect (K), (c) net surface shortwave radiative flux (K), and (d) latent heat flux due to the atmosphere (K). The energy units have been converted to units of temperature (K) based on Eq. (2).

HIST), in which the model was driven by the historical radiative forcing for 1850–2005 and the representative concentration pathway 4.5 (RCP4.5) for 2006–40, based on phase 5 of the Coupled Model Intercomparison Project 5 (CMIP5; Taylor et al. 2012). The second is a pacemaker experiment (hereafter HIST-IPO). The SST in the tropical central-eastern Pacific (20°N–20°S, 175°E–75°W) was restored to the model's climatological mean plus observed anomaly, but the oceanic and atmospheric components were freely coupled in other ocean areas. The restoring time scale was set to 10 days. A buffer zone (zonal and meridional ranges are both 5°) was built between the inner and outer boxes, with the restoration weight linearly reduced from one to zero (Boer et al. 2016; Zhou et al. 2016). In the

HIST-IPO, the radiative forcing is identical to HIST. Both the HIST and HIST-IPO cover the period of 1900–2012 and consist of eight independent runs starting from different initial conditions. The SST data used as observational evidence for restoration are those of HadISST1.1.

c. Data preprocessing

In this study, we focus on the interdecadal variability internally generated in the climate system. Hence, all the data have been preprocessed before the analysis through the following two steps. First, the interannual variability was filtered out through the 8-yr low-pass Lanczos filter (Duchon 1979). Second, signals associated with external radiative forcing were removed.

FIG. 3. Seasonal evolutions of the intensity of 3-month running-mean ID-IOBM indices (K) derived from the ensemble mean of four observational datasets (HadISST1.1, ERSST5, Kaplan2, and HadSST3; blue solid line), the CESM results (black solid line), and the individual observational datasets (blue dashed lines). For each dataset, the seasonal ID-IOBM intensity is calculated by regressing seasonal-mean area-averaged TIO SSTAs onto the corresponding normalized annual-mean ID-IOBM index shown in Fig. 1e. The red line denotes the seasonal evolution of tendency of the ID-IOBM intensity in the CESM results (K month^{-1}). Values in the upper left and bottom-right corners are variation ranges (maximum minus minimum).

For the observational datasets, the external forcing signals were represented by the global mean low-pass-filtered SSTAs. The low-pass filtered observational fields were first regressed onto the global-mean SSTAs time series, and the residual terms of the regression analyses are used in this study (Ting et al. 2009). For the CESM data, the external forcing signals were represented by the multimember ensemble mean of the HIST (hereafter HIST_ENS) (Wang et al. 2017). It is estimated that the SST variance in the tropical central-eastern Pacific (20°N – 20°S , 175°E – 75°W) in the HIST_ENS is only approximately 10.0%–12.6% of that in the individual members, suggesting that the internal interdecadal variability has been smoothed out in the HIST_ENS. The differences between the multimember ensemble mean of the HIST-IPO (hereafter HIST-IPO_ENS) and the HIST_ENS are used in this study, which represents pure IPO-related climate variability. For simplicity, the differences between two experiments are referred to as the CESM results hereafter without further explanation.

d. Definitions of the ID-IOBM and IPO

The ID-IOBM index is defined as the annual-mean SSTAs averaged over the TIO (25°N – 30°S , 40° – 110°E) from the preprocessed data. It is worth noting that although the ID-IOBM is defined based on the filtered SSTAs, it is definitely a physical mode. There are interdecadal spectral peaks at the period of 20–30 years in the spectral analyses of the raw SSTAs in the TIO for both the observational data and the CESM results (figure not shown). The ID-IOBM-related patterns are obtained through regressing variable anomalies onto the

ID-IOBM index. The IPO is defined as the first EOF mode of the annual-mean SSTAs in the Pacific (60°N – 40°S , 100°E – 70°W) after the data preprocessing. The corresponding normalized principal component time series is taken as the IPO index.

e. Significance test

We used a nonparametric method, referred to as the random-phase test, to test the statistical significances of correlation and regression analysis (Ebisuzaki 1997; Wu et al. 2016). For instance, we tested the significance of the correlation between time series A and B [$r(A, B)$]. The core of the random-phase test is to construct N time series, all of which have the same power spectrum but involve different temporal phases with the time series A . Then, the significance level of $r(A, B)$ can be estimated through comparing $r(A, B)$ with the correlations of all these N constructed time series with time series B . More detailed descriptions of the method can be found in Ebisuzaki (1997).

3. Temporal and spatial characteristics of the ID-IOBM

Spatial patterns of the ID-IOBM from the three observational SST datasets all show a basinwide warming in the TIO, with the maximum SSTAs centers located in the southwestern TIO (Figs. 1a–c). However, the detailed features and intensities of the ID-IOBM are different among the three observational SST datasets because reliable observational records were extremely sparse in the TIO before the 1950s (Han et al. 2014b; Zhang 2016).

FIG. 4. Spatial distributions of ID-IOBM in (a) JJA, (b) ASO, (c) NDJ, and (d) FMA derived from the CESM results, which are calculated by regressing the seasonal-mean SSTAs onto the annual-mean ID-IOBM index (K). White dots represent values reaching the 5% significance level.

The CESM results reproduce the major features of ID-IOBM in the observations, such as the basinwide warming pattern and maximum SSTAs centered in the southwestern TIO (Fig. 1d). The major discrepancies of simulations are that 1) the asymmetry of SSTA intensity between the northern and southern TIO is weaker than that in the observations, 2) the maximum SSTAs center is shifted equatorward by approximately 5° – 10° , and 3) the warm SSTAs in the far southeastern Indian Ocean is absent, as had been mentioned in previous studies (Yang et al. 2015; Zhang et al. 2018).

The temporal evolutions of ID-IOBM in three observational datasets are highly consistent (Fig. 1e). Their ensemble mean is simultaneously correlated with the observational IPO index (Fig. 1e). For the period of 1900–2012, the correlation coefficient with the IPO index is 0.49. It is noted that the correlation varies with

time. Specifically, the correlation coefficient with the IPO index changes from positive (0.57) over the period of 1900–1990 to negative (-0.73) after 1990. The change in the relationship between ID-IOBM and IPO in the observations was attributed to the fact that the conventional regression method cannot cleanly separate interdecadal internal variability from the external forcing signals (Dong and McPhaden 2017b; Zhang et al. 2018). In contrast, the evolution of ID-IOBM index in the CESM results is more synchronous with that of the IPO index, with the correlation coefficient reaching 0.76 over the period of 1900–2012. This is because the external forcing signals, which are represented by the HIST_ENS, have been cleanly removed from the HIST-IPO_ENS.

The mechanisms for the formation of ID-IOBM can be understood through the following approach. The

FIG. 5. Spatial distributions of the ID-IOBM-related SSTA tendency in (a) NDJ and (b) JJA derived from the CESM results, which are calculated by regressing seasonal-mean SSTA tendencies onto the annual-mean ID-IOBM index (K month^{-1}). White dots represent values reaching the 5% significance levels. The position of the purple box in (a) is 3°N – 15°S , 40° – 100°E .

mixed layer temperature (MLT) tendency equation can be written as

$$C \frac{\partial T'}{\partial t} = D'_o + Q'_{\text{net}}, \quad (1)$$

where T' is the MLT anomaly converted to the energy unit, Q'_{net} denotes the anomalous net surface heat flux, and D'_o denotes the anomalous ocean dynamic effects due to three-dimensional advection and mixing. Note that Q'_{net} consists of the anomalous net surface shortwave radiative flux Q'_{sw} , longwave radiative flux Q'_{lw} , sensible heat flux Q'_{sh} , and latent heat flux Q'_{lh} . On the interdecadal time scale, the MLT anomalies tendency term on the left-hand side of Eq. (1) is smaller than the two right-hand-side terms by at least one order of magnitude and can thus be neglected (Dong et al. 2014; Xie et al. 2010; Zhang and Li 2014). We decompose the anomalous latent heat flux term into anomalous latent heat flux due to the atmospheric forcing effect Q'^a_{lh} and Newtonian damping effect Q'^o_{lh} . Then, Eq. (1) is changed to

$$T' = \frac{D'_o + Q'_a}{\alpha Q'_{\text{lh}}}, \quad (2)$$

where $Q'_a = Q'_{\text{sw}} + Q'_{\text{lw}} + Q'_{\text{sh}} + Q'^a_{\text{lh}}$, which is dominant by the Q'_{sw} and Q'^a_{lh} , and α denotes the Newtonian damping coefficient (approximately 0.06 K^{-1} ; Xie et al. 2010; Zhang and Li 2014). More detailed derivation can be found in Xie et al. (2010).

The ID-IOBM pattern calculated by using Eq. (2) is similar to that which was directly output from the model

(Figs. 2a and 1d), suggesting that the formula can be used to estimate the relative contributions of anomalous ocean dynamic effects D'_o , shortwave radiative flux Q'_{sw} , and latent heat flux due to atmospheric forcing Q'^a_{lh} to the formation of ID-IOBM. It is seen that the basinwide warming results from the combined effects of three terms (Figs. 2b–d). Meanwhile, the SSTAs warming center is located in the southwestern TIO, asymmetric about the equator, which is mainly due to the effect of ocean dynamics (Fig. 2b). However, in this approach, the ocean dynamic effect is estimated based on its balance with the anomalous net surface heat flux Q'_{net} , rather than diagnosed explicitly, which limits our further understanding of detailed ocean dynamic processes.

4. Season-dependent growth mechanisms of the ID-IOBM

In this section, we focus on the seasonal evolutions of ID-IOBM, which helps us to further understand the mechanisms responsible for the formation of ID-IOBM from a perspective that is different from the balance between atmospheric forcing and Newtonian damping effects used in the previous studies (Dong et al. 2014; Xie et al. 2010; Zhang and Li 2014) and section 3.

a. Seasonal evolution of the ID-IOBM

It is found that the intensity of ID-IOBM shows a seasonally evolving feature in all four observational SST datasets (Fig. 3). For each dataset, the seasonal ID-IOBM intensity is

FIG. 6. Spatial distribution of the ID-IOBM-related (a) SSTAs in NDJ. (b) As in (a), but for the anomalous precipitation (shading; mm day⁻¹), 850-hPa wind anomalies (vectors; m s⁻¹), and 200-hPa velocity potential anomalies (contours; 10⁶ m s⁻¹). The interval of the contours is 1.0. White dots represent values of the shading reaching the 5% significance level. The vectors are shown when either the zonal or the meridional component reaches the 10% significance level. In (a), the black dashed box represents the restoring region for HIST-IPO experiments.

calculated by regressing seasonal-mean area-averaged TIO SSTAs onto the corresponding normalized annual-mean ID-IOBM index shown in Fig. 1e. Although there are uncertainties in the phases of the seasonal evolutions among the four datasets, the ID-IOBM tends to be stronger in boreal winter and spring than in summer and fall (Fig. 3). For the ensemble mean of the four datasets, the magnitude of seasonal change of ID-IOBM reaches 0.032 K (maximum minus minimum), which is approximately 43.1% of the annual-mean amplitude of ID-IOBM. The CESM results reproduce the seasonal evolution of ID-IOBM intensity, with its seasonal change magnitude (0.038 K) close to the observations. The ID-IOBM in the CESM results reaches its peak (trough) in early spring (fall), approximately consistent with the observations (Fig. 3). The similarities between the CESM results and observations give us confidence to further investigate the physical processes modulating the

seasonal evolution of ID-IOBM based on the CESM results.

The seasonal evolution of spatial patterns of ID-IOBM in the CESM results is given in Fig. 4. Here, we select the peak [February–April (FMA)] and trough [August–October (ASO)] of ID-IOBM amplitude and two seasons with maximum and minimum time tendencies (NDJ and JJA). The TIO is dominated by the basinwide warming in all four seasons. The ID-IOBM intensity in FMA is much stronger than in the other three seasons, with warm SSTA centers located approximately 10°S. The ID-IOBM-related SSTAs in FMA and June–August (JJA) are generally zonally uniform, while those in ASO and November–January (NDJ) exhibit a significant zonal gradient, with strong warm SSTAs in the western TIO and weak warm SSTAs in the east.

To investigate the growth and decay of ID-IOBM in the seasonal cycle, the spatial distributions of SSTA tendencies in NDJ and JJA are shown in Fig. 5. For

FIG. 7. As in Fig. 6, but for the (a) net surface shortwave radiative flux anomalies (W m^{-2}) and (b) latent heat flux (shading; W m^{-2}) and 10-m wind speed anomalies (contours with interval of 0.03 m s^{-1} ; negative contours are dashed). White dots represent values of the shading reaching the 5% significance level. The purple boxes are as in Fig. 5, but are divided into the western TIO (3°N – 15°S , 40° – 70°E) and eastern TIO (3°N – 15°S , 70° – 100°E).

NDJ, the TIO is dominated by a band of warming tendencies from 3°N to 15°S , with the maximum located on the western coast of Sumatra (Fig. 5a). The SSTA tendency in JJA is generally opposite to that in NDJ, but the magnitude is smaller (Fig. 5b).

b. Growth mechanisms of the ID-IOBM in early winter

In this subsection, we investigate what physical processes cause the band of warming tendencies in NDJ (purple box in Fig. 5a) through examining the ID-IOBM-related SSTAs and large-scale circulations in the Indo-Pacific region. The IPO-like warm SSTAs in the equatorial central–eastern Pacific stimulate an anomalous Walker circulation. Its downwelling branch suppresses convection over the tropical Indo-Pacific warm pool region. This suppressed convection over the eastern TIO drives easterly anomalies and twin anticyclone anomalies residing on both sides of the equator in the TIO (Fig. 6), according to the Gill model (Gill 1980). The negative precipitation anomalies and associated anomalous circulations play dominant roles in warming the TIO SSTAs.

In section 3, the warm SSTAs in the TIO were understood via the analysis model under the assumption that the MLT anomalies tendency term is one order of magnitude smaller than the dynamic and thermodynamic forcing terms on the interdecadal time scale. The assumption is invalid when the seasonal evolution is involved, because this time scale is far smaller than the interdecadal time scale. The mixed layer heat budget for the ID-IOBM-related variability in NDJ is explicitly diagnosed. The linearized MLT tendency equation is written as

$$\left(\frac{\partial T'}{\partial t}\right) = -(\bar{\mathbf{V}} \cdot \nabla_h T' + \mathbf{V}' \cdot \nabla_h \bar{T}) - \left(\bar{w} \frac{\partial T'}{\partial z} + w' \frac{\partial \bar{T}}{\partial z}\right) + \left(\frac{Q'_{\text{net}}}{\rho C_p H}\right) + R, \quad (3)$$

where T denotes the MLT; $\mathbf{V}$ and w denote horizontal and vertical velocity of oceanic currents; ∇_h and $\partial/\partial z$ denote the horizontal and vertical gradient operators, respectively; and the bar and prime in the equation denote climatological-mean and ID-IOBM-related anomalous variables, respectively. Also, Q_{net} denotes the net surface heat flux; ρ ($= 1000 \text{ kg m}^{-3}$), C_p ($= 4000 \text{ J kg K}^{-1}$), and H denote density of seawater, specific heat of water, and mixed layer depth, respectively. For the CESM data, H is defined as the shallowest depth where the local interpolated buoyancy gradient matches the maximum buoyancy gradient between the surface and any discrete depth within that water column (Large et al. 1997). The nonlinear terms and other subgrid processes that cannot be resolved by the model are denoted by R .

The diagnosis results indicate that the warming tendencies in the eastern and western TIO are caused by different processes. For the eastern TIO, the positive net surface shortwave radiative flux anomalies have a dominant contribution to this warming tendency (relative contribution is 83.4%; Fig. 7a). Spatial distribution of the net surface shortwave radiative flux anomalies is similar to that of the precipitation anomalies (Fig. 6b), indicating that suppressed convection reduces local cloud cover and thus enhances the absorbed shortwave radiation in the eastern TIO. The mechanism is also seen

FIG. 8. (a) The ID-IOBM-related mixed layer heat budget terms (K month^{-1}) area-averaged over the western TIO in NDJ. From left to right, the bars represent the time tendency of mixed layer temperature (MLT) anomalies derived from the model outputs, the time tendency of MLT anomalies calculated from the right-hand side of Eq. (3), contributions from ocean dynamic effects and net surface heat flux anomalies. (b) As in (a), but for individual linearized horizontal and vertical oceanic advection anomalies.

in the formation and maintenance of the conventional IA-IOBM (Klein et al. 1999; Du et al. 2009; Wu and Kinter 2010). In early winter, the mean westerly (easterly) prevails in the equatorial (off equatorial) Indian Ocean in both the observation and CESM results (figure not shown). The easterly anomalies over the TIO lead to the decrease (increase) of the wind speed and thus the positive (negative) latent heat flux anomalies in the equatorial (off equatorial) Indian Ocean (Fig. 7b). Hence, the latent heat flux term has fewer net contributions to the warming tendency in the eastern TIO.

The mixed layer heat budget analysis indicates that the warming in the western TIO is significantly attributed to the ocean dynamic effects (relative contribution is 80.8%), while the net surface heat fluxes term has only a small contribution (Fig. 8a). The positive latent heat fluxes due to the easterly wind anomalies are mostly offset by the decreased absorbed surface short-wave radiative fluxes in the western TIO (Fig. 7). Given the importance of ocean dynamic effects, a further examination of the individual oceanic advection terms is conducted in Eq. (3).

The $-\mathbf{V}' \cdot \nabla_h \bar{T}$ term is dominant in the warming tendency in the western TIO (Figs. 8b and 9a), with a contribution larger than 55%. Climatological NDJ-mean MLT has a warm center located in the equatorial central Indian Ocean, with a positive zonal gradient of MLT in the western TIO (Fig. 9b). As a result, anomalous westward ocean currents driven by the equatorial

FIG. 9. (a) As in Figs. 6 and 7, but for horizontal advection of mean temperatures by anomalous currents (K month^{-1}). (b) Corresponding NDJ-mean mixed layer temperatures (shading; K) and anomalous currents (vectors; cm s^{-1}). The purple box is as in Fig. 7 for the western TIO.

easterly wind stress anomalies advect the warm water to the equatorial western TIO and thus warm this region (Fig. 9). Meanwhile, anomalous southward Ekman currents cause warm advection anomalies in the off-equatorial southwestern TIO.

Compared with the $-\mathbf{V}' \cdot \nabla_h \bar{T}$ term, the $-w' \partial \bar{T} / \partial z$ term plays a secondary role in the warming tendency of MLT in the western TIO (relative contribution is 18.6%; Figs. 8b and 10a). The equatorial easterly anomalies and anticyclonic anomalies over the southern TIO drive an anomalous meridional overturning circulation through Ekman pumping, with anomalous upwelling along the

FIG. 10. (a) As in Fig. 9, but for vertical advection of mean temperatures by anomalous currents (K month^{-1}). (b) Depth-latitude section of NDJ-mean ocean temperatures (shading; K) and anomalous vertical currents (vectors; cm s^{-1} and $10^{-4} \text{ cm s}^{-1}$), which is zonally averaged over 40°–80°E. The purple box is as in Fig. 9.

equator and downwelling along approximately 10°S (Fig. 10b). The anomalous vertical motions cause negative (positive) vertical advection anomalies in the equator (the off-equatorial southwestern TIO) (Fig. 10a). The southwestern TIO has a thermocline dome associated with climatological upwelling driven by local negative wind stress curl (Xie et al. 2002; Du et al. 2009), which is reproduced by the CESM (figure not shown). The presence of the mean upwelling is the necessary condition under which the positive vertical advection anomalies can affect the southwestern TIO SST. Though the $-w'\partial\bar{T}/\partial z$ term has a smaller net contribution to the warming tendency in the western TIO, it leads to the maximum of the warming tendency along approximately 10°S (Fig. 5a).

FIG. 11. The ID-IOBM-related mixed layer heat budget terms (K month^{-1}) area-averaged over the TIO (15°N–15°S, 40°–100°E) in JJA. From left to right, the bars represent the time tendency of mixed layer temperature anomalies derived from the model outputs, the time tendency of MLT anomalies calculated from the right-hand side of Eq. (3), contributions from ocean dynamic effects, surface shortwave radiative flux, and latent heat flux anomalies.

c. Decay mechanisms of the ID-IOBM in summer

The SSTA tendency associated with the ID-IOBM in JJA is negative, opposite to that in NDJ, though the magnitude is smaller (Fig. 5b). To understand why the ID-IOBM decays in JJA, a parallel mixed layer heat budget analysis is conducted. It demonstrates that the increased upward latent heat flux anomalies play a dominant role in causing the negative MLT tendency in the TIO (Fig. 11). The increased upward latent heat flux is dominated by the Newtonian damping effect instead of atmospheric forcing (figure not shown).

The intensities of precipitation and wind anomalies in JJA over the TIO are far weaker than those in NDJ (Fig. 12). The difference is associated with two factors. First, the SSTAs intensity in the equatorial central-eastern Pacific (black box in Fig. 12a) decreases by approximately 22.1% relative to that in NDJ (Figs. 6a and 12a), corresponding to the weaker IPO remote forcing. Second, and more importantly, the enhanced convective heating over the tropical central-eastern Pacific is shifted eastward by approximately 50° compared with that in NDJ, which corresponds to the eastward shift of the anomalous Walker circulation (Fig. 12b). As a result, the TIO is out of the control of the descending branch of the anomalous Walker circulation. The ID-IOBM-related SSTAs are gradually damped through the Newtonian damping effect.

5. Conclusions and discussion

a. Conclusions

In this study, we explore how the IPO drives ID-IOBM based on the observational data and pacemaker experiments of CESM, in which the SSTAs in the

FIG. 12. As in Fig. 6, but for JJA. The position of the black box denotes the equatorial central–eastern Pacific region (5°N – 5°S , 170° – 120°W).

equatorial central–eastern Pacific are restored to the observations. Our major findings are summarized as follows:

- 1) After the external forcing signals are removed by subtracting the HIST_ENS, the HIST-IPO_ENS can simulate the spatial pattern of ID-IOBM to a large extent, suggesting that the ID-IOBM can be generated by the air–sea integrations in the Indo-Pacific region. The temporal evolution of ID-IOBM in the CESM results is highly synchronous with that of the equatorial central–eastern Pacific SSTAs, suggesting that it is a passive response to the IPO remote forcing.
- 2) It is found that the intensity of ID-IOBM shows a seasonal evolution characteristic in all four SST observational datasets; that is, the ID-IOBM tends to be stronger in winter and spring than in summer and fall. The CESM results largely reproduce this feature. In the CESM results, the peak (trough) of ID-IOBM intensity is reached in the early spring

(fall), indicating that the fastest growth (decay) rate of ID-IOBM occurs in early winter (summer).

- 3) The mechanisms responsible for the growth and decay of ID-IOBM are summarized in a schematic diagram (Fig. 13). In early winter (NDJ), the ID-IOBM is greatly intensified by the distinct processes in the eastern and western TIO. The eastern TIO is dominated by the descending branch of the anomalous Walker circulation associated with the IPO. The anomalous downward motion suppresses local convection, which increases incoming solar radiation and leads to SST warming. Other physical processes have smaller contributions to the warming tendency in the eastern TIO. The warming tendency in the western TIO is dominated by ocean dynamic effects. The equatorial easterly anomalies drive westward anomalous equatorial ocean currents and southward off-equatorial Ekman currents as a response to the suppressed convection over the southeastern TIO. In NDJ, the warmest climatological TIO SST is located in the equatorial

FIG. 13. Schematic diagram of the mechanisms responsible for the (a) growth and (b) decay of ID-IOBM. The closed black dashed lines denote the ID-IOBM-related anomalous Walker circulation. In (a), the brown dashed vector represents the ID-IOBM-related low-level wind anomalies. The gray solid vector represents the NDJ-mean climatological westerly in the equatorial Indian Ocean. The red dashed and solid contours are the 28° and 29°C isotherms of the climatological SST in NDJ, respectively.

central Indian Ocean. As a result, the anomalous ocean currents advect warm water to the equatorial western Indian Ocean and off-equatorial southwestern Indian Ocean. Meanwhile, the equatorial easterly anomalies and anticyclone anomalies over the southern TIO drive an anomalous meridional overturning circulation, with anomalous upwelling and downwelling along the equator and approximately 10°S, respectively. The anomalous vertical motions lead to cold (warm) vertical advection along the equator (approximately 10°S), which shifts the center of the warming tendency in the western TIO to approximately 10°S. Except for the ocean dynamic processes, the decreased upward latent heat flux anomalies due to the weakened climatological westerlies also contribute to this warming tendency.

- 4) In summer (JJA), the TIO is out of the control of the descending branch of the anomalous Walker circulation. This is because the IPO-related enhanced convective heating over the tropical central–eastern Pacific shifts eastward by approximately 50° relative to that in NDJ. As a result, the ID-IOBM is

weakened through the Newtonian damping effect and reaches its trough in early fall.

b. Discussion

1) DISCREPANCIES IN THE CESM SIMULATIONS

The temporal evolution of ID-IOBM derived from the CESM results is not exactly consistent with that in the observation (Fig. 1e). Except for the uncertainties in the observational data (Han et al. 2014b; Zhang 2016) and the impact of the external forcing that cannot be cleanly removed in the observational analysis (Dong and McPhaden 2017b; Zhang et al. 2018), inherent issues in the artificially designed pacemaker experiments may contribute to the difference in the temporal evolution between the simulated and observed ID-IOBM. First, in the HIST-IPO experiments, the IPO-related SSTAs are specified, and the ID-IOBM is a passive response to the IPO forcing. In the observations, the ID-IOBM may have active forcing on the SSTAs in the equatorial central–eastern Pacific, especially after the 1950s when

FIG. 14. Spatial distribution of the IPO-related (a) SSTAs (shading; $^{\circ}\text{C}$) in NDJ. (b) As in (a), but for the anomalous precipitation (shading; mm day^{-1}) and 200-hPa velocity potential anomalies (contours with an interval of $1.0 \times 10^6 \text{ m}^2 \text{ s}^{-1}$; negative contours are dashed). (c),(d) As in (a) and (b), respectively, but for JJA. White dots represent values of the shading reaching the 5% significance level. The SST is from the HadISST1.1 dataset, and the precipitation and velocity potential are from the 20CR data.

the TIO SST increasingly warmed (Arora et al. 2016; Luo et al. 2012). Second, the Atlantic multidecadal oscillation, another dominant interdecadal variability mode, may modulate Indo-Pacific SST variability in terms of previous studies. It was suggested that the positive AMO favors the basinwide warm SSTAs in the TIO (Li et al. 2016; Kucharski et al. 2016). The contribution of the AMO to the ID-IOBM may be studied through North Atlantic pacemaker experiments in the future.

The ID-IOBM is forced by the IPO-driven convective heating anomalous over the equatorial central-eastern Pacific and associated anomalous Walker circulation. The discrepancies in simulating the heating anomalies may influence the simulation skill for the ID-IOBM. The ID-IOBM-related SSTAs in the equatorial Pacific in the CESM results are stronger than those in the observations. Meanwhile, the overlying positive precipitation anomalies in the CESM results are shifted westward relative to those in the reanalysis data (figure not shown). These biases are associated with the excessive westward extension of the cold tongue in the climatology, a common discrepancy of coupled models (Wang et al. 2017). However, these biases are not very severe in the CESM.

2) COMPARISONS BETWEEN ID-IOBM AND IA-IOBM

Both the IA-IOBM and ID-IOBM are driven by remote forcing from the equatorial central-eastern Pacific, and have season-dependent features. However, the seasonal dependences in the two modes have different

specific meaning. First, the IA-IOBM has a season-dependent life cycle, which generally forms in winter, reaches the peak in spring, and persists to following summer [Klein et al. 1999; Alexander et al. 2002; Du et al. 2009; also see Xie et al. (2016) for a review]. However, the ID-IOBM exists in all year-round. Its intensity shows the seasonal dependence, with the peak in early spring and the trough in early fall (Fig. 3). Second, the season-dependent life cycle of the IA-IOBM is closely associated with the seasonal phase locking of both ENSO forcing [Klein et al. 1999; Alexander et al. 2002; also see Xie et al. (2016) for a review] and preceding Indian Ocean dipole (IOD) (Li et al. 2003; Hong et al. 2010). The IOD tends to turn into IA-IOBM in winter due to the changes in direction of the mean wind over the coast of Sumatra (Li et al. 2003; Tokinaga and Tanimoto 2004) and the eastward propagation of the equatorial downwelling Kelvin waves over the TIO (Yu et al. 2005; Yuan and Liu 2009). The development of the IA-IOBM during a mature El Niño winter is driven by the descending branch of El Niño-driven anomalous Walker circulation (Klein et al. 1999; Alexander et al. 2002). Its maintenance in following spring primarily relies on local wind–evaporation–SST feedback (Du et al. 2009). To El Niño decaying summer, the remote forcing from the central-eastern Pacific has disappeared, the IA-IOBM gradually decays. This study indicated that the growth (decay) of the ID-IOBM occurs in early winter (summer). The season-dependent growth–decay feature is associated with the seasonal change in the zonal position of the IPO-driven anomalous Walker circulation.

3) THE SEASONALLY ZONAL SHIFT OF ANOMALOUS WALKER CIRCULATION FORCED BY THE IPO

The seasonally zonal shift of the anomalous Walker circulation forced by the IPO leads to the season-dependent growth of ID-IOBM in the CESM. This is a robust feature that is also seen in the 20CR data (Compo et al. 2011) (Fig. 14). The zonal position of the maximum center of the warm SSTAs does not show evident changes between summer and early winter (Figs. 14a,c). However, the responses of the overlying precipitation anomalies are completely different (Figs. 14b,d). In early winter, the center of the positive precipitation anomalies is located on the equator and to west of the warm SSTA center (Figs. 14a,b). In summer, the center of the precipitation anomalies is off the equator and over the eastern Pacific, generally in phase with the center of the warm SSTAs (Figs. 14c,d). Corresponding to the zonal shift of the enhanced convective heating, the anomalous Walker circulation is shifted zonally. The seasonally zonal shift of anomalous Walker circulation is relevant to the study by Han et al. (2017), who revealed that on the interdecadal time scale, the Pacific and Indian Ocean Walker cells covary during winter but are uncorrelated during summer.

To our knowledge, there is no explanation for the seasonally zonal shift of the anomalous convective heating over the equatorial central–eastern Pacific forced by the IPO so far. It is interesting to note that the positive precipitation anomalies during a developing El Niño summer is located to the west of that during mature El Niño winter, instead of to the east (Wu et al. 2017). The mechanism causing the differences in the responses of precipitation to SST deserves further study.

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REFERENCES

- Alexander, M. A., I. Bladé, M. Newman, J. R. Lanzante, N.-C. Lau, and J. D. Scott, 2002: The atmospheric bridge: The influence of ENSO teleconnections on air–sea interaction over the global oceans. *J. Climate*, **15**, 2205–2231, [https://doi.org/10.1175/1520-0442\(2002\)015<2205:TABTIO>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<2205:TABTIO>2.0.CO;2).
- Arora, A., S. A. Rao, R. Chattopadhyay, T. Goswami, G. George, and C. T. Sabeerali, 2016: Role of Indian Ocean SST variability on the recent global warming hiatus. *Global Planet. Change*, **143**, 21–30, <https://doi.org/10.1016/j.gloplacha.2016.05.009>.
- Boer, G. J., and Coauthors, 2016: The Decadal Climate Prediction Project (DCPP) contribution to CMIP6. *Geosci. Model Dev.*, **9**, 3751–3777, <https://doi.org/10.5194/gmd-9-3751-2016>.
- Chiang, J. C. H., and A. H. Sobel, 2002: Tropical tropospheric temperature variations caused by ENSO and their influence on the remote tropical climate. *J. Climate*, **15**, 2616–2631, [https://doi.org/10.1175/1520-0442\(2002\)015<2616:TTVCB>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<2616:TTVCB>2.0.CO;2).
- Compo, G. P., and Coauthors, 2011: The Twentieth Century Reanalysis Project. *Quart. J. Roy. Meteor. Soc.*, **137**, 1–28, <https://doi.org/10.1002/qj.776>.
- Dong, L., and T. Zhou, 2014: The Indian Ocean sea surface temperature warming simulated by CMIP5 models during the 20th century: Competing forcing roles of GHGs and anthropogenic aerosols. *J. Climate*, **27**, 3348–3362, <https://doi.org/10.1175/JCLI-D-13-00396.1>.
- , and M. J. McPhaden, 2016: Interhemispheric SST gradient trends in the Indian Ocean prior to and during the recent global warming hiatus. *J. Climate*, **29**, 9077–9095, <https://doi.org/10.1175/JCLI-D-16-0130.1>.
- , and —, 2017a: The effects of external forcing and internal variability on the formation of interhemispheric sea surface temperature gradient trends in the Indian Ocean. *J. Climate*, **30**, 9077–9095, <https://doi.org/10.1175/JCLI-D-17-0138.1>.
- , and —, 2017b: Why has the relationship between Indian and Pacific Ocean decadal variability changed in recent decades? *J. Climate*, **30**, 1971–1983, <https://doi.org/10.1175/JCLI-D-16-0313.1>.
- , T. Zhou, and B. Wu, 2014: Indian Ocean warming during 1958–2004 simulated by a climate system model and its mechanism. *Climate Dyn.*, **42**, 203–217, <https://doi.org/10.1007/s00382-013-1722-z>.
- , T. J. Zhou, A. Dai, F. Song, B. Wu, and X. Chen, 2016: The footprint of the inter-decadal Pacific oscillation in Indian Ocean sea surface temperatures. *Sci. Rep.*, **6**, 21251, <https://doi.org/10.1038/srep21251>.
- Du, Y., and S.-P. Xie, 2008: Role of atmospheric adjustments in the tropical Indian Ocean warming during the 20th century in climate models. *Geophys. Res. Lett.*, **35**, L08712, <https://doi.org/10.1029/2008GL033631>.
- , —, G. Huang, and K. Hu, 2009: Role of air–sea interaction in the long persistence of El Niño–induced north Indian Ocean warming. *J. Climate*, **22**, 2023–2038, <https://doi.org/10.1175/2008JCLI2590.1>.
- Duchon, C., 1979: Lanczos filtering in one and two dimensions. *J. Appl. Meteor.*, **18**, 1016–1022, [https://doi.org/10.1175/1520-0450\(1979\)018<1016:LFIOAT>2.0.CO;2](https://doi.org/10.1175/1520-0450(1979)018<1016:LFIOAT>2.0.CO;2).
- Ebisuzaki, W., 1997: A method to estimate the statistical significance of a correlation when the data are serially correlated. *J. Climate*, **10**, 2147–2153, [https://doi.org/10.1175/1520-0442\(1997\)010<2147:AMTETS>2.0.CO;2](https://doi.org/10.1175/1520-0442(1997)010<2147:AMTETS>2.0.CO;2).
- Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.*, **106**, 447–462, <https://doi.org/10.1002/qj.49710644905>.
- Han, W., and Coauthors, 2014a: Intensification of decadal and multi-decadal sea level variability in the western tropical Pacific during recent decades. *Climate Dyn.*, **43**, 1357–1379, <https://doi.org/10.1007/s00382-013-1951-1>.

- , J. Vialard, M. J. McPhaden, T. Lee, Y. Masumoto, M. Feng, and W. P. M. de Ruijter, 2014b: Indian Ocean decadal variability: A review. *Bull. Amer. Meteor. Soc.*, **95**, 1679–1703, <https://doi.org/10.1175/BAMS-D-13-00028.1>.
- , G. A. Meehl, A. Hu, J. Zheng, J. Kenigson, J. Vialard, B. Rajagopalan, and Yanto, 2017: Decadal variability of the Indian and Pacific Walker cells since the 1960s: Do they covary on decadal time scales? *J. Climate*, **30**, 8447–8468, <https://doi.org/10.1175/JCLI-D-16-0783.1>.
- Henley, B. J., and Coauthors, 2017: Spatial and temporal agreement in climate model simulations of the Interdecadal Pacific Oscillation. *Environ. Res. Lett.*, **12**, 044011, <https://doi.org/10.1088/1748-9326/aa5cc8>.
- Hong, C.-C., T. Li, LinHo, and Y.-C. Chen, 2010: Asymmetry of the Indian Ocean basinwide SST anomalies: Roles of ENSO and IOD. *J. Climate*, **23**, 3563–3576, <https://doi.org/10.1175/2010JCLI3320.1>.
- Huang, B. H., and J. L. Kinter III, 2002: Interannual variability in the tropical Indian Ocean. *J. Geophys. Res.*, **107**, 3199, <https://doi.org/10.1029/2001JC001278>.
- Huang, B. Y., and Coauthors, 2017: Extended Reconstructed Sea Surface Temperature, version 5 (ERSSTv5): Upgrades, validations, and intercomparisons. *J. Climate*, **30**, 8179–8205, <https://doi.org/10.1175/JCLI-D-16-0836.1>.
- Hurrell, J. W., and Coauthors, 2013: The Community Earth System Model: A framework for collaborative research. *Bull. Amer. Meteor. Soc.*, **94**, 1339–1360, <https://doi.org/10.1175/BAMS-D-12-00121.1>.
- Kaplan, A., M. A. Cane, Y. Kushnir, A. C. Clement, M. B. Blumenthal, and B. Rajagopalan, 1998: Analyses of global sea surface temperature 1856–1991. *J. Geophys. Res.*, **103**, 18 567–18 589, <https://doi.org/10.1029/97JC01736>.
- Kennedy, J. J., N. A. Rayner, R. O. Smith, D. E. Parker, and M. Saunby, 2011a: Reassessing biases and other uncertainties in sea surface temperature observations measured in situ since 1850: 1. Measurement and sampling uncertainties. *J. Geophys. Res.*, **116**, D14103, <https://doi.org/10.1029/2010JD015218>.
- , —, —, —, and —, 2011b: Reassessing biases and other uncertainties in sea surface temperature observations measured in situ since 1850: 2. Biases and homogenization. *J. Geophys. Res.*, **116**, D14104, <https://doi.org/10.1029/2010JD015220>.
- Klein, S. A., B. J. Soden, and N.-C. Lau, 1999: Remote sea surface temperature variations during ENSO: Evidence for a tropical atmospheric bridge. *J. Climate*, **12**, 917–932, [https://doi.org/10.1175/1520-0442\(1999\)012<0917:RSSTVD>2.0.CO;2](https://doi.org/10.1175/1520-0442(1999)012<0917:RSSTVD>2.0.CO;2).
- Krishnamurthy, V., and B. P. Kirtman, 2003: Variability of the Indian Ocean: Relation to monsoon and ENSO. *Quart. J. Roy. Meteor. Soc.*, **129**, 1623–1646, <https://doi.org/10.1256/qj.01.166>.
- Kucharski, F., and Coauthors, 2016: The teleconnection of the tropical Atlantic to Indo-Pacific sea surface temperatures on inter-annual to centennial time scales: A review of recent findings. *Atmosphere*, **7**, 29, <https://doi.org/10.3390/atmos7020029>.
- Lamarque, J. F., and Coauthors, 2012: CAM-chem: Description and evaluation of interactive atmospheric chemistry in the Community Earth System Model. *Geosci. Model Dev.*, **5**, 369–411, <https://doi.org/10.5194/gmd-5-369-2012>.
- Large, W., G. Danabasoglu, S. Doney, and J. McWilliams, 1997: Sensitivity to surface forcing and boundary layer mixing in the NCAR CSM ocean model: Annual-mean climatology. *J. Phys. Oceanogr.*, **27**, 2418–2447, [https://doi.org/10.1175/1520-0485\(1997\)027<2418:STSFA>2.0.CO;2](https://doi.org/10.1175/1520-0485(1997)027<2418:STSFA>2.0.CO;2).
- Latif, M., and T. P. Barnett, 1995: Interactions of the tropical oceans. *J. Climate*, **8**, 952–964, [https://doi.org/10.1175/1520-0442\(1995\)008<0952:IOTTO>2.0.CO;2](https://doi.org/10.1175/1520-0442(1995)008<0952:IOTTO>2.0.CO;2).
- Lau, N.-C., and M. J. Nath, 2004: Coupled GCM simulation of atmosphere–ocean variability associated with zonally asymmetric SST changes in the tropical Indian Ocean. *J. Climate*, **17**, 245–265, [https://doi.org/10.1175/1520-0442\(2004\)017<0245:CGSOAV>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<0245:CGSOAV>2.0.CO;2).
- Lee, S.-K., W. Park, M. O. Baringer, A. L. Gordon, B. Huber, and Y. Liu, 2015: Pacific origin of the abrupt increase in Indian Ocean heat content during the warming hiatus. *Nat. Geosci.*, **8**, 445–449, <https://doi.org/10.1038/ngeo2438>.
- Li, T., B. Wang, C.-P. Chang, and Y. S. Zhang, 2003: A theory for the Indian Ocean dipole–zonal mode. *J. Atmos. Sci.*, **60**, 2119–2135, [https://doi.org/10.1175/1520-0469\(2003\)060<2119:ATFTIO>2.0.CO;2](https://doi.org/10.1175/1520-0469(2003)060<2119:ATFTIO>2.0.CO;2).
- , —, B. Wu, T. Zhou, C.-P. Chang, and R. Zhang, 2017: Theories on formation of an anomalous anticyclone in western North Pacific during El Niño: A review. *J. Meteor. Res.*, **31**, 987–1006, <https://doi.org/10.1007/s13351-017-7147-6>.
- Li, X., S.-P. Xie, S. T. Gille, and C. Yoo, 2016: Atlantic-induced pan-tropical climate change over the past three decades. *Nat. Climate Change*, **6**, 275–279, <https://doi.org/10.1038/nclimate2840>.
- Liu, Z., and M. Alexander, 2007: Atmospheric bridge, oceanic tunnel, and global climatic teleconnections. *Rev. Geophys.*, **45**, RG2005, <https://doi.org/10.1029/2005RG000172>.
- Luo, J. J., W. Sasaki, and Y. Masumoto, 2012: Indian Ocean warming modulates Pacific climate change. *Proc. Natl. Acad. Sci. USA*, **109**, 18 701–18 706, <https://doi.org/10.1073/pnas.1210239109>.
- Mantua, N., S. Hare, Y. Zhang, J. Wallace, and R. Francis, 1997: A Pacific interdecadal oscillation with impacts on salmon production. *Bull. Amer. Meteor. Soc.*, **78**, 1069–1079, [https://doi.org/10.1175/1520-0477\(1997\)078<1069:APICOW>2.0.CO;2](https://doi.org/10.1175/1520-0477(1997)078<1069:APICOW>2.0.CO;2).
- Neale, R. B., J. Richter, S. Park, P. H. Lauritzen, S. J. Vavrus, P. J. Rasch, and M. Zhang, 2013: The mean climate of the Community Atmosphere Model (CAM4) in forced SST and fully coupled experiments. *J. Climate*, **26**, 5150–5168, <https://doi.org/10.1175/JCLI-D-12-00236.1>.
- Newman, M., and Coauthors, 2016: The Pacific decadal oscillation, revisited. *J. Climate*, **29**, 4399–4427, <https://doi.org/10.1175/JCLI-D-15-0508.1>.
- Power, S., T. Casey, C. Folland, A. Colman, and V. Mehta, 1999: Inter-decadal modulation of the impact of ENSO on Australia. *Climate Dyn.*, **15**, 319–324, <https://doi.org/10.1007/s003820050284>.
- Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan, 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.*, **108**, 4407, <https://doi.org/10.1029/2002JD002670>.
- Roxy, M., S. Gualdi, H. K. L. Drbohlav, and A. Navarra, 2011: Seasonality in the relationship between El Niño and Indian Ocean dipole. *Climate Dyn.*, **37**, 221–236, <https://doi.org/10.1007/s00382-010-0876-1>.
- Saji, N. H., and T. Yamagata, 2003: Structure of SST and surface wind variability during Indian Ocean dipole mode events: COADS observations. *J. Climate*, **16**, 2735–2751, [https://doi.org/10.1175/1520-0442\(2003\)016<2735:SOSASW>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<2735:SOSASW>2.0.CO;2).
- , B. N. Goswami, P. N. Vinayachandran, and T. Yamagata, 1999: A dipole mode in the tropical Indian Ocean. *Nature*, **401**, 360–363.

- Schott, F. A., S.-P. Xie, and J. P. McCreary, 2009: Indian Ocean circulation and climate variability. *Rev. Geophys.*, **47**, RG1002, <https://doi.org/10.1029/2007RG000245>.
- Smith, R. D., and Coauthors, 2010: The Parallel Ocean Program (POP) reference manual. Los Alamos Unclassified Rep. LAUR-10-01853, 140 pp.
- Taylor, K. E., R. J. Stouffer, and G. A. Meehl, 2012: An overview of CMIP5 and the experiment design. *Bull. Amer. Meteor. Soc.*, **93**, 485–498, <https://doi.org/10.1175/BAMS-D-11-00094.1>.
- Ting, M., Y. Kushnir, R. Seager, and C. Li, 2009: Forced and internal twentieth-century SST trends in the North Atlantic. *J. Climate*, **22**, 1469–1481, <https://doi.org/10.1175/2008JCLI2561.1>.
- Tokinaga, H., and Y. Tanimoto, 2004: Seasonal transition of SST anomalies in the tropical Indian Ocean during El Niño and Indian Ocean dipole years. *J. Meteor. Soc. Japan*, **82**, 1007–1018, <https://doi.org/10.2151/jmsj.2004.1007>.
- Tozuka, T., J. Luo, S. Masson, and T. Yamagata, 2007: Decadal modulations of the Indian Ocean dipole in the SINTEX-F1 coupled GCM. *J. Climate*, **20**, 2881–2894, <https://doi.org/10.1175/JCLI4168.1>.
- Wang, C.-Y., S.-P. Xie, Y. Kosaka, Q. Liu, and X.-T. Zheng, 2017: Global influence of tropical Pacific variability with implications for global warming slowdown. *J. Climate*, **30**, 2679–2695, <https://doi.org/10.1175/JCLI-D-15-0496.1>.
- Webster, P. J., A. M. Moore, J. P. Loschnigg, and R. R. Leben, 1999: Coupled ocean–atmosphere dynamics in the Indian Ocean during 1997–98. *Nature*, **401**, 356–360, <https://doi.org/10.1038/43848>.
- Wu, B., and T. Zhou, 2008: Oceanic origin of the interannual and interdecadal variability of the summertime western Pacific subtropical high. *Geophys. Res. Lett.*, **35**, L13701, <https://doi.org/10.1029/2008GL034584>.
- , —, and T. Li, 2009: Seasonally evolving dominant interannual variability modes of East Asian climate. *J. Climate*, **22**, 2992–3005, <https://doi.org/10.1175/2008JCLI2710.1>.
- , —, and —, 2010: Relative contributions of the Indian Ocean and local SST anomalies to the maintenance of the western North Pacific anomalous anticyclone during El Niño decaying summer. *J. Climate*, **23**, 2974–2986, <https://doi.org/10.1175/2010JCLI3300.1>.
- , —, and —, 2012: Two distinct modes of tropical Indian Ocean precipitation in boreal winter and their impacts on equatorial western Pacific. *J. Climate*, **25**, 921–938, <https://doi.org/10.1175/JCLI-D-11-00065.1>.
- , —, and —, 2016: Impacts of the Pacific–Japan and circumglobal teleconnection patterns on the interdecadal variability of the East Asian summer monsoon. *J. Climate*, **29**, 3253–3271, <https://doi.org/10.1175/JCLI-D-15-0105.1>.
- , —, and —, 2017: Atmospheric dynamic and thermodynamic processes driving the western North Pacific anomalous anticyclone during El Niño. Part I: Maintenance mechanisms. *J. Climate*, **30**, 9621–9635, <https://doi.org/10.1175/JCLI-D-16-0489.1>.
- Wu, R., and J. L. Kinter, 2010: Atmosphere–ocean relationship in the midlatitude North Pacific: Seasonal dependence and east–west contrast. *J. Geophys. Res.*, **115**, D06101, <https://doi.org/10.1029/2009JD012579>.
- Xie, S.-P., H. Annamalai, F. A. Schott, and J. P. McCreary Jr., 2002: Structure and mechanisms of South Indian Ocean climate variability. *J. Climate*, **15**, 864–878, [https://doi.org/10.1175/1520-0442\(2002\)015<0864:SAMOSI>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<0864:SAMOSI>2.0.CO;2).
- , K. Hu, J. Hafner, H. Tokinaga, Y. Du, G. Huang, and T. Sampe, 2009: Indian Ocean capacitor effect on Indo-western Pacific climate during the summer following El Niño. *J. Climate*, **22**, 730–747, <https://doi.org/10.1175/2008JCLI2544.1>.
- , C. Deser, G. A. Vecchi, J. Ma, H. Y. Teng, and A. T. Wittenberg, 2010: Global warming pattern formation: Sea surface temperature and rainfall. *J. Climate*, **23**, 966–986, <https://doi.org/10.1175/2009JCLI3329.1>.
- , Y. Kosaka, Y. Du, K. M. Hu, J. S. Chowdary, and G. Huang, 2016: Indo-western Pacific Ocean capacitor and coherent climate anomalies in post-ENSO summer: A review. *Adv. Atmos. Sci.*, **33**, 411–432, <https://doi.org/10.1007/s00376-015-5192-6>.
- Yamagata, T., S. Behera, J.-J. Luo, S. Masson, M. Jury, and S. A. Rao, 2004: Coupled ocean–atmosphere variability in the tropical Indian Ocean. *Earth’s Climate: The Ocean–Atmosphere Interaction, Geophys. Monogr.*, Vol. 147, Amer. Geophys. Union, 189–212.
- Yang, Y., S.-P. Xie, L. Wu, Y. Kosaka, N.-C. Lau, and G. A. Vecchi, 2015: Seasonality and predictability of the Indian Ocean dipole mode: ENSO forcing and internal variability. *J. Climate*, **28**, 8021–8036, <https://doi.org/10.1175/JCLI-D-15-0078.1>.
- Yu, W. D., B. Q. Xiang, L. Liu, and N. Liu, 2005: Understanding the origins of interannual thermocline variations in the tropical Indian Ocean. *Geophys. Res. Lett.*, **32**, L24706, <https://doi.org/10.1029/2005GL024327>.
- Yuan, D. L., and H. L. Liu, 2009: Long-wave dynamics of sea level variations during Indian Ocean dipole events. *J. Phys. Oceanogr.*, **39**, 1115–1132, <https://doi.org/10.1175/2008JPO3900.1>.
- Zhang, L., 2016: The roles of external forcing and natural variability in global warming hiatuses. *Climate Dyn.*, **47**, 3157–3169, <https://doi.org/10.1007/s00382-016-3018-6>.
- , and T. Li, 2014: A simple analytical model for understanding the formation of sea surface temperature patterns under global warming. *J. Climate*, **27**, 8413–8421, <https://doi.org/10.1175/JCLI-D-14-00346.1>.
- , W. Han, and F. Siens, 2018: Unraveling causes for the changing behavior of the tropical Indian Ocean in the past few decades. *J. Climate*, **31**, 2377–2388, <https://doi.org/10.1175/JCLI-D-17-0445.1>.
- Zhou, T. J., and Coauthors, 2016: GMMIP (v1.0) contribution to CMIP6: Global Monsoons Model Inter-Comparison Project. *Geosci. Model Dev.*, **9**, 3589–3604, <https://doi.org/10.5194/gmd-9-3589-2016>.