Indian Ocean Dipole Modes Associated with Different Types of ENSO Development

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(Manuscript received 3 June 2016, in final form 20 November 2016)

ABSTRACT

This study identifies several modes of coevolution of various types of El Niño–Southern Oscillation (ENSO) and Indian Ocean dipole (IOD) by performing rotated season-reliant empirical orthogonal function (S-EOF) analysis with consideration of ENSO asymmetry. The first two modes reveal that early-onset ENSO is associated with subsequent strong IOD development, whereas late-onset ENSO forces an obscure IOD pattern with marginal SST anomalies in the western Indian Ocean. Further studies show that El Niño starting before early summer can more easily force an IOD event than that starting in late summer or fall, even when they are of equivalent magnitudes. This is because the atmospheric responses over the Indian Ocean to the eastern Pacific warming are in sharp contrast between early and late summer. Early-onset (late onset) El Niño can (cannot) cause favorable atmospheric circulation conditions over the Indian Ocean for inducing the western Indian Ocean warming, which facilitates the subsequent IOD development. In addition, the different propagations of ocean dynamic Rossby waves during the early- or late-onset types of ENSO are also accountable for the different IOD development. For the higher-order modes, the rotated S-EOF of “Niño only” cases shows a coevolution between a negative IOD mode and a date line Pacific El Niño, with warm sea surface temperature anomalies originating from the northern Pacific meridional mode.

1. Introduction

El Niño–Southern Oscillation (ENSO) is the largest interannual variation of the coupled atmosphere–ocean system, and it has remote influence on sea surface temperature (SST) anomalies over other ocean basins via

the atmospheric bridge (Lau and Nath 1996; Klein et al. 1999). A typical example is the Indian Ocean basinwide warming mode (IOB) in boreal spring due to an eastward shift of the Walker cell and resultant subsidence with downward surface heat fluxes (Latif and Barnett 1995; Klein et al. 1999). As the dominant mode of the interannual variability of Indian Ocean SST anomalies, the IOB has been studied extensively by the research community because of its significant relationship with ENSO. The second mode of Indian Ocean SST is the Indian Ocean dipole (IOD; Saji et al. 1999; Webster et al. 1999), which has a relatively less robust relationship with ENSO. The IOD usually occurs in boreal summer and fall, and it decays rapidly in winter. A positive IOD is characterized by strong negative SST

Denotes content that is immediately available upon publication as open access.

Supplemental information related to this paper is available at the Journals Online website: <http://dx.doi.org/10.1175/JCLI-D-16-0426.s1>.

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DOI: 10.1175/JCLI-D-16-0426.1

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anomalies off Java and Sumatra and weak positive SST anomalies in the western Indian Ocean. Numerous studies have investigated the influences of the IOD on many regions, including East Africa and Indonesia (Saji et al. 1999), the Asian summer monsoon region (Ashok et al. 2001), and Australia (Cai and Cowan 2008; Ummenhofer et al. 2009).

The consensus is that the IOD, unlike the ENSO-related IOB, has a self-generating mechanism due to the internal atmosphere–ocean coupling of the Indian Ocean (Webster et al. 1999; Allan et al. 2001; Li et al. 2003; Yamagata et al. 2004; Hong et al. 2008a,b). However, more than half the IOD events are induced by external forcings such as ENSO (Ashok et al. 2003; Meyers et al. 2007; Luo et al. 2010) and the southern annular mode (Lau and Nath 2004; Cai et al. 2011). Positive (negative) IOD events usually co-occur with El Niño (La Niña) events in the same year (Saji et al. 1999). ENSO events can influence the entire life span of IOD events (Ueda and Matsumoto 2001; Fischer et al. 2005), and the induced IOD can also provide feedback to the developing phase of ENSO (Behera and Yamagata 2003; Annamalai et al. 2005). This ENSO–IOD connection has strengthened since the 1970s because of the enhancement of the Walker circulation (Yuan and Li 2008).

Recently, increasing numbers of studies have grouped IOD events based on different features. Du et al. (2013) detected a new type of IOD called the “unseasonable” IOD. This type of IOD matures in late summer and is not followed by ENSO, unlike canonical IOD events. Wu et al. (2012) reported contrast rainfall patterns in response to IOD with different onset time. Endo and Tozuka (2015) found a new IOD pattern named “IOD Modoki,” which has negative SST anomalies in the eastern and western tropical Indian Ocean and positive SST anomalies in central parts of the Indian Ocean. The influence on surrounding areas of IOD Modoki differs from canonical IOD events, but its connection with the Pacific Ocean is unclear. Guo et al. (2015) identified a new type of positive (negative) IOD that evolves from nonuniform basinwide warming (cooling) in the tropical Indian Ocean (triggered by the zonal gradient of SST anomaly of the Indian Ocean) during the decaying phase of El Niño (La Niña) events. Both IOD and El Niño can be classified as different types depending either on the zonal location of the SST anomaly between the central and eastern Pacific (Ashok et al. 2007; Kao and Yu 2009; Chen et al. 2015; Capotondi et al. 2015) or on the onset time (Xu and Chan 2001). Different IODs might interact in different ways with different types of ENSO. Although the IOD is correlated with most ENSO events, its relationship with central Pacific (CP) events (or El Niño Modoki; Ashok et al. 2007) is somewhat obscure

and under debate. Zhang et al. (2015) highlighted that the relationship between the eastern Pacific (EP) El Niño and the IOD depends on the strength of ENSO, whereas for the central Pacific El Niño, its horizontal location is more important to forcing an IOD. Wang and Wang (2013, 2014) detected a new type of central Pacific El Niño based on its different impact on rainfall in southern China, and they then classified El Niño Modoki events into Modoki I and II. They further found that this type of central Pacific El Niño is associated with negative IOD events, which contrasts with the traditionally recognized ENSO–IOD relation. The complexity of the ENSO–IOD relation might occur because ENSO is a diverse continuum that exhibits substantial variations with regionally different feedbacks (Capotondi et al. 2015). To understand the relationship between various types of ENSO and IOD, we need to improve the understanding of the interbasin coupling of the Indo-Pacific Ocean during the developing phase of ENSO (Kug et al. 2006) and to consider the Indo-Pacific covariation. For example, the classification of IOD by Guo et al. (2015) was not based on different features of spatial patterns but on different evolutions of SST anomaly patterns. For Indo-Pacific coevolution of climate modes, how the various types of the IOD are related to the different types of ENSO remains unclear.

This study considered the Indian and Pacific SST covariation by performing rotated season-reliant empirical orthogonal function (S-EOF) analysis (Wang and An 2005). S-EOF analysis is a powerful method that can be used to diagnose the seasonal evolution of spatial structures; however, the orthogonal constraints among different modes sometimes lead to complexity of the spatial structures and difficulty in physical interpretation (Hannachi 2007). Any seasonal-evolution modes of ENSO would peak in boreal winter, which could cause nonorthogonality among the modes; hence, orthogonality was not the primary consideration. Thus, the rotated S-EOF is an appropriate choice in this study (Lian and Chen 2012). The rotated S-EOF relaxes the orthogonality constraints and can slightly modulate the results, making them more interpretable. In addition, the ENSO cycle has been recognized to have asymmetry in either amplitude or duration between El Niño and La Niña (Okumura and Deser 2010), which highlights the subject of determining the nature of the relationship between these asymmetric ENSO and IOD events in this study.

The remainder of this paper is organized as follows. Section 2 provides a brief introduction of the data, method, and models. Section 3 presents the two leading modes of ENSO development and the associated Indian Ocean variations. Section 4 shows the high-order modes of the Indo-Pacific SST evolution that reflects the

TABLE 1. CMIP5 models used to perform the S-EOF analysis of SST for the historical simulations. (Expansions of acronyms are available online at <http://www.ametsoc.org/PubsAcronymList>.)

Model	Institution and country
ACCESS1.0	Commonwealth Scientific and Industrial Research Organisation (CSIRO) and Bureau of Meteorology (BoM), Australia
CanESM2	Canadian Centre for Climate Modelling and Analysis, Canada
CCSM4	National Center for Atmospheric Research, United States
CESM1(BGC)	Community Earth System Model contributors, United States
CESM1(CAM5)	Community Earth System Model contributors, United States
CMCC-CESM	Centro Euro-Mediterraneo per I Cambiamenti Climatici, Italy
CMCC-CM	Centro Euro-Mediterraneo per I Cambiamenti Climatici, Italy
CMCC-CMS	Centro Euro-Mediterraneo per I Cambiamenti Climatici, Italy
CNRM-CM5	Centre National de Recherches Météorologiques, France
GFDL CM3	NOAA/Geophysical Fluid Dynamics Laboratory, United States
GFDL-ESM2G	NOAA/Geophysical Fluid Dynamics Laboratory, United States
GFDL-ESM2M	NOAA/Geophysical Fluid Dynamics Laboratory, United States
GISS-E2-H	NASA Goddard Institute for Space Studies, United States
GISS-E2-R	NASA Goddard Institute for Space Studies, United States
HadGEM2-ES	Met Office Hadley Centre, United Kingdom
HadGEM2-CC	Met Office Hadley Centre, United Kingdom
INM-CM4.0	Institute of Numerical Mathematics, Russia
IPSL-CM5A-LR	Institute Pierre-Simon Laplace, France
MIROC-ESM	Atmosphere and Ocean Research Institute (AORI), National Institute for Environmental Studies (NIES), Japan Agency for Marine-Earth Science and Technology (JAMSTEC), Japan
MPI-ESM-LR	Max Planck Institute for Meteorology, Germany
MPI-ESM-MR	Max Planck Institute for Meteorology, Germany
MRI-CGCM3	Meteorological Research Institute, Japan
NorESM1-M	Norwegian Climate Centre, Norway

asymmetric features between the IOD and El Niño or La Niña. Finally, a summary and discussion are presented in section 5.

2. Data, models, and method

This study used the observational SST dataset from the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST; Rayner et al. 2003). Other observational datasets include the monthly 850-hPa winds and sea level pressure from the National Centers for Environmental Prediction (NCEP)–NCAR reanalysis dataset (<http://www.esrl.noaa.gov/psd/>) and the absolute dynamic topography data from AVISO (<http://www.aviso.oceanobs.com/>). The time period for the observational data used in this study is 1948–2014.

For cross validation, SST data from historical simulations of 22 coupled general circulation models that participated in phase 5 of the Coupled Model Intercomparison Project (CMIP5) were also used. The details of these models are listed in Table 1. In addition, a 500-yr SST dataset from the CCSM4 model under the preindustrial scenario of CMIP5 was also analyzed. Two atmospheric general circulation models (AGCMs) were used to perform a series of sensitivity experiments. These were the Max Planck Institute for Meteorology (MPIM) ECHAM5

(Roeckner et al. 2003) and the National Center for Atmospheric Research Community Climate Model, version 3 (NCAR CCM3; Kiehl et al. 1998).

In climate research, EOF analysis (Lorenz 1956) is a tool commonly used to investigate the temporal evolution of a fixed spatial pattern; however, many climatic events have spatial patterns that vary temporally. S-EOF is an effective method for such nonstationary spatial patterns. Each eigenvector comprises several spatial patterns in consecutive months (e.g., from March to December) that share the same principal component. Thus they can represent the evolution with time. Otherwise, if we perform EOF for each month instead of doing S-EOF, each month would have a spatial pattern and a time series. These patterns or series may be similar, but each is different from others. Since there is no connection among the patterns, we cannot obtain evolution information of ENSO development. Therefore, performing S-EOF is the right approach. The S-EOF modes were further rotated to adjust the spatial patterns to make them more interpretable because the orthogonality among the modes was not our prime consideration.

Given the asymmetry between El Niño and La Niña evolutions, the rotated S-EOF analysis was applied to El Niño and La Niña events separately. Considering the

FIG. 1. (a)–(j) First mode of the rotated S-EOF for tropical Indo-Pacific SST anomalies (shading) and 850-hPa winds (vectors) regressed onto PC-1, (k) PC-1.

“Niño only” S-EOF analysis as an example, the SST anomaly data for all El Niño years were selected and then appended (but with reversed sign) to the Niño SST anomaly data along the temporal dimension before applying the S-EOF. Similar procedures were undertaken for the “Niña only” S-EOF. To test the effects, S-EOF analysis was also performed for the “all data” case. It was found that the first two modes of the Niño-only and Niña-only cases were very similar to the all-data case, whereas for the high-order modes, discrepancies exist between Niño-only and Niña-only cases, reflecting an asymmetry. Therefore, we analyze the leading two modes derived from the S-EOF analysis of the all-data case in section 3 and discuss the high-order modes for Niño-only and Niña-only cases in section 4.

3. Contrast between the two leading modes of Indo-Pacific SST anomalies

a. Different features

The first mode (mode 1) of the Indo-Pacific SST covariation, shown in Fig. 1, explains 47% of the total variance. Also drawn in Fig. 1 are the regressed 850-hPa wind anomalies onto the principal component (PC) time series. This mode shows a developing phase of ENSO starting from boreal spring (Fig. 1a) and peaking in winter. SST anomalies first appear near the South American coast (Figs. 1a–c) before extending westward

over the central-eastern Pacific. In the Indian Ocean, SST anomalies of the IOD pattern begin to form in early summer. Then, the IOD gradually develops with the developing ENSO and reaches its peak in boreal fall (Figs. 1g,h) and finally decays after October (Figs. 1i,j). This is a typical pattern of ENSO–IOD coevolution (Saji et al. 1999). El Niño excites an atmospheric teleconnection over the Indian Ocean and causes easterly wind anomalies conducive to the IOD during boreal summer and fall (Klein et al. 1999; Alexander et al. 2002).

The second mode (mode 2), which explains 15% of the total variance, exhibits an ENSO transition pattern within one year (Fig. 2). The El Niño SST warming first emerges in the equatorial Niño-3.4 region in late summer or early fall (Figs. 1f,g) before it intensifies and extends to the whole central-eastern Pacific. The S-EOF mode 2 is distinguished from mode 1 not only by its peak spatial pattern but also by its onset time. In mode 2, El Niño starts in July–August, much later than that in mode 1 (Fig. 1). A more noteworthy feature is that over the Indian Ocean mode 1 has a clearly positive IOD pattern starting in early summer and peaking in fall, but mode 2 does not show a well-formed IOD pattern concurrent with ENSO development. In fall (Figs. 2g–i), mode 2 has cold SST anomalies over the eastern Indian Ocean but no positive SST anomalies in the western Indian Ocean; thus it is not considered as an IOD. The disparity between the two modes implies that the early-onset El Niño events tend to be concurrent with positive IOD in

FIG. 2. As in Fig. 1, but for the second mode.

the following summer and fall, whereas late-onset El Niño does not.

The above conclusion is from observational data. To validate its robustness, we conducted a parallel S-EOF analysis using SST data from the historical scenario (during the period 1951–2005) of 22 CMIP5 models. We extracted the Niño-3.4 and IOD index from the spatial patterns of S-EOF for each model and showed them in Fig. 3. Each index was calculated from the spatial pattern of consecutive months (from March to December) by spatially averaging the SST anomalies over the region of index definition. The IOD index was calculated as the difference of SST anomalies between the west and the east pole of the Indian Ocean relevant to IOD events (Saji et al. 1999). Each thin gray line represents an individual model, and the thick black solid line with circles indicates the multimodel mean. It clearly shows that mode 1 is characterized by an early-onset El Niño associated with a remarkable IOD, whereas mode 2 is characterized by a late-onset El Niño with neutral IOD conditions. These results are similar to the observations, confirming the robustness of the two modes of ENSO–IOD covariation.

Although the mature phase of ENSO is firmly locked in boreal winter, the timing of its onset is somewhat irregular. Many previous studies have identified this, and some have classified ENSO into different types based on its timing of onset (Neelin et al. 2000; Xu and Chan 2001; Horii and Hanawa 2004). To further examine the relationship between ENSO onset time and the IOD development, the observed ENSO events during the

period 1950–2014 were grouped by their seasons of onset from boreal spring to fall. Then for each season, the percentages of ENSO cases with and without ensuing IOD events were calculated. These results are shown as stacked bars in Fig. 4a, where the black bars are for pure ENSO cases and the gray bars represent ENSO–IOD coexistence. This survey was based on the historical ENSO cases listed in Table 2. The ENSO year is identified by the normalized November–January (NDJ) averaged SST anomaly ≥ 0.8 . This is also reasonable for CMIP5 models because we have checked that all models used in this study have ENSO peaking during NDJ. The IOD occurrence is judged by the normalized IOD index ≥ 1.0 , and the criterion of ENSO onset is 3 consecutive months of the averaged Niño-3.4 index ≥ 0.5 . According to this survey, the numbers of ENSO starting in the three seasons are comparable: 12 events in spring, 16 in summer, and 14 in fall. As the onset time of ENSO gets later, the following IOD events are fewer. Above 90% of spring-type ENSO events are accompanied by IOD events, while the percentage falls to 50% and 20% for summer- and fall-type events, respectively. We also did such a survey for the historical run (1951–2005) of each CMIP5 model and calculated the multimodel mean of the percentages and ENSO intensities. These results are shown in Fig. 4b, which shows the same relationship between ENSO onset time and IOD occurrence as in Fig. 4a.

Hereafter, ENSO events are classified into two groups according to mode 1 and mode 2: the early-onset events

FIG. 3. (a),(c) Niño-3.4 and (b),(d) IOD indices from the monthly spatial pattern of S-EOF (a),(b) mode 1 and (c),(d) mode 2 using the SST from preindustrial simulations of 25 CMIP5 models (gray lines). The multimodel mean is denoted by the thick black solid line, and the red dashed line is the zero line.

forming in boreal spring–early summer and the late-onset events forming in late summer–fall. It might be speculated that the reason why most of the early-onset ENSO events tend to be accompanied by IOD events is that such events generally have larger intensity than late-onset ENSO events (Horii and Hanawa 2004; Sooraj et al. 2009). This is because the time of ENSO development through Bjerknes feedback is longer before the ENSO event reaches its peak in boreal winter;¹ therefore, ENSO events with larger magnitude could easily force an IOD. This can also be seen from the red line in Fig. 4, which denotes the group-averaged intensity of the Niño-3.4 index in the peak season (NDJ). This is also the reason why early-onset and late-onset ENSO events manifest in the first and second modes, respectively.

Is the contrast in ENSO magnitude the only reason for the distinct IOD developments with the two types of ENSO? To answer this question, we analyzed a 500-yr coupled simulation using the CCSM4 model under the preindustrial control scenario of CMIP5. The preindustrial dataset is chosen because it has a longer period than observations and the historical run. To examine the influence of mere ENSO timing on the IOD, the cases with Niño-3.4 indices of equal magnitude in August for the early onset and later onset of the ENSO event were selected and their associated IOD evolutions

were plotted (Fig. 5). The El Niño (La Niña) events with a Niño-3.4 value in April above (below) zero were selected as spring-type events, while those with a Niño-3.4 value in April below (above) zero were selected as late-onset events. Then, for both types, a further selection of those events with Niño-3.4 values in August of 0.5–1.2 was performed. This ensured that all ENSO events had

FIG. 4. Percentages of ENSO–IOD concurrent cases (black bars) and ENSO-only cases (gray bars) in ENSO events starting in three seasons, and averaged winter (DJF) ENSO intensity (red line) for the three season-starting types of ENSO. (a) Observation and (b) multimodel (historical run; 1951–2005) averaged percentage and ENSO intensity.

¹ Thus the value of Niño index in late spring has often been referred to as the indication of the amplitude that ENSO can finally reach by the end of that year.

TABLE 2. Year list of IOD concurrence or not for ENSO events starting from different seasons during 1950–2014.

	Spring type (12 events)		Summer type (16 events)		Fall type (14 events)	
	With IOD	Without IOD	With IOD	Without IOD	With IOD	Without IOD
El Niño	1972	1957	1951	1965	2006	1968
	1982		1963	1969		1979
	1987		1977	1976		1986
	1997		1994	2004		1991
			2002	2009		2014
La Niña	1950		1970	1973	1974	1949
	1954		1998	2007	1984	1967
	1955		2010	2011		1983
	1964					1995
	1971					1999
	1975					2000
	1988					

comparable magnitudes during boreal fall to winter, which controlled the interference of ENSO magnitude, and thus produced a better comparison between the two types of ENSO events. For early-onset ENSO events, Fig. 5b shows that the indices of positive IOD (associated with El Niño) clearly deviate from negative IOD (associated with La Niña), indicating evident IOD development from summer to fall. However, for the late-onset type of ENSO events, although their intensities after late summer have the same magnitude as those of the spring-type events, the positive and negative IOD indices are quite diverse and the multicase average is neutral. The contrast between Figs. 5b and 5d strongly indicates that even when the two types of ENSO events have the same intensity in fall and winter, the spring-onset events are more conducive to the development of IOD events than summer-onset events. Therefore, the difference in magnitude between the two types of ENSO events is not the only cause of different IOD behaviors, and the timing of ENSO onset is also important.

b. Mechanisms

We have shown that early-onset and late-onset ENSO events can cause different IOD development even when the two types of ENSO events have reached to the same degree of intensity by boreal fall. A natural question arises: Why is an early-onset ENSO event more inducible to the subsequent IOD than a late-onset ENSO event of comparable magnitude? Since the main difference between the two types of ENSO events shown in Figs. 5a and 5c is the summer Niño index, we can thus speculate that the summer state might be important in causing this disparity of the IOD. For the formation of an IOD, an initial SST anomaly triggering the subsequent positive feedback is necessary. It can be seen from Fig. 1 that during the period of early-onset ENSO development, the Indian Ocean warm SST

anomalies first emerge over the Arabian Sea in early summer. These warm SST anomalies sustain and extend from the northwestern Indian Ocean to most of the western Indian Ocean, persisting into fall (around October) when the western Pacific cyclonic anomaly in summer converts to an anticyclonic anomaly. The easterlies on the south flank of the anticyclone enhance surface easterlies over the Indian Ocean and intensify eastern Indian Ocean cooling, thereby bringing the IOD to its peak. In the case of late-onset ENSO (Fig. 2); however, when El Niño is initially established (Fig. 2g) in late summer or early fall, the western Indian Ocean warming was never fully established throughout the fall. Without the collaboration of the western Indian Ocean warming, the IOD pattern is not significant, though the easterlies are already developed by fall (Figs. 2h,i). Based on the above analysis, it can be seen that the initial summer warming over the northwestern Indian Ocean might be a key to the development of an ENSO-forced IOD. Two further questions are the following: How does the timing of ENSO onset cause different IOD responses? What is the relation between the initial northwestern Indian Ocean warming and ENSO forcing in different seasons? These might be explained through two possible mechanisms: 1) different atmospheric responses over the Indian Ocean to the timing of ENSO onset and 2) different Indian Ocean dynamic response to different timing of ENSO onset.

1) ATMOSPHERE BRIDGES

To examine the atmospheric teleconnections under the two types of modes, a series of simulations were performed using two AGCMs: ECHAM5 and CCM3. We noted that the two types of ENSO differ not only in the onset time but also in the spatial pattern of the eastern Pacific SST. The maximum SST center over the eastern Pacific in mode 2 locates more westward than that in mode 1. To clarify these two factors' (ENSO

FIG. 5. (a),(c) Niño-3.4 evolution of developing El Niño (red) and La Niña (blue) cases and (b),(d) corresponding IOD indices. Both types of indices are normalized. The ENSO events are classified into two groups; (a),(b) spring type and (c),(d) summer type, based on their onset time. For both types of ENSO, a screening procedure was performed to select cases of equal magnitude of Niño-3.4 index in August (within the span 0.5–1.2). The SST data are from a 500-yr simulation within the CMIP5 preindustrial experiment using CCSM4.

onset time and SST spatial pattern) contributions to the Indian Ocean, we designed two types of experiments and each is forced by different SSTs. One is forced by the SST pattern of early El Niño as in S-EOF-1, which is the same as Fig. 1d but only over the equatorial central-eastern Pacific region (10°S – 10°N , 160°E – 60°W ; see Fig. 6e). The other is forced by the SST pattern of later-onset ENSO (S-EOF-2) as in Fig. 2f but only over the equatorial central-eastern Pacific region (10°S – 10°N , 160°E – 60°W ; see Fig. 6f). To make the two kinds of experiments generate atmospheric responses of comparable magnitude, we set the two SST patterns to have the same root-mean-square (RMS) before running the models. The SST anomaly was steadily imposed over the central-eastern Pacific, while elsewhere seasonally varying climatological SST was imposed. For each model, 20 realizations were performed and each was initialized with different atmospheric conditions. To check the seasonal dependency, the atmospheric response to the SST anomaly in every month was examined, and it was found that the Indian Ocean atmospheric responses to central-eastern Pacific warming in late spring to early summer are quite distinct (almost opposite) in comparison with those in late summer to fall. Thus we selected the results of June and August to represent early summer and late summer results, respectively. The results for ECHAM5 and CCM3 are shown in Fig. 6 and Fig. S1 in the supplemental material, respectively. The variables drawn are the responses of lower-level (850 hPa) wind vectors, wind speeds

(shading), and sea level pressure (contours). These results are very consistent between the two models. It turns out that the two different SST patterns produce very similar atmospheric response patterns in the same season (see the similarity between Figs. 6a,c and Figs. 6b,d), while each SST pattern produces contrasting atmospheric responses in different seasons (see the contrast between Figs. 6a,b and Figs. 6c,d). The results of the other model, CCM3 (Fig. S1), show the same conclusion. Thus the seasonal dependence of atmospheric teleconnections is more important than the El Niño SST pattern differences between S-EOF-1 and S-EOF-2.

In early summer (Figs. 6a,b), the central-eastern Pacific warming can produce an anticyclonic circulation anomaly over the Saudi Arabian peninsula and the Arabian Sea. This wind mode is reminiscent of the asymmetric mode of spring Indian Ocean rainfall and wind variability reported by Wu et al. (2008) but with greater emphasis on the northern Indian Ocean. The strong northeasterly wind anomalies on the southeast flank have the opposite direction to the climatological monsoonal winds (Somali jet) in early summer; hence, weakened summer monsoonal wind speeds can reduce the surface evaporation and cause relatively less latent heat loss from the ocean. Furthermore, reduced southwesterly winds can suppress upwelling off East Africa (Yuan et al. 2008) and are favorable for the warming of the northwestern Indian Ocean. In conjunction with the anomalous circulation, negative rainfall and positive outgoing longwave radiation anomalies (figure not

FIG. 6. (a),(b) Early summer (June) and (c),(d) late summer (August) responses of 850-hPa wind (vectors), wind speed (shading), and sea level pressure (contours) anomalies to El Niño SST anomalies (SSTA) from the two S-EOF modes. (a),(c),(e) The SSTA pattern is from S-EOF-1 (Fig. 1d) over the central-eastern Pacific region of 10°S–10°N, 160°E–60°W, while (b),(d),(f) the SSTA pattern is from S-EOF-2 (Fig. 2f) over the region of 10°S–10°N, 160°E–60°W. The SSTA pattern in (f) is scaled to ensure the two patterns in (e),(f) have the same magnitude (measured by root-mean-square). These results are obtained using the model ECHAM5. Positive and negative values are denoted as yellow and green lines, respectively. The contour interval is 20 Pa and the zero lines are omitted.

shown) were also found in the northwestern Indian Ocean, which might also facilitate the western Indian Ocean SST warming. The 850-hPa wind anomaly pattern over the Indian Ocean in June (Figs. 6a,b) is similar to the first EOF mode of the corresponding months (Figs. 1c,d). This indicates that initial northwestern Indian Ocean warming in late spring and early summer is indeed the result of remote El Niño forcing before air-sea coupling of the Indian Ocean takes effect.

In contrast to early summer, the late-summer responses exhibit a completely different situation consistently shown by both models (Figs. 6c,d). The Arabian Sea anticyclone disappears, and it is replaced by cyclonic circulation with negative sea level pressure anomalies over northeastern India and the Bay of Bengal. Correspondingly, over the northern Indian Ocean, westerly anomalies are evident to the south of the negative sea level pressure center, which is unfavorable for the western Indian Ocean warming. It is noteworthy that the cyclonic anomaly over northeastern India is associated with warm SST in the eastern Pacific. In fact, this does not contradict the well-known negative correlation

between Indian rainfall and El Niño. The observational correlation map with El Niño in late summer also shows a cyclone as well as a positive rainfall anomaly over northeastern India (figure not shown); however, all Indian rainfall is negatively correlated with El Niño.

The seasonal disparity between the atmospheric responses (Figs. 6a–d) might be attributable to seasonally different mean climatic states. Boreal fall is the season when the climatological eastern Pacific SST reaches its minimum (Rasmusson and Carpenter 1982); thus, SST anomalies of the same magnitude would produce weaker convection anomalies than in early summer. This can be seen from a further analysis of the 200-hPa geopotential height anomalies on a broad scale in Fig. 7, which shows the simulated atmospheric response to the SST anomaly pattern as in Fig. 6e using the model ECHAM5. Since the two models (ECHAM5 and CCM3) and the two SST anomaly patterns (Figs. 6e,f) produce similar responses patterns, Fig. 7 shows only the ECHAM5 responses for simplicity. We can see much weaker anomalies in late summer than in early summer, and the remote Pacific–North American (PNA) pattern is even indiscernible in

FIG. 7. (a) June and (b) August responses of 200-hPa geopotential height (shading), sea level pressure (SLP; contours), and 850-hPa wind (vectors) to the same El Niño SSTA pattern as in Fig. 6e, simulated by ECHAM5. Positive and negative SLP anomaly values are denoted as red and green lines, respectively.

August (Fig. 7b). The disappearance of PNA may be due to the absence of the westerly jet in late summer (Kumar and Hoerling 1998). The August responses (Fig. 7a) are in contrast with early summer when the PNA response still exists (Fig. 7a), similar to the situation in boreal spring (figure not shown). A noteworthy feature is over the northwestern tropical Pacific and the north Indian Ocean: the barotropic response over the northwestern Pacific in June becomes a baroclinic response in August, with lower-level cyclone anomalies extending from the northwestern Pacific to the northern Indian Ocean. The teleconnection between the central-eastern Pacific warming (cooling) and the northwestern Pacific cyclone (anticyclone) anomaly in late summer is in line with previous studies of Fan et al. (2013) and Xiang et al. (2013). This seasonal transition may be related to the seasonal variation of the vertical shear of this region between June and August. However, the exact reason is not clear and needs further study. The seasonality of the ENSO–Indian Ocean teleconnection implies that if an El Niño starts in late summer or fall, it might lose the opportunity of heating the northwestern Indian Ocean SST sufficiently to trigger a positive IOD event.

The above analysis highlights the influence of ENSO timing on IOD development via an atmospheric bridge. It is suggested that ENSO events starting before early summer could easily warm the northwestern Indian Ocean and provide the necessary IOD preconditions. Subsequently, the summer mean state is also favorable for the further development of the IOD via the strengthening of the equatorial zonal wind anomaly and

dynamic Bjerknes feedback. This has also been elaborated by Xiang et al. (2011). They emphasized that the mean precipitation in boreal summer has greater equatorial restriction than in winter, which causes an equatorward shift of the anomalous precipitation center relative to the cold SST center. Moreover, the vertical shear over the Indian Ocean in summer is the monsoonal easterly vertical shear, which can promote a greater lower-level wind response than in other seasons. This is because the easterly vertical shear might induce a barotropic mode response (Wang and Xie 1996; Li 2006), which superposes on the mean baroclinic mode and leads to a reduction (increase) of intensity of the upper-level (lower level) Rossby wave response. Xiang et al. (2011) explained why the IOD always develops in boreal summer rather than in winter. The present study demonstrates that, for a developed eastern Pacific SST anomaly to trigger an IOD event, the mean state in early summer or spring is more favorable than late summer or fall.

2) EFFECTS OF OCEAN DYNAMIC WAVES

In addition to the atmospheric bridge, ocean dynamics also play a role in modulating the interannual SST variability in the Indian Ocean (Masumoto and Meyers 1998; Chambers et al. 1999; Xie et al. 2002). Both ENSO and IOD are always associated with the propagation of ocean Rossby waves from east to west across the Indian Ocean (Xie et al. 2002; Yu et al. 2005). In particular, Rossby waves around Seychelles in the tropical southwestern Indian Ocean (6° – 10° S) are very important in air–sea coupling because the thermocline in that region

FIG. 8. Longitude–time section of ADT anomalies (0.5-cm contours; zero lines omitted) and SST anomalies (shading) averaged over 6°–10°S of the tropical Indian Ocean, corresponding to S-EOF (a) mode 1 and (b) mode 2. The SST values are from the S-EOF results shown in Figs. 1 and 2, while the ADT values are interannual ADT anomalies in consecutive months regressed onto the S-EOF PC-1 (Fig. 1k) and PC-2 (Fig. 2k). The solid lines denote positive values while the dashed denote negative values.

is shallow (hence it is called the Seychelles dome), and thus ocean wave anomalies can easily influence SST. Considering the fall situation of Figs. 2g–i, the main feature is the cold western Indian Ocean. Therefore, it is speculated that downwelling ocean Rossby waves might not arrive in the western Indian Ocean (or do arrive but at an inappropriate time) to be manifested in the western Indian Ocean SST. Based on this conjecture, the propagation of absolute dynamic topography (ADT) anomalies in the two S-EOF modes were examined.

Figure 8 shows the longitude–time section of ADT anomalies (contours) and SST anomalies (shading) averaged in the belt of 6°–10°S in the tropical Indian Ocean, corresponding to the first (Fig. 8a) and second (Fig. 8b) S-EOF modes. The SST values are from the S-EOF results shown in Figs. 1 and 2, while the ADT values are the interannual ADT anomalies in consecutive months regressed onto the S-EOF first principal component (PC-1) (Fig. 1k) and PC-2 (Fig. 2k). It can be seen from Fig. 8a that the positive ADT anomaly (downwelling Rossby wave) first exists in March and April near 90°E, excited by the easterly wind anomalies over the eastern Indian Ocean (see Figs. 1a,b). It slowly propagates westward and then rapidly reaches the western Indian Ocean in June and July. Such an abrupt acceleration implies that it is not simply a freely

propagating Rossby wave but a wind-forced Rossby wave. The downwelling Rossby wave arrives in the Seychelles dome region and causes the warming of the southeastern Indian Ocean (Figs. 1d,e) in boreal summer. Therefore, the warm SST extends from the northwestern Indian Ocean to the southwest, favoring an IOD development. For the second mode, however, the downwelling Rossby wave seems more like a free Rossby wave: it propagates slowly and does not arrive in the southwestern Indian Ocean by boreal summer; thus, warm SST anomalies do not appear in the western Indian Ocean. Even in fall when the Rossby wave reaches the Seychelles dome region, positive SST anomalies are not immediately manifested. This might be because the depth of the thermocline here is seasonally deeper in late fall and winter than in early summer (Rao and Behera 2005). Consequently, the positive anomalies are mainly confined over the central-southern Indian Ocean (see Figs. 2g–i) rather than the western Indian Ocean, and hence the IOD is not fully developed.

In summary, the seasonal dependence of the atmosphere responses over the Indian Ocean to different timing of ENSO onset together with the different ocean dynamic Rossby waves causes the disparity in IOD developments in the two leading modes. The atmospheric responses to eastern Pacific warming have seasonal

FIG. 9. As in Figs. 1 and 2, but for the spatial pattern of the third mode of rotated S-EOF for the Niño-only case.

dependence. If El Niño is formed by early summer, it can produce a lower-level anticyclone over the northwestern Indian Ocean with a weakened Somali jet, which warms the northwestern Indian Ocean and facilitates the IOD development. This is the reason why most IOD events concurrent with ENSO have initial SST warming in the northwestern Indian Ocean (Yuan et al. 2008). In addition, the westward-propagating ocean Rossby waves arrive in the Seychelles dome region in the southwestern Indian Ocean by summer with the aid of atmosphere–ocean coupling, thereby inducing the southwestern Indian Ocean warming and further promoting IOD development. However, if an El Niño of the same magnitude starts to form in late summer or fall, it will maintain a lower-level cyclone over the South Asian subcontinent, which is not instrumental in warming the northwestern Indian Ocean. For the southwestern Indian Ocean, westward-propagating Rossby waves cannot arrive at the Seychelles dome region until late fall; thus, positive SST anomalies do not appear in the southwestern Indian Ocean, and consequently an IOD pattern does not completely develop. Therefore, early-onset El Niño events can more easily force northwestern Indian Ocean warming and trigger IOD development than late-onset events. This explains why the IOD events have quite different (Figs. 5b,d) responses to ENSO events of the same magnitude (Figs. 5a,c).

4. Other ENSO–IOD covariation modes

The spatial patterns of the first two rotated S-EOF modes are largely symmetric for El Niño and La Niña

events. As for those of high-order modes, however, the results for El Niño and La Niña cases are different because of the asymmetry of ENSO evolution. This study obtained the third and fourth modes for both El Niño and La Niña events. Among these, the third mode (Fig. 9) for the Niño-only case (which accounts for 7.7% of the total variance) is of particular interest. It depicts a process of development of a special type of El Niño, which has asymmetric warm SST extending from the northeastern Pacific to the equatorial central Pacific. During its peak phase (Figs. 9i,j), the warm SST locates far west beyond the date line, and there is no significant warming in the eastern Pacific. The SST propagation within the northeastern subtropics to central tropics pathway has been reported by previous researchers (Yu and Kao 2010), and this type of far-west CP El Niño is like the EOF-2 of the tropical Pacific SST as shown by Ashok et al. (2007). Wang and Wang (2013) have classified El Niño Modoki events into Modoki I and II. According to their classification, the S-EOF-2 mode in this study (Fig. 2) is like El Niño Modoki I. Their recent study (Wang and Wang 2014) reported that most of the observed El Niño Modoki II cases are associated with negative IOD in boreal summer and fall. The present results in this study also reveal such a relationship between ENSO and the IOD, which confirms the previous results based on observational composite analysis. In addition, the regression of the preceding Indian Ocean SST onto this PC also exhibits such a negative IOD pattern (figure not shown).

The CP El Niño–IOD relationship has long been considered to be obscure because there appears to be no

significant statistical correlation between them. Zhang et al. (2015) suggested that the CP El Niño–IOD relationship depends mainly on the zonal locations of the SST anomalies rather than on their magnitudes. This implies that the classification of El Niño based on the location of the SST anomaly center is instrumental in illuminating the CP El Niño–IOD relationship. In this regard, the El Niño Modoki II has a distinguishing feature—the westernmost location of the warm center among all types of El Niños. This feature might be caused by the special formation mechanism of El Niño Modoki II. The SST warming originates from the northeastern Pacific in boreal spring (Figs. 9a,b), before extending gradually to the equatorial central Pacific through the wind–evaporation–SST (WES) feedback mechanism (Chang et al. 2007; Wu et al. 2010). According to Wang and Wang (2014), El Niño Modoki II is characterized by the displaced Walker circulation with an ascending branch over the eastern Indian Ocean. This ascending branch brings greater-than-normal precipitation to the west coast of Sumatra (see Fig. 6c of Wang and Wang 2014). The rainfall anomaly excites a pair of cyclones symmetrically about the equator over the eastern Indian Ocean (Figs. 9f–j). Then, the equatorial westerly anomalies trigger the Bjerknes feedback of the Indian Ocean basin, sustaining the negative IOD. Under such a situation, it can be noted from Fig. 9 that the northwestern Pacific cyclone in summer does not turn into the fall anticyclone and that this is a feature distinct from other types of El Niño events.

It is noteworthy that in Fig. 9, the negative IOD already exists in spring and summer, while the El Niño Modoki II forms in fall or winter. In this line of reasoning, El Niño Modoki II might not be the initial forcing of the negative IOD, which prompts the question of whether there is a physical connection between the northern Pacific meridional mode and the IOD. To date, only four concurrent cases of El Niño Modoki II and negative IOD have been observed, and thus a further study of this topic is needed.

As for the negative phase of the teleconnection, a parallel S-EOF analysis was performed for Niña-only cases, but the relation between La Niña Modoki II and positive IOD was not observed (Fig. S2 in the supplemental material; accounting for 10.4% of the total variance). This Niña-only mode slightly resembles the negative phase of ENSO Modoki II but with weaker magnitude in its peak season. Whether the relationship between El Niño Modoki II and negative IOD also holds for La Niña Modoki II is open to further research.

The fourth EOF mode for Niño-only cases (Fig. S3 in the supplemental material; accounting for 5.6% of the total variance) shows a CP El Niño developing from the

southeastern Pacific Ocean. The warming center in its peak season is located at 150°W. This type of El Niño can be classified as El Niño Modoki I according to Wang and Wang (2013); however, Fig. S3 does not show an IOD accompanying the El Niño event. This might be because the central Pacific warming starts in late summer and loses the opportunity of heating the western Indian Ocean SST. Therefore, although the equatorial easterlies over the eastern Indian Ocean are developed in fall, they fail to extend to the western Indian Ocean because of the lack of influences from western Indian Ocean warming. This reasoning is in agreement with the argument regarding the relationship between ENSO onset season and IOD development in section 3.

The fourth mode for La Niña only (Fig. S4 in the supplemental material; which accounts for 3.7% of the total variance) shows a persistent La Niña throughout the year. The atmospheric and SST anomalies over the Pacific and Indian Oceans remain steady, indicating that no positive or negative IOD events develop during the persistence of La Niña. It is reasonable that persistent La Niña instead of persistent El Niño features in the rotated S-EOF mode because there is asymmetry in the durations of El Niño and La Niña events. After the peak season of ENSO, most El Niño events decay rapidly by the following summer; however, many La Niña events persist throughout the following year and often re-intensify in the subsequent winter (Okumura and Deser 2010), as shown in Fig. S4. This further confirms that the S-EOF results for Niño-only and Niña-only cases are reasonable.

5. Summary and discussion

To examine the coevolution between various types of El Niño and the IOD, we have shown several modes of season-reliant EOF by taking the Pacific and Indian Oceans together using both observations and coupled model data. The leading two modes indicate that the early-onset ENSO can more easily force an IOD event than late-onset ENSO, even if they are of the same magnitude. Both the atmospheric bridge and ocean dynamic waves are responsible for this observed feature. A set of model experiments using two AGCMs with SST prescription over the central-eastern Pacific in different seasons reveal that the atmospheric responses to the central-eastern Pacific SST anomalies have contrasting seasonal dependence. If an El Niño event has already formed by early summer, it can produce a lower-level anticyclone over the northwestern Indian Ocean with weakened monsoonal winds over Somalia, thereby warming the northwestern Indian Ocean by reduced evaporation and wind-driven upwelling along the coast

of East Africa. The warmed northwestern Indian Ocean SST facilitates subsequent IOD development through the Bjerknes feedback over the Indian Ocean. Meanwhile, the westward-propagating downwelling ocean Rossby waves reach the thermocline dome region in the southwestern Indian Ocean by summer, further favoring the southwestern Indian Ocean warming. Conversely, if an El Niño event starts in late summer or fall, it produces an anomalous lower-level cyclone over the southern Indian subcontinent without weakening the Somali southwesterly jet, hence losing the opportunity of warming the western Indian Ocean. Furthermore, westward-propagating ocean waves cannot reach the thermocline dome region in the southwestern Indian Ocean by fall; hence, warm SST anomalies are not found in the western Indian Ocean and the IOD does not develop. Ultimately, late spring and early summer represent the optimum window for ENSO formation to affect IOD development, whereas late-onset ENSO events are not favorable for allowing the Bjerknes feedback in the Indian Ocean to develop an IOD.

The above relationship between the timing of the ENSO onset and the IOD is applicable to both El Niño and La Niña. In consideration of the ENSO asymmetry, a rotated S-EOF analysis was also performed separately for Niño-only and Niña-only cases. It was found that El Niño and La Niña events exhibit different evolutionary features that are manifested in high-order modes, among which an interesting mode of El Niño–IOD coevolution was identified. The third mode of El Niño evolution, characterized by a central Pacific El Niño pattern, developed from the northern Pacific meridional mode with a warm center over the far-west date line previously identified as El Niño Modoki II (Wang and Wang 2013). Interestingly, it is accompanied by a negative IOD event, which is the opposite of most other ENSO events. Previous studies (Wang and Wang 2014) have also reported such cases in their studies, and the present S-EOF results confirmed this teleconnection from another perspective.

AGCM experiments were used in this study, though the air–sea coupling is essential for IOD development. However, the AGCM modeling approach is fine because our focal point is not the process of IOD development. Instead, we focus on the beginning when the Indian Ocean atmosphere initially responds to the Niño SST forcing. This initial atmospheric response to the El Niño SST anomalies can be reasonably illustrated by AGCMs. It is found that the early-onset (late onset) ENSO can (cannot) produce atmospheric circulations favorable for warming the northwestern Indian Ocean, which is also seen in the S-EOF analyses of observations. The western Indian Ocean warming is an important

trigger for the subsequent IOD development via the air–sea coupling process.

The impact of the timing of ENSO onset has long been a topic of research. Many previous studies (Xu and Chan 2001; Sooraj et al. 2009) have noted its influence on ENSO strength, subsequent ENSO duration, and the Indian Ocean SST. Some previous studies have reported that early-onset ENSO events tend to have larger influence on the Indian Ocean than late-onset events, but they have all attributed this to the disparity in ENSO strength. However, the present study shows that, even when the two types of ENSO are of the same magnitude, they still can generate distinct IOD responses. The reason is the seasonal dependence of the atmospheric responses to El Niño due to different mean climatic states. An El Niño that forms before early summer can more easily warm the northeastern Indian Ocean. Although the mechanism of ENSO influence on the Indian Ocean has been elaborated by previous studies (Yuan et al. 2008; Wu et al. 2008), the present study has emphasized that this mechanism could vary with season. It is concluded that late spring to early summer is the important time window. El Niño events that start to form after this period might not warm the western Indian Ocean sufficiently and hence not trigger the mechanism of IOD development. Xiang et al. (2011) highlighted that the summer mean state is more important to IOD development than the winter mean state, whereas the present study focuses on the initial triggering, before air–sea coupling takes effect and leads to the development of an IOD. Comparison of the periods of early summer and late summer shows that the former period has the greater advantage in helping an ENSO event trigger an IOD.

The present study investigates the Indo-Pacific SST from a holistic perspective and obtains seasonal coevolution modes of ENSO and IOD during the ENSO development phase. It links various types of ENSO and IOD events that have been documented separately in previous studies. For example, the second S-EOF mode has a central Pacific El Niño pattern, which is similar to El Niño Modoki I identified by Wang and Wang (2013) but with a late onset time. The Indian Ocean pattern in mode 2 is reminiscent of IOD Modoki documented by Endo and Tozuka (2015), with cold SST over the eastern and western Indian Ocean but warm SST over the central-southern Indian Ocean; however, it is not identified as a normal IOD event in this study. We also identified El Niño Modoki II (Wang and Wang 2014) and confirm its relationship with negative IOD events.

One may note that the El Niño in mode 2 evolves from the peak phase of La Niña and the Indian Ocean

experiences a decaying process of basinwide cooling. The preceding cold conditions of the Indian Ocean may also obstruct the positive IOD events. The Indian Ocean cooling can force an anomalous cyclone anomaly located over the northwestern Pacific from spring to summer (see Figs. 2a–g; see also Xie et al. 2009). This cyclone anomaly is not instrumental for easterly wind anomalies over the eastern Indian Ocean. In addition, the cold Indian Ocean condition increases the difficulty of the emergence of the western Indian Ocean warming. Further study of this issue is under way in a separate paper on different post-IOB evolutions by the authors of this paper.

We note that most of the spring type ENSO events occurred before the 1990s, whereas more summer and fall type occurred after the late 1990s (see Table 2). We speculate that this phenomenon is related to the frequent occurrence of central Pacific (CP) El Niño after 1990 (Yeh et al. 2009), as the onset time of CP El Niño is usually later than that of eastern Pacific (EP) El Niño (Kao and Yu 2009). This shift may be caused by the weakening of the zonal slope of the equatorial thermocline and a weaker eastern Pacific cold tongue, which is shown in the CMIP5 simulations under global warming scenarios (Yeh et al. 2009; Xie et al. 2010). In addition, North Pacific Oscillation (NPO)-like atmospheric circulations during spring become more effective in initiating a subsequent CP El Niño since the 1990s (Yu and Kao 2010; Yeh et al. 2015). However, given the substantial uncertainty regarding the anthropogenic impacts and natural decadal variability (Capotondi et al. 2015), it is not yet clear to what extent observed ENSO modulation is a cause or a consequence of either anthropogenic or intrinsic decadal-scale changes in the equatorial Pacific mean state.

The results presented in this study improve the understanding of the ENSO–IOD relationship and the documents on the relationship between ENSO onset and subsequent IOD, which can provide the basis for forecasting the ensuing IOD based on El Niño/La Niña information. However, this study had some limitations. The S-EOF procedure mainly addresses those concurrent IOD–ENSO events or pure ENSO events without significant anomalies in the Indian Ocean, because the Pacific SST anomalies have larger magnitudes and greater persistency than Indian Ocean SST anomalies. Therefore, this study does not deal with the evolution of pure IOD events that are not accompanied by a persistent SST evolution in the Pacific. In addition, the AGCMs used in this study can only deal with the triggering mechanism but cannot involve the processes of air–sea coupling during the IOD development.

Additional studies are needed to further reveal the relationship between the Pacific Ocean and IOD.

Acknowledgments. This work is supported by the Ministry of Science and Technology of China [National Basic Research Program of China (2012CB955603 and 2013CB430301)], the Strategic Priority Research Program of the Chinese Academy of Sciences (XDA11010203), the Natural Science Foundation of China (41406001, 41476003, and 41490643), the Fundamental Research Funds for the Central Universities (201562030), the National Programme on Global Change and Air–Sea Interaction (GASI-IPOVAI-04), NSFC-Shandong Joint Fund for Marine Science Research Centers (U1406401), and the Pioneer Hundred Talents Program of the Chinese Academy of Sciences.

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