Dominant Interannual Covariations of the East Asian–Australian Land Precipitation during Boreal Winter

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ABSTRACT

The present study applies the empirical orthogonal function (EOF) method to investigate the interannual covariations of East Asian–Australian land precipitation (EAALP) during boreal winter based on observational and reanalysis datasets. The first mode of EAALP variations is characterized by opposite-sign anomalies between East Asia (EA) and Australia (AUS). The second mode features an anomaly pattern over EA similar to the first mode, but with a southwest–northeast dipole structure over AUS. El Niño–Southern Oscillation (ENSO) is found to be a primary factor in modulating the interannual variations of land precipitation over EA and western AUS. By comparison, the Indian Ocean subtropical dipole mode (IOSD) plays an important role in the formation of precipitation anomalies over northeastern AUS, mainly through a zonal vertical circulation spanning from the southern Indian Ocean (SIO) to northern AUS. In addition, the ENSO-independent cold sea surface temperature (SST) anomalies in the western North Pacific (WNP) impact the formation of the second mode. Using the atmospheric general circulation model ECHAM5, three 40-yr numerical simulation experiments differing in specified SST forcings verify the impacts of the IOSD and WNP SST anomalies. Further composite analyses indicate that the dominant patterns of EAALP variability are largely determined by the out-of-phase and in-phase combinations of ENSO and IOSD. These results suggest that in addition to ENSO, IOSD should be considered as another crucial factor influencing the EAALP variability during the boreal winter, which has large implications for improved prediction of EAALP land precipitation on the interannual time scale.

1. Introduction

The Asian–Australian monsoon (AAM) is one of the most powerful monsoon systems, forming due to the pronounced thermal contrast between the Eurasian continent and the Indo-Pacific Ocean (e.g., Wang et al. 2003; T. Zhou et al. 2009). The variability of the AAM critically influences the economy and society across the region (e.g., Wang 2006). Thus, a better understanding of the AAM variability can help to improve the accuracy of prediction and has the potential to reduce the

socioeconomic loss from extreme events associated with abnormal AAM.

The AAM system consists of some mutually connected regional monsoons (e.g., Chen et al. 2000; Wu and Wang 2000; Hamada et al. 2002; Wang et al. 2003; Wang et al. 2009; Gong et al. 2015). During the boreal winter (December–February), the dry and cold monsoon flows from East Asia (EA) cross the equator to reach Australia (AUS), during which the air mass picks up moisture from the warm oceans and feeds strong monsoon rains over the Australian continent and surrounding regions (e.g., Zhang and Zhang 2010). The lower-tropospheric cross-equatorial flows, therefore, link regional monsoons between two hemispheres and induce the large-scale atmospheric circulation and

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precipitation covariations over the East Asia and Australia (EA–AUS) region. The monsoonal precipitation in the EA–AUS region is also large during boreal winter especially in the Maritime Continent and northern Australia compared with that boreal summer (e.g., Cai and van Rensch 2013). Though the total winter precipitation in southern China accounts for more than 10% of annual total precipitation, it has a profound impact on the traffic, agriculture, economy, and daily life due to its relatively large amplitudes of interannual variability (Wang and Feng 2011; Ge et al. 2016). For example, in 2007/08 winter, southern China suffered from extreme freezing rain and snow, which induced large losses of life and property (e.g., W. Zhou et al. 2009).

Many efforts have been devoted to understanding the interannual variability of AAM precipitation during boreal summer (e.g., Webster and Yang 1992; Lau and Nath 2000; Meehl and Arblaster 2002; Wu and Wang 2002; Saji and Yamagata 2003; Wang et al. 2003; Zhai et al. 2005; Cai et al. 2009; Xie et al. 2009; T. Zhou et al. 2009; Guo et al. 2014, 2016, 2017, 2018). Although there are also many previous studies that analyzed the boreal winter precipitation variations in the EA–AUS region (e.g., Zhang et al. 1996; Chen et al. 2000; Wu et al. 2003; England et al. 2006; Wu and Kirtman 2007; Wang et al. 2009; Ummenhofer et al. 2008; Taschetto et al. 2011; Gong et al. 2014, 2019), most of these studies focused on the individual regions of the AAM region. In view of the integrity of the AAM, a regional approach is unable to identify coherent interannual variability and overriding large-scale control (e.g., Yim et al. 2014; Wang et al. 2018). Therefore, the interannual variability of all AAM precipitation during boreal winter needs to be further investigated to reveal its major patterns and the associated influential factors.

Land precipitation is the most relevant factor for addressing societal concerns and needs. Nevertheless, it remains unclear how the precipitation varies over land in the AAM monsoon region during boreal winter. Since the oceanic monsoon precipitation far exceeds that over land, the leading modes of precipitation over both land and ocean cannot represent the real dominant precipitation variability over land alone (Wang et al. 2018). Therefore, in this study, we focus on the major modes of interannual variations in East Asian–Australian land precipitation (EAALP) during boreal winter. The main purposes of the present study are 1) to identify the interannual major modes of EAALP during boreal winter and 2) to unravel which factors are responsible for these major modes. The revelation of the factors and associated physical processes helps us better understand the mechanisms of climate variability in the EA–AUS region during boreal winter.

2. Data and methods

a. Data

Given that the quality of the precipitation data was much improved after 1979 due to the merging of satellite observations (Robertson et al. 2003), the present analysis is limited to the time period from 1979 to 2016 unless otherwise noted. The observational monthly global land precipitation data are obtained from the Climatic Research Unit (CRU) of the University of East Anglia (CRU_TS version 4.01) with a horizontal resolution of $0.5^\circ \times 0.5^\circ$ covering the time period from 1901 to 2016 (Harris and Jones 2017). Also, the observational land precipitation dataset from Global Precipitation Climatology Centre (GPCC) with $1^\circ \times 1^\circ$ horizontal resolution from 1948 to the present (Schneider et al. 2017) is used in attempt to verify the robustness of the results from CRU. For the convenience of comparison, the CRU datasets are bilinearly interpolated to the same resolution of $1^\circ \times 1^\circ$ in accordance with GPCC. Meanwhile, the large-scale reanalysis precipitation data over oceans are employed from Global Precipitation Climatology Project (GPCP) with a resolution of $2.5^\circ \times 2.5^\circ$ from 1979 to the present (Huffman et al. 2009). The reanalysis atmospheric variables are provided by the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) dataset, which has a horizontal resolution of $2.5^\circ \times 2.5^\circ$ from 1948 to the present (Kalnay et al. 1996). The sea surface temperature (SST) data are obtained from the monthly mean Extended Reconstructed SST, version 5 (ERSST v5), dataset (Huang et al. 2017). Based on the ERSST data, the Niño-3.4 index defined as the SST anomaly averaged over the equatorial east-central Pacific region (5°S – 5°N , 170° – 120°W) during boreal winter is adopted to describe the ENSO variability. In addition, the Indian Ocean subtropical dipole index (IOSD) is employed, which is defined as the difference of SST anomalies between the southwestern (20° – 35°S , 70° – 110°E) and northeastern (10° – 20°S , 90° – 120°E) areas in the southern Indian Ocean (SIO) (Gong et al. 2019).

b. Atmospheric general circulation model

To examine the impact of regional SST anomalies on the interannual variability of land monsoon precipitation in the EA–AUS region, the atmospheric general circulation model of ECHAM, version 5 (ECHAM5), is employed to conduct sensitivity experiments. ECHAM5 was developed by the Max Planck Institute for Meteorology (Roeckner et al. 2006), which evolved originally from the spectral weather prediction model of the European Centre for Medium-Range Weather Forecasts. The model was run at a horizontal resolution of spectral

triangular 63 (T63) and 19 vertical levels in a sigma-pressure coordinate system. ECHAM5 has been sufficiently proved to have excellent performance in the simulation of ENSO-related climate anomalies (e.g., Song and Zhou 2014; B. Liu et al. 2018).

c. Methods

Here our analysis focuses on boreal winter, and the winter means are constructed by averaging monthly data of December, January, and February (DJF). Our convention is that the winter of 1979 refers to the 1978/79 winter. Additionally, the long-term linear trends are removed from the original datasets prior to further analyses. The empirical orthogonal function (EOF) method (e.g., Lorenz 1956; von Storch and Zwiers 1999) is applied to identify the dominant covariation modes of land precipitation variations over the EA–AUS region. The principal component (PC) represents the time evolution of the corresponding EOF mode. Our main technique of analysis is linear regression and correlation for diagnosing the relevant signals associated with an index. We use partial regression and correlation techniques to isolate one index from another. Partial regression (correlation) involves computation of the linear regression (correlation) of a predictand upon a predictor, after the linear relationship with a second predictor has been removed from both the predictand and predictor. For example, the ENSO-independent PC indices are obtained through the following equation:

$$PCi_{RM_NINO3.4} = PCi - r_i \times NINO3.4, \quad (1)$$

where i denotes the order of the EOF mode, and r_i is the regression coefficient of the PCi with respect to the Niño-3.4 index. We use $PCi_{RM_NINO3.4}$ to indicate the ENSO-independent PC index. $PCi_{RM_NINO3.4}$ is then used to calculate the partial regression (without ENSO's influences) with the climate anomaly fields, in which the ENSO variability is removed. Similarly, we use $IOSD_{RM_Nino3.4}$ to indicate the residual IOSD without the influence of Niño-3.4. It is noted that this method has been widely used in previous similar studies (e.g., Nicholls 1989; Cai and van Rensch 2013; Gong et al. 2017; L. Liu et al. 2018). Furthermore, a two-tailed t test is used to evaluate the statistical significance of the correlation coefficients and the regression coefficients.

3. Dominant covariation modes of the EAALP during boreal winter

a. Spatial distribution of EAALP in climatology

Figure 1 presents the distributions of climatological EAALP during boreal winter. There is abundant

precipitation in the Maritime Continent and northern part of AUS above 500 mm per season with the maximum values exceeding 1300 mm per season in both the CRU and GPCC datasets (Figs. 1a,b). The large precipitation over the Maritime Continent and AUS during boreal winter is related to strong convective activities triggered by the interaction between the two hemispheres through the intrusion of northerly wind from the East Asian winter monsoon (e.g., Chen et al. 2000; Wang et al. 2009; Wang and Chen 2010). Meanwhile, climatological precipitation exceeds 180 mm per season over most of southern China where the maximum values exceed 250 mm per season in both the CRU and GPCC datasets (Figs. 1a,b). The winter precipitation over southern China accounts for more than 10% of annual total precipitation amount (e.g., Wu et al. 2003; Wang and Feng 2011; Ge et al. 2016).

The spatial distribution of the standard deviation of interannual precipitation variations over the EA–AUS region shows a pattern similar to that of climatological mean, albeit with relatively large amplitudes in the Maritime Continent, the northern part of AUS, and southern China (Figs. 1c,d). This large precipitation variability implies a potential for frequent occurrence of extreme precipitation in these regions on the interannual time scale.

b. Spatial and temporal features of the dominant modes of EAALP during boreal winter

We perform an EOF analysis in the region of 40°S–40°N, 90°–155°E to reveal the dominant covariation modes of interannual variability of the EAALP for the period 1979–2016 during boreal winter. The EOF analysis of the EAALP anomalies has been carried out by constructing an area-weighted covariance matrix to account for the decrease of area toward the pole. The first two leading modes are analyzed, which together account for about 40% of the total variances of EAALP in both the CRU and GPCC datasets. These two leading modes are well separated from the other modes according to the criteria proposed by North et al. (1982).

The first EOF mode (EOF1) accounts for 26.2% and 24.2% in the CRU and GPCC datasets, respectively. The spatial pattern of the EOF1 mode is characterized by an opposite loading between EA and AUS in both the CRU and GPCC datasets (Figs. 2a,b). During the positive phases of the EOF1 mode, positive precipitation anomalies are mainly located over southern China, whereas there are negative precipitation anomalies mainly located over AUS and the Maritime Continent, especially in the northern part of AUS (Figs. 2a,b). Meanwhile, the pattern correlation coefficient of the EOF1 mode and the temporal correlation coefficient of PC1 (Fig. 2c) between CRU and GPCC are 0.98 and 0.96, respectively. This

FIG. 1. (a) The climatological mean and (c) the interannual standard deviation of EAALP during boreal winter based on the CRU dataset from 1979 to 2016. (b),(d) As in (a) and (c), respectively, but for the GPCC dataset.

suggests that the EOF1 pattern is highly consistent between the two datasets.

The second EOF mode (EOF2) explains 15.4% and 13% of total variance of EAALP in the CRU and GPCC

datasets, respectively. The spatial feature of EOF2 in the CRU dataset also demonstrates a positive loading of precipitation anomalies in the southern part of EA, but with a relatively smaller amplitude and a slightly

FIG. 2. (a) The spatial pattern of EOF1 in EAALP during boreal winter based on the CRU dataset. (b) As in (a), but for the GPCC dataset. (c) The normalized principal component (PC) of EOF1 in the CRU and GPCC datasets. (d)–(f) As in (a)–(c), respectively, but for EOF2. The percentages of the explained variance of EOF1 and EOF2 are given at the top right of (a) and (b) and (d) and (e), respectively.

northward extension compared to that in EOF1 mode (Fig. 2d). These characteristics are reproduced in the GPCC dataset (Fig. 2e). Meanwhile, the precipitation anomalies are basically consistent between the EOF1 and EOF2 modes over most of the Maritime Continent in both the CRU and GPCC datasets. The pattern correlation coefficient of EOF2 and the temporal correlation coefficient of PC2 (Fig. 2f) between CRU and GPCC are 0.96 and 0.93, respectively. These results suggest that the EOF1 and EOF2 patterns are robust between the two datasets.

4. Role of ENSO and IOSD in the dominant patterns of EAALP variability

To understand the mechanisms behind the formation of the dominant patterns of EAALP during boreal winter, we examine the associated atmospheric circulation and SST anomalies over the Indo–Pacific Ocean in an attempt to elucidate the potential individual and combined influences of ENSO and IOSD on precipitation

variability over the EA–AUS region. Since both EOF1 and EOF2 between the CRU and GPCC datasets presented in section 3 are highly consistent, only the results based on the CRU data are shown in the following sections.

a. Role of ENSO in EAALP variability during boreal winter

The large-scale precipitation anomalies over the Indo–Pacific Ocean are obtained based on the GPCP dataset. As shown in Fig. 3, the SST anomaly pattern associated with EOF1 is quite similar to that associated with EOF2 over the tropical Pacific, with significant warming in the tropical central–eastern Pacific and cooling in the western North Pacific (WNP). This SST anomaly pattern bears a resemblance to that of ENSO that is a dominant mode of interannual climate variability. The correlation coefficients of PC1 and PC2 with Niño-3.4 index reach 0.49 and 0.56, respectively, during 1979–2016, both exceeding the 99% confidence level. The negative SST anomalies related to the EOF2 are

FIG. 3. (a) Regression maps of SST (color shading) and 850-hPa winds (vectors; m s^{-1}) against the normalized PC1 in the CRU dataset for 1979–2016. (c) As in (a), but for precipitation (color shading), 200-hPa velocity potential (contours; $0.2 \times 10^6 \text{ m}^2 \text{ s}^{-1}$), and divergent winds (vectors; m s^{-1}). (b), (d) As in (a) and (c), respectively, but for PC2. Stippled areas in (a) and (b) and in (c) and (d) indicate the statistical significance of SST and precipitation at the 95% confidence level, respectively. Regressed winds with speed less than 0.2 m s^{-1} are not shown.

somewhat stronger than those related to EOF1 in the WNP (Figs. 3a,b). In contrast, the SST anomaly pattern in the Indian Ocean associated with the EOF1 and EOF2 is obviously different, especially over the SIO (Figs. 3a,b). Corresponding to EOF1, significant warming is observed in most parts of the tropical Indian Ocean, but cooling occurs in the southwestern Indian Ocean (Fig. 3a). In term of EOF2, however, there are significant cold SST anomalies extending northwestward from the western coast of AUS to the equatorial Indian Ocean and warm SST anomalies extending southeastward from the east of Madagascar to the southeastern Indian Ocean (Fig. 3b).

Corresponding to the SST anomaly pattern in the Indo-Pacific Ocean, pronounced anticyclonic anomalies are observed west of the cold SST anomalies in the WNP related to both EOF1 and EOF2 due to a Rossby wave response to the cold SST anomalies (e.g., Zhang et al. 1996; Wang et al. 2000). The anomalous southwesterlies over EA not only weaken the East Asian winter monsoon (EAWM), but also transport warm and moist air from the low latitudes to EA, thereby inducing an increase in precipitation therein (Figs. 3a,b). The center of the anomalous anticyclone related to EOF2 shifts

slightly northward compared to that associated with the EOF1, which induces the northward extension of precipitation anomalies (Figs. 3c,d). The cold SST and accompanied anticyclonic anomalies in the WNP are generally regarded as the results of ENSO SST forcing. In other words, ENSO may play a dominant role in the formation of precipitation anomaly pattern in the EA region for both EOF1 and EOF2.

AUS is controlled by a strong anomalous anticyclone over the eastern Indian Ocean, which spreads farther eastward, reaching all the way to the entire AUS region in the lower troposphere. Meanwhile, an anomalous convergence exists in the upper troposphere of AUS associated with EOF1 (Fig. 3c). The resultant anomalous descending motion suppresses precipitation in most part of AUS (Figs. 3a,c). On the other hand, the southerly anomalies over AUS bring cold and dry air from high latitudes to AUS, inducing a dryer than normal climate in AUS (Figs. 3a,c). These characteristics are quite similar to those in the negative phases of the Australian summer monsoon (ASM; Kajikawa et al. 2010). The correlation coefficient between PC1 and ASM index defined by Kajikawa et al. (2010) is -0.76 , exceeding the 99.9% confidence level. This result

FIG. 4. As in Figs. 3a and 3c, but based on the normalized Niño-3.4 index.

suggests that the EOF1 can also largely reflect the ASM variability.

Corresponding to EOF2, there are significant cyclonic anomalies over the northern part of AUS in the lower troposphere, accompanied by anomalous divergence in the upper troposphere, which leads to anomalous ascending motion and increased precipitation over the northeast part of AUS (Figs. 3b,d). Interestingly, the anomalous anticyclone and divergence also shift westward to the Indian Ocean, inducing negative precipitation anomalies over the eastern tropical Indian Ocean and western AUS compared with those related to EOF1 (Fig. 3d). These features are quite different from those in ASM. The correlation coefficient between PC2 and ASM index is -0.06 , suggesting the independence between EOF2 and the ASM. Note that the SST anomaly pattern related to EOF1 and EOF2 is quite different in the SIO (Figs. 3a,b). These results imply that the SIO SST may also play a role in the formation of

different EAALP anomaly patterns during boreal winter. Moreover, the divergent wind anomalies over the Maritime Continent are quite consistent, corresponding to EOF1 and EOF2, which results in a similar precipitation anomaly pattern over the Maritime Continent.

In view of high correlations of both PC1 and PC2 with the Niño-3.4 index, ENSO may be a primary factor contributing to the EAALP variability during boreal winter. This is in good agreement with previous findings that precipitation anomalies over the southern part of EA and western AUS are both influenced by ENSO to a great extent (e.g., Ropelewski and Halpert 1987; Wu et al. 2003; Wu and Kirtman 2007; Zhou and Wu 2010). To illustrate the effect of ENSO on the precipitation anomaly in the EA–AUS region in detail, the SST, 850-hPa winds, precipitation, and 200-hPa velocity potential and corresponding divergent winds anomalies related to ENSO are shown in Fig. 4. During the warm

FIG. 5. As in Fig. 3, but based on the ENSO-independent PC time series.

phases of ENSO, the anomalous southerlies on the west side of the anticyclone associated with El Niño over the WNP bring warm and moist air from the low latitudes to EA and induce the increase in precipitation (Fig. 4a). Coincidentally, a large anomalous anticyclone, extending from the central Indian Ocean to the western AUS generated by the weakened Walker circulation, causes southerly anomalies in western AUS (Fig. 4). This tends to transport the dry and cold air from high latitudes to western AUS, resulting in reduced precipitation therein. Thus, the two dominant patterns of EAALP variability are both closely tied to ENSO's impact.

b. Role of IOSD in EAALP variability during boreal winter

In addition to ENSO, there may be other factors responsible for the distinct patterns revealed in EOF1 and EOF2, especially in AUS. To figure out the other possible climate factors of the dominant variability of EAALP during boreal winter, we obtain the anomaly patterns associated with ENSO-independent PC1 and PC2 indices by removing the ENSO variability from the original PC time series.

Figures 5a and 5b display the anomalies of SST and 850-hPa winds associated with the ENSO-independent PC1 and PC2 time series, respectively, which are obtained

by regressing the ENSO-independent SST and 850-hPa wind anomalies on the corresponding $PC1_{RM_NINO3.4}$ and $PC2_{RM_NINO3.4}$ indices, respectively. The large-scale anomalous warming in the tropical eastern-central Pacific has disappeared with respect to the $PC1_{RM_NINO3.4}$ and $PC2_{RM_NINO3.4}$ indices (Figs. 5a,b). The cold SST anomalies in the WNP are also weakened, especially with respect to $PC1_{RM_NINO3.4}$. The resultant anomalous anticyclone over the WNP is weakened as well (Figs. 5a,b), accompanied by a decrease in precipitation anomalies over EA (Figs. 5c,d). This confirms the dominant role of ENSO in modulating the precipitation anomaly over EA.

Meanwhile, the ENSO-independent circulation and precipitation anomalies associated with $PC1_{RM_NINO3.4}$ and $PC2_{RM_NINO3.4}$ over AUS remain similar to those related to original PC1 and PC2 (Figs. 3 and 5) to a certain extent. The anticyclonic anomalies associated with $PC1_{RM_NINO3.4}$ are somewhat weaker than those related to PC1 and mainly located on AUS (Fig. 5a). The cyclonic anomalies associated with $PC2_{RM_NINO3.4}$ are somewhat stronger than those related to PC2 (Figs. 3b and 5b). The resultant precipitation anomalies over AUS are almost oppositely associated with $PC1_{RM_NINO3.4}$ and $PC2_{RM_NINO3.4}$ indices. This result implies that apart from ENSO, there should be other factors influencing the precipitation variability in the AAM region during boreal winter.

FIG. 6. (a) Regressed vertical circulation consisting of vertical velocity (0.01 Pa s^{-1}) and zonal wind (m s^{-1}) against $\text{PC1}_{\text{RM_NINO3.4}}$ averaged over 10° – 20°S at different pressure levels. (b) As in (a), but for $\text{PC2}_{\text{RM_NINO3.4}}$. Shaded regions pass the 95% confidence level.

The ENSO-independent SST anomalies in the subtropical Indian Ocean associated with both $\text{PC1}_{\text{RM_NINO3.4}}$ and $\text{PC2}_{\text{RM_NINO3.4}}$ indices manifest a pronounced dipole structure, but with almost opposite polarity (Figs. 5a,b). Also, the resultant precipitation over the SIO displays opposite signs to a great extent, with significant positive and negative precipitation anomalies corresponding to $\text{PC1}_{\text{RM_NINO3.4}}$ and $\text{PC2}_{\text{RM_NINO3.4}}$ indices, respectively (Figs. 5c,d). The positive precipitation anomalies over the SIO accompanied by anomalous ascending motion trigger an anomalous zonal vertical circulation and in turn induce an anomalous descending motion over the northern part of AUS, suppressing precipitation in this region (Fig. 6a). In contrast, the negative precipitation anomalies over the SIO accompanied by anomalous descending motion generate a reversed anomalous zonal vertical circulation and in turn induce an anomalous ascending motion over the northern part of AUS, enhancing the precipitation in this region (Fig. 6b). This result suggests that the dipole structure of SST anomaly pattern in the SIO may contribute to the formation of precipitation anomaly pattern in the AAM regions, especially in AUS.

The IOSD has been well recognized to be a key climate mode over the SIO in modulating the precipitation variability through teleconnections (e.g., Reason 2001, 2002; Yang 2009; Kataoka et al. 2012; Cao et al. 2014; Gong et al. 2019). It is generally triggered by latent heat flux anomalies associated with the variations in the Mascarene high (Behera and Yamagata 2001) and maintained by the positive feedback of air–sea interaction over the subtropical SIO (e.g., Wang et al. 2003; Suzuki et al. 2004). As shown in Fig. 5, the opposite

sign of the IOSD-like SST anomaly pattern associated with $\text{PC1}_{\text{RM_NINO3.4}}$ and $\text{PC2}_{\text{RM_NINO3.4}}$ is distinctly observed, indicating that the IOSD could be considered as another factor influencing the EAALP variations. Following Gong et al. (2019), the IOSD index is defined as the difference of SST anomalies between the southwestern area (20° – 35°S , 70° – 110°E) and the northeastern area (10° – 20°S , 90° – 120°E) over the SIO to quantitatively verify the impact of IOSD on EAALP. The correlation coefficients of $\text{PC1}_{\text{RM_NINO3.4}}$ and $\text{PC2}_{\text{RM_NINO3.4}}$ with the ENSO-independent IOSD index reach -0.53 and 0.46 , respectively, during 1979–2016, both of which exceed the 99% confidence level. This implies that the IOSD may play a certain role in the formation of the dominant patterns of EAALP during boreal winter.

Figure 7a shows the anomalies of SST and 850-hPa winds associated with the IOSD index. A pronounced dipole SST anomaly pattern is observed over the SIO, with significant positive SST anomalies over the southwestern SIO and significant negative SST anomalies over the northeastern SIO (Fig. 7a). Accompanying the dipole SST anomaly pattern, there are obvious anticyclonic anomalies over the SIO, consistent with those in Gong et al. (2019). In the presence of climatological southeasterly winds over the subtropical SIO (not shown), the increased (decreased) total wind speed to the northeast (southwest) of the anomalous anticyclone center induces strengthening (weakening) of evaporative cooling. Meanwhile, the increased southeasterly off the northwest coast of AUS also strengthens the coastal upwelling and induces the further cooling of SST northwest of AUS. As such, the anticyclonic wind

FIG. 7. (a) Regression maps of SST (color shading) and 850-hPa winds (vectors; m s^{-1}) against the IOSD index. (b) As in (a), but for the $\text{IOSD}_{\text{RM_NINO3.4}}$ index. (c) As in (b), but for the precipitation (color shading), 200-hPa velocity potential (contours; $0.2 \times 10^6 \text{ m}^2 \text{ s}^{-1}$), and divergent winds (vectors; m s^{-1}). (d) Anomalies of the vertical circulation consisting of vertical velocity (0.01 Pa s^{-1}) and zonal wind (m s^{-1}) related to $\text{IOSD}_{\text{RM_NINO3.4}}$ averaged over $10^\circ\text{--}20^\circ\text{S}$ at different pressure levels. Stippled areas in (a)–(c) and shading in (d) indicate the statistical significance of SST, precipitation, and vertical circulation anomalies at the 95% confidence level, respectively. Regressed winds with speeds less than 0.2 m s^{-1} are not shown in (a)–(c).

anomalies trigger the SST dipole with cold SST anomalies to the east and warm SST anomalies to the west of the anticyclone. The cooling to the east of the anticyclone, in turn, suppresses convection and reduces latent heat release, which excites the atmospheric Rossby waves that propagate westward and reinforce the anomalous anticyclone.

This positive thermodynamic feedback between the anomalous anticyclone and the underlying ocean most likely explains the pronounced IOSD pattern, which is similar to the mechanism underlying the maintenance of the anomalous anticyclone associated with ENSO over the tropical WNP and tropical SIO proposed by Wang et al. (2003). In addition, there are weak cold SST

anomalies in the WNP and warm SST anomalies in the central Pacific. However, the correlation coefficient is only 0.18 between the IOSD and Niño-3.4 indices. The insignificant correlation between IOSD and ENSO suggests that the IOSD is independent of ENSO events to a great extent, which in general agrees with previous findings (e.g., Reason 2001; Yang 2009; Kataoka et al. 2012).

To interpret the independent effect of IOSD on the EAALP more clearly, the tropical SST variability is linearly removed from the IOSD index using a linear regression method; that is, we remove the Niño-3.4 variability from the original IOSD index and the residual IOSD can be represented as an ENSO-independent

IOSD_{RM_NINO3.4}. The results are basically similar to those with the original IOSD index with a pronounced dipole structure of SST anomaly pattern, accompanied by an anomalous anticyclone and cyclone in the SIO and northern AUS, respectively (Fig. 7b). The suppressed convection due to the anticyclonic anomalies over the SIO generates an anomalous zonal vertical circulation and induces anomalous ascending motion over the northeast part of AUS and increases the precipitation in this region (Figs. 7c,d). In addition, there are significant negative precipitation anomalies confined to the southeastern part of China during the positive phases of the IOSD events (Fig. 7c). This probably has something to do with the propagation of the low-frequency wave train from the SIO to the Northern Hemisphere (Yang 2009). Thus, the findings further give us more confidence that the IOSD cannot be ignored in accounting for the interannual covariations of EAALP.

c. Combined effect of ENSO and IOSD in EOF1 and EOF2

The aforementioned results indicate that both ENSO and IOSD can significantly influence the precipitation variability over the EA–AUS. During the positive phases of ENSO events, there are significant positive precipitation anomalies over the southern EA and significant negative precipitation mainly in the western and northern part of AUS. In contrast, during the positive phases of IOSD, pronounced positive precipitation anomalies are observed in the northern AUS and significant negative precipitation anomalies are located in southeastern China. Comparing these results with the EOF1 and EOF2 patterns, this implies that the combination based on the different phases of the ENSO and IOSD events may form the interannual dominant patterns of EAALP variability during boreal winter. To verify this hypothesis, all the 38 winters during 1979–2016 are divided into four combinations based on the magnitudes of Niño-3.4 and IOSD indices. The out-of-phase composites are constructed by the differences of the positive phases of Niño-3.4 index (Niño-3.4 > 0) coupled with negative phases of IOSD index (IOSD < 0) minus the negative phases of Niño-3.4 index (Niño-3.4 < 0) coupled with positive phases of IOSD index (IOSD > 0). Similarly, the in-phase composites are constructed by the differences of the positive phases of Niño-3.4 index (Niño-3.4 > 0) coupled with positive phases of IOSD index (IOSD > 0) minus the negative phases of Niño-3.4 index (Niño-3.4 < 0) coupled with negative phases of IOSD index (IOSD < 0). It should be noted that the results are basically similar if the criterion of 0.3 standard deviations is used (not shown). The composite

anomalies based on the out-of-phase and in-phase status between the Niño-3.4 and IOSD indices are presented in Fig. 8. The time series of Niño-3.4 and IOSD indices are presented in Fig. 8a. The out-of-phase composite differences of SST and 850-hPa wind anomalies shown in Fig. 8b between the Niño-3.4 and IOSD indices indicate that the SST anomaly pattern is quite similar to that corresponding to EOF1, with pronounced warming in the tropical central–eastern Pacific, cooling in the WNP, and a negative IOSD-like SST anomaly pattern in the SIO (Fig. 8b). Meanwhile, the atmospheric responses to the SST anomaly pattern are also similar to those related to EOF1, with an anomalous anticyclone over the WNP and AUS, which induces significant positive and negative precipitation anomalies over southern China and AUS, respectively (Figs. 8b,d).

Furthermore, the in-phase composite differences of climate anomalies between the Niño-3.4 and IOSD indices are displayed in Figs. 8c and 8e. The SST and circulation anomaly pattern is similar to that corresponding to EOF2, with large-scale warming in the tropical central–eastern Pacific, cooling in the WNP, and a positive IOSD-like SST anomaly pattern in the SIO. Moreover, the anticyclonic and cyclonic anomalies over the WNP and northeastern AUS are also reproduced, inducing positive precipitation anomalies over southern China and northeastern AUS (Figs. 8c,e). The feature of relatively weaker positive precipitation anomalies in southern China is also similar to that related to EOF2, which could be attributed to the damping effect of IOSD on the precipitation anomalies over southeastern China during its positive phases. Meanwhile, the southerly anomalies west of the anomalous anticyclone in the SIO bring dry and cold air from high latitudes to western AUS, inducing the decrease of precipitation in this region. These results confirm the hypothesis that the interannual dominant patterns of EAALP during boreal winter are formed by different combinations with opposite and same phases of ENSO and IOSD events to a large extent. This result also suggests that IOSD actually plays an equivalent role to ENSO in the variability of the EAALP during the boreal winter.

5. Discussion

We have shown that the EOF1 and EOF2 patterns of interannual variations of EAALP during boreal winter are mainly formed by the combination of different phases between ENSO and IOSD. Meanwhile, when the ENSO variability is removed from the PC2, cold SST anomalies still exist in the WNP associated with the PC2_{RM_NINO3.4} index, albeit with a relatively weaker

FIG. 8. (a) The time series of normalized Niño-3.4 (red line) and IOSD (blue line) indices during the period 1979–2016. (b) Composite differences of boreal winter SST (color shading) and 850-hPa winds (vectors; m s^{-1}) between the positive phases of Niño-3.4 index with negative phases of IOSD index and the negative phases of Niño-3.4 index with positive phases of IOSD index. (d) As in (b), but for the precipitation (color shading), 200-hPa velocity potential (contours; $0.4 \times 10^6 \text{ m}^2 \text{ s}^{-1}$), and divergent winds (vectors; m s^{-1}). (c),(e) As in (b) and (d), but for the composite differences between the same warm and cold phases of the Niño-3.4 and IOSD indices. Stippled areas in (b) and (c) and in (d) and (e) indicate that the difference of SST and precipitation anomalies is statistically significant at the 95% confidence level, respectively. Regressed winds with speeds less than 0.3 m s^{-1} are not shown in (b)–(e).

amplitude. The WNP cold SST anomalies not only favor the maintenance of the anomalous anticyclone over the WNP, which induces the increase of precipitation in EA, but also may strengthen the cross-equatorial flows and induce the convergence and ascending motion over northern AUS. To illustrate the possible effect of the WNP cold SST anomalies on circulation and precipitation anomalies over northern AUS, the regressed

vertical circulation averaged over 120° – 150°E , consisting of vertical velocity and meridional winds, against $\text{PC2}_{\text{RM_NINO3.4}}$ is shown in Fig. 9. Indeed, there is an anomalous meridional vertical circulation between the WNP and northern AUS with descending and divergent wind anomalies over the WNP and convergent and ascending anomalies over northern AUS (Fig. 9). This result confirms the relationship of vertical circulation

FIG. 9. Anomalies of the vertical circulation consisting of vertical velocity (0.01 Pa s^{-1}) and meridional wind (m s^{-1}) related to $\text{PC2}_{\text{RM_NINO3.4}}$ averaged over 120° – 150°E at different pressure levels. Shaded regions pass the 95% confidence level.

between the WNP and northern AUS. The WNP cold SST anomalies may generate anomalous downward motion and below-normal precipitation over the WNP and facilitate anomalous upward motion and above-normal precipitation over northern AUS via an anomalous meridional vertical cell (Figs. 9 and 5b,d). Thus, the ENSO-independent cold SST anomalies in the WNP may exert a significant impact on the formation of EOF2 pattern of EAALP during boreal winter.

To confirm the impact of the SST anomalies in SIO and WNP on the EAALP during boreal winter, numerical simulation is employed with ECHAM5 model. Our analysis of the ECHAM5 simulation with observational SST specified confirms that the model captures well the spatial distribution of the standard deviation and the dominant modes of interannual variations of winter precipitation over the AAM region (see Figs. S1 and S2 in the supplemental material). We carry out three 40-yr numerical simulation experiments using the ECHAM5 model, named EXP_CON, EXP_IOSD-WNP, and EXP_WNP. The EXP_CON experiment is a control run with climatological SST forcing specified in the global ocean. The last 30 years of a 40-yr control run are used in the analysis. The multiple-year mean winter values are equivalent to those from a multi-ensemble of sensitivity experiments with the same SST forcing but starting from different initial conditions (e.g., Chen et al. 2014; B. Liu et al. 2018). In the EXP_IOSD-WNP experiment, IOSD- and $\text{PC2}_{\text{RM_NINO3.4}}$ -related SST anomalies over the SIO and WNP are added onto climatological SST. The EXP_WNP experiment is the same as EXP_IOSD-WNP experiment, but with only WNP anomalies added onto climatological SST. Figures 10a and 10c display the distribution of imposed

SST anomalies that are multiplied by a factor of 4 in the original SST anomalies for the EXP_IOSD-WNP and EXP_WNP experiments, respectively. For a fair comparison with observations, the model sensitivity results are rescaled by a factor of 1/4 (Sutton and Hodson 2007). Composite differences of precipitation and 850-hPa winds between the EXP_IOSD-WNP and the EXP_CON experiments are presented in Fig. 10b.

Corresponding to the forcing of dipole SST anomalies in the SIO and cold SST anomalies in the WNP, the atmospheric and precipitation responses are similar to those associated with $\text{PC2}_{\text{RM_NINO3.4}}$ (Figs. 10b and 5b,d). The positive precipitation anomalies over EA are well reproduced, which is attributed to the transport of warm and moist air by southerly anomalies west of the anomalous anticyclone. At the same time, the positive precipitation anomalies over northeastern AUS are also reproduced. The ascending motion associated with the cyclonic anomalies triggered by anomalous descending and divergent winds from the SIO and WNP induce the increase in precipitation over northeastern AUS. To further verify the effect of WNP SST anomalies, composite differences of precipitation and 850-hPa winds between the EXP_WNP and the EXP_CON experiments are also presented in Fig. 10d. There is an anomalous anticyclone over the WNP, which brings warm and moist air from low latitudes to EA and induces the increase in precipitation in EA. Meanwhile, there are obvious cross-equatorial flows from WNP to northern AUS, which induce a relatively weaker convergence and positive precipitation anomalies in northern AUS. This supports that the cross-equatorial flows triggered by the anomalous divergence in the WNP can partly contribute to the positive precipitation anomalies in northern AUS. Those features further support that the IOSD SST anomalies together with the WNP cold SST anomalies largely contribute to the formation of EOF2 pattern of EAALP during the boreal winter through anomalous vertical circulations.

6. Concluding remarks

The present study investigates the interannual variations of EAALP during boreal winter during 1979–2016 based on the latest version of CRU and GPCC datasets. The distribution of the interannual standard deviation of EAALP shows large values in the Maritime Continent, the northern part of AUS, and southern China. This implies a potential possibility of the occurrence of extreme precipitation over this region on an interannual time scale. Therefore, the interannual variations of EAALP during boreal winter are further examined.

FIG. 10. (a) The distribution of imposed SST anomalies that are multiplied by a factor of 4 in the original SST anomalies for the IOSD-WNP SST experiment. (b) Composite differences of precipitation (color shading) and 850-hPa winds (vectors; m s^{-1}) in the boreal winter between the IOSD-WNP SST experiment (EXP_IOSD-WNP) and the control run (EXP_CON), which are multiplied by a factor of 1/4 in the model results. (c),(d) As in (a) and (b), respectively, but for composite differences between the WNP SST experiment (EXP_WNP) and EXP_CON.

The empirical orthogonal function (EOF) method is used to extract the dominant patterns with coherent temporal variations of EAALP during boreal winter. The leading mode of EAALP variations is characterized by opposite-sign anomalies between the southern EA and most of AUS. The second mode features a similar pattern over EA to that in the first mode, but with a relatively weaker amplitude and a slightly northward extension. Meanwhile, a southwest–northeast dipole structure of land precipitation anomalies is found over AUS in the EOF2 pattern. Analysis shows that positive (negative) precipitation anomalies in southern China and negative (positive) precipitation anomalies in western AUS during the positive (negative) phases of EOF1 and EOF2 are largely forced by ENSO. During the warm phases of ENSO, the southerly anomalies west of the anomalous anticyclone over the WNP bring warm and moist air from low latitudes to EA and induce the increase in precipitation. Meanwhile, a large anomalous anticyclone, extending from the central Indian Ocean to

western AUS, which is generated by the weakened Walker circulation associated with warm ENSO events, induces southerly anomalies in western AUS and transports dry and cold air from high latitudes to western AUS, reducing the precipitation in this region.

Correlation/regression analysis shows that the ENSO-independent IOSD plays an important role in the formation of precipitation anomalies in northeastern AUS and southern China. The broadly opposite precipitation and atmospheric circulation anomalies are associated with the opposite SST anomaly dipole pattern in the SIO corresponding to ENSO-independent PC1 and PC2. The dominant dynamical mechanisms behind the pattern pertaining to the first two leading EOF modes are schematically summarized in Fig. 11. The anomalous ascending (descending) motion associated with the positive (negative) precipitation anomalies generated by the negative (positive) phases of IOSD over the SIO triggers an anomalous zonal vertical circulation and induces an anomalous descending (ascending) motion

FIG. 11. Schematic diagrams showing the dynamical mechanisms associated with the (left) first and (right) second modes of boreal winter land precipitation over the EA–AUS region. Red upward and blue downward shaded areas indicate positive and negative SST anomalies, respectively. Dashed red and blue arrows correspond to ascending and descending motions, respectively. The lower-level red and blue curved arrows denote the anomalous anticyclone circulation (AAC) and anomalous cyclone circulation (AC), respectively. Solid and dashed circles denote convergent and divergent wind anomalies in the upper levels, respectively.

over northern AUS, reducing (increasing) the precipitation in this region. Meanwhile, the cold SST anomalies in the WNP also play a role for the formation of EOF2. They not only favor the maintenance of the anomalous anticyclone over the WNP, but also strengthen the cross-equatorial flows in the lower troposphere and facilitate the positive precipitation in northern AUS. This mechanism of IOSD and WNP SST anomalies on the formation of EOF2 pattern is verified by numerical experiments using ECHAM5 with specified SST forcing. Further composite analysis suggests that the dominant patterns of EAALP variability are largely formed by the opposite and same combinations of the ENSO and IOSD events. These results indicate that in addition to ENSO, the IOSD should be considered as another crucial factor in the prediction of the EAALP variability on the interannual time scale during boreal winter.

Recently, Sekizawa et al. (2018) indicated that interannual variations in precipitation over northwestern AUS are not forced by tropical SST anomalies. Since their results are obtained from several station data confined in northwestern AUS, this indicates that northwestern AUS precipitation may be influenced by local factors or internal variability. Meanwhile, it should be noted that this study mainly focuses on the large-scale monsoon precipitation in the AAM regions and the

local station data cannot fully represent the large-scale monsoon variations. In fact, the large-scale AAM monsoon variations, whether defined based on the monsoonal winds or on monsoonal precipitation, are found to be tightly connected to tropical SST anomalies (not shown). Therefore, tropical SST anomalies can influence the large-scale monsoonal precipitations in the EA–AUS region. Recent studies have proposed a new SST anomaly pattern off the west coast of AUS, named Ningaloo Niño (e.g., Kataoka et al. 2014; Tozuka et al. 2014). The correlation coefficients of PC1 and PC2 with the Ningaloo Niño index during boreal winter are -0.1 and -0.39 , respectively, implying a possible effect of Ningaloo Niño on the formation of EOF2. However, the variability of Ningaloo Niño is found to be not fully independent of ENSO. Thus, the possible connection between the Ningaloo Niño and EAALP is worth further exploration in the future. Moreover, some previous studies found that the boreal winter land precipitation in EA and AUS both experienced the obvious interdecadal changes during past several decades (e.g., Ge et al. 2016; Arblaster et al. 2002). The possible role of ENSO and IOSD in the interdecadal changes of EAALP needs to be further investigated in the future.

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REFERENCES

- Arblaster, J. M., G. A. Meehl, and A. M. Moore, 2002: Interdecadal modulation of Australian rainfall. *Climate Dyn.*, **18**, 519–531, <https://doi.org/10.1007/s00382-001-0191-y>.
- Behera, S. K., and T. Yamagata, 2001: Subtropical SST dipole events in the southern Indian ocean. *Geophys. Res. Lett.*, **28**, 327–330, <https://doi.org/10.1029/2000GL014451>.
- Cai, W., and P. van Rensch, 2013: Austral summer teleconnections of Indo-Pacific variability: Their nonlinearity and impacts on Australian climate. *J. Climate*, **26**, 2796–2810, <https://doi.org/10.1175/JCLI-D-12-00458.1>.
- , T. Cowan, and A. Sullivan, 2009: Recent unprecedented skewness towards positive Indian Ocean dipole occurrences and its impact on Australian rainfall. *Geophys. Res. Lett.*, **36**, L11705, <https://doi.org/10.1029/2009GL037604>.
- Cao, J., P. Yao, L. Wang, and K. Liu, 2014: Summer rainfall variability in low-latitude highlands of China and subtropical Indian Ocean dipole. *J. Climate*, **27**, 880–892, <https://doi.org/10.1175/JCLI-D-13-00121.1>.
- Chen, J., Z. Wen, R. Wu, Z. Chen, and P. Zhao, 2014: Interdecadal changes in the relationship between southern China winter–spring precipitation and ENSO. *Climate Dyn.*, **43**, 1327–1338, <https://doi.org/10.1007/s00382-013-1947-x>.
- Chen, W., H. F. Graf, and R. Huang, 2000: The interannual variability of East Asian winter monsoon and its relation to the summer monsoon. *Adv. Atmos. Sci.*, **17**, 48–60, <https://doi.org/10.1007/s00376-000-0042-5>.
- England, M., C. Ummenhofer, and A. Santoso, 2006: Interannual rainfall extremes over southwest Western Australia linked to Indian Ocean climate variability. *J. Climate*, **19**, 1948–1969, <https://doi.org/10.1175/JCLI3700.1>.
- Ge, J., X. Jia, and H. Lin, 2016: The interdecadal change of the leading mode of the winter precipitation over China. *Climate Dyn.*, **47**, 2397–2411, <https://doi.org/10.1007/s00382-015-2970-x>.
- Gong, H., L. Wang, W. Chen, R. Wu, K. Wei, and X. Cui, 2014: The climatology and interannual variability of the East Asian winter monsoon in CMIP5 models. *J. Climate*, **27**, 1659–1678, <https://doi.org/10.1175/JCLI-D-13-00039.1>.
- , —, —, D. Nath, G. Huang, and W. Tao, 2015: Diverse influences of ENSO on the East Asian–western Pacific winter climate tied to different ENSO properties in CMIP5 models. *J. Climate*, **28**, 2187–2202, <https://doi.org/10.1175/JCLI-D-14-00405.1>.
- , —, —, X. Chen, and D. Nath, 2017: Biases of the wintertime Arctic Oscillation in CMIP5 models. *Environ. Res. Lett.*, **12**, 014001, <https://doi.org/10.1088/1748-9326/12/1/014001>.
- , W. Zhou, W. Chen, L. Wang, Y.-T. Leung, K.-Y. Cheung, and Y. Zhang, 2019: Modulation of the southern Indian Ocean dipole on the impact of El Niño–Southern Oscillation on Australian summer rainfall. *Int. J. Climatol.*, **39**, 2484–2490, <https://doi.org/10.1002/joc.5941>.
- Guo, J., M. Deng, J. Fan, Z. Li, Q. Chen, P. Zhai, Z. Dai, and X. Li, 2014: Precipitation and air pollution at mountain and plain stations in northern China: Insights gained from observations and modeling. *J. Geophys. Res. Atmos.*, **119**, 4793–4807, <https://doi.org/10.1002/2013JD021161>.
- , and Coauthors, 2016: Delaying precipitation and lightning by air pollution over the Pearl River Delta. Part I: Observational analyses. *J. Geophys. Res.*, **121**, 6472–6488, <https://doi.org/10.1002/2015JD023257>.
- , and Coauthors, 2017: Declining frequency of summertime local-scale precipitation over eastern China from 1970 to 2010 and its potential link to aerosols. *Geophys. Res. Lett.*, **44**, 5700–5708, <https://doi.org/10.1002/2017GL073533>.
- , and Coauthors, 2018: Aerosol-induced changes in the vertical structure of precipitation: A perspective of TRMM precipitation radar. *Atmos. Chem. Phys.*, **18**, 13 329–13 343, <https://doi.org/10.5194/acp-18-13329-2018>.
- Hamada, J. I., M. D. Yamanaka, J. Matsumoto, S. Fukao, P. A. Winarso, and T. Sribimawati, 2002: Spatial and temporal variations of the rainy season over Indonesia and their link to ENSO. *J. Meteor. Soc. Japan*, **80**, 285–310, <https://doi.org/10.2151/jmsj.80.285>.
- Harris, I., and P. Jones, 2017: CRU TS4.00: Climatic Research Unit (CRU) Time-Series (TS) version 4.00 of high resolution gridded data of month-by-month variation in climate (Jan. 1901–Dec. 2015). Centre for Environmental Data Analysis, University of East Anglia Climatic Research Unit, accessed 2018, <https://doi.org/10.5285/edf8febfaad48abb2cbaf7d7e846a86>.
- Huang, B., and Coauthors, 2017: NOAA Extended Reconstructed Sea Surface Temperature (ERSST), version 5. NOAA/NCEI, accessed 2018, <https://doi.org/10.7289/V5T72FNM>.
- Huffman, G. J., R. F. Adler, D. T. Bolvin, and G. J. Gu, 2009: Improving the global precipitation record: GPCP version 2.1. *Geophys. Res. Lett.*, **36**, L17808, <https://doi.org/10.1029/2009GL040000>.
- Kajikawa, Y., B. Wang, and J. Yang, 2010: A multi-time scale Australian monsoon index. *Int. J. Climatol.*, **30**, 1114–1120, <https://doi.org/10.1002/joc.1955>.
- Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471, [https://doi.org/10.1175/1520-0477\(1996\)077<0437:TNYRP>2.0.CO;2](https://doi.org/10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2).
- Kataoka, T., T. Tozuka, Y. Masumoto, and T. Yamagata, 2012: The Indian Ocean subtropical dipole mode simulated in the CMIP3 models. *Climate Dyn.*, **39**, 1385–1399, <https://doi.org/10.1007/s00382-011-1271-2>.
- , —, S. Behera, and T. Yamagata, 2014: On the Ningaloo Niño/Niña. *Climate Dyn.*, **43**, 1463–1482, <https://doi.org/10.1007/s00382-013-1961-z>.
- Lau, N.-C., and M. J. Nath, 2000: Impact of ENSO on the variability of the Asian–Australian monsoons as simulated in GCM experiments. *J. Climate*, **13**, 4287–4309, [https://doi.org/10.1175/1520-0442\(2000\)013<4287:IOEOTV>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<4287:IOEOTV>2.0.CO;2).
- Liu, B., G. Huang, K. Hu, R. Wu, H. Gong, P. Wang, and G. Zhao, 2018: The multidecadal variations of the interannual relationship between the East Asian summer monsoon and ENSO in a coupled model. *Climate Dyn.*, **51**, 1671–1686, <https://doi.org/10.1007/s00382-017-3976-3>.
- Liu, L., J. Guo, W. Chen, R. Wu, L. Wang, H. Gong, W. Xue, and J. Li, 2018: Large-scale pattern of the diurnal temperature range changes over East Asia and Australia in boreal winter: A perspective of atmospheric circulation. *J. Climate*, **31**, 2715–2728, <https://doi.org/10.1175/JCLI-D-17-0608.1>.

- Lorenz, E. N., 1956: Empirical orthogonal functions and statistical weather prediction. Statistical Forecasting Project Rep. 1, MIT Department of Meteorology, 49 pp.
- Meehl, G. A., and J. Arblaster, 2002: The troposphere biennial oscillation and Asian–Australian monsoon rainfall. *J. Climate*, **15**, 722–744, [https://doi.org/10.1175/1520-0442\(2002\)015<0722:TTBOAA>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<0722:TTBOAA>2.0.CO;2).
- Nicholls, N., 1989: Sea surface temperatures and Australian winter rainfall. *J. Climate*, **2**, 965–973, [https://doi.org/10.1175/1520-0442\(1989\)002<0965:SSTAAW>2.0.CO;2](https://doi.org/10.1175/1520-0442(1989)002<0965:SSTAAW>2.0.CO;2).
- North, G. R., T. L. Bell, R. F. Cahalan, and F. J. Moeng, 1982: Sampling errors in the estimation of empirical orthogonal functions. *Mon. Wea. Rev.*, **110**, 699–706, [https://doi.org/10.1175/1520-0493\(1982\)110<0699:SEITEO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1982)110<0699:SEITEO>2.0.CO;2).
- Reason, C., 2001: Subtropical Indian Ocean SST dipole events and southern African rainfall. *Geophys. Res. Lett.*, **28**, 2225–2227, <https://doi.org/10.1029/2000GL012735>.
- , 2002: Sensitivity of the southern African circulation to dipole sea-surface temperature patterns in the South Indian Ocean. *Int. J. Climatol.*, **22**, 377–393, <https://doi.org/10.1002/joc.744>.
- Robertson, F. R., D. E. Fitzjarrald, and C. D. Kummerow, 2003: Effects of uncertainty in TRMM precipitation radar path integrated attenuation on interannual variations of tropical oceanic rainfall. *Geophys. Res. Lett.*, **30**, 1180, <https://doi.org/10.1029/2002GL016416>.
- Roeckner, E., and Coauthors, 2006: Sensitivity of simulated climate to horizontal and vertical resolution in the ECHAM5 atmosphere model. *J. Climate*, **19**, 3771–3791, <https://doi.org/10.1175/JCLI3824.1>.
- Ropelewski, C. F., and M. S. Halpert, 1987: Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation. *Mon. Wea. Rev.*, **115**, 1606–1626, [https://doi.org/10.1175/1520-0493\(1987\)115<1606:GARSPP>2.0.CO;2](https://doi.org/10.1175/1520-0493(1987)115<1606:GARSPP>2.0.CO;2).
- Saji, N. H., and T. Yamagata, 2003: Possible impacts of Indian Ocean dipole mode events on global climate. *Climate Res.*, **25**, 151–169, <https://doi.org/10.3354/cr025151>.
- Schneider, U., P. Finger, A. Meyer-Christoffer, E. Rustemeier, M. Ziese, and A. Becker, 2017: Evaluating the hydrological cycle over land using the newly-corrected precipitation climatology from the Global Precipitation Climatology Centre (GPCC). *Atmosphere*, **8**, 52, <https://doi.org/10.3390/atmos8030052>.
- Sekizawa, S., H. Nakamura, and Y. Kosaka, 2018: Interannual variability of the Australian summer monsoon system internally sustained through wind–evaporation feedback. *Geophys. Res. Lett.*, **45**, 7748–7755, <https://doi.org/10.1029/2018GL078536>.
- Song, F., and T. Zhou, 2014: Interannual variability of East Asian summer monsoon simulated by CMIP3 and CMIP5 AGCMs: Skill dependence on Indian Ocean–western Pacific anticyclone teleconnection. *J. Climate*, **27**, 1679–1697, <https://doi.org/10.1175/JCLI-D-13-00248.1>.
- Sutton, R., and D. Hodson, 2007: Climate response to basin-scale warming and cooling of the North Atlantic Ocean. *J. Climate*, **20**, 891–907, <https://doi.org/10.1175/JCLI4038.1>.
- Suzuki, R., S. K. Behera, S. Iizuka, and T. Yamagata, 2004: Indian Ocean subtropical dipole simulated using a coupled general circulation model. *J. Geophys. Res.*, **109**, C09001, <https://doi.org/10.1029/2003JC001974>.
- Taschetto, A. S., A. Sen Gupta, H. H. Hendon, C. C. Ummenhofer, and M. H. England, 2011: The contribution of Indian Ocean sea surface temperature anomalies on Australian summer rainfall during El Niño events. *J. Climate*, **24**, 3734–3747, <https://doi.org/10.1175/2011JCLI3885.1>.
- Tozuka, T., T. Kataota, and T. Yamagata, 2014: Locally and remotely forced atmospheric circulation anomalies of Ningaloo Niño/Niña. *Climate Dyn.*, **43**, 2197–2205, <https://doi.org/10.1007/s00382-013-2044-x>.
- Ummenhofer, C., A. Sen Gupta, M. J. Pook, and M. H. England, 2008: Anomalous rainfall over southwest Western Australia forced by Indian Ocean sea surface temperatures. *J. Climate*, **21**, 5113–5134, <https://doi.org/10.1175/2008JCLI2227.1>.
- von Storch, H., and F. W. Zwiers, 1999: *Statistical Analysis in Climate Research*. Cambridge University Press, 845 pp., <https://doi.org/10.1007/3-540-37722-0>.
- Wang, B., Ed., 2006: *The Asian Monsoon*. Springer Praxis, 779 pp.
- , R. Wu, and X. Fu, 2000: Pacific–East Asian teleconnection: How does ENSO affect East Asian climate? *J. Climate*, **13**, 1517–1536, [https://doi.org/10.1175/1520-0442\(2000\)013<1517:PEATHD>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<1517:PEATHD>2.0.CO;2).
- , —, and T. Li, 2003: Atmosphere–warm ocean interaction and its impact on Asian–Australian monsoon variation. *J. Climate*, **16**, 1195–1211, [https://doi.org/10.1175/1520-0442\(2003\)16<1195:AOIAH>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)16<1195:AOIAH>2.0.CO;2).
- , and Coauthors, 2018: Toward predicting changes in the land monsoon rainfall a decade in advance. *J. Climate*, **31**, 2699–2714, <https://doi.org/10.1175/JCLI-D-17-0521.1>.
- Wang, L., and W. Chen, 2010: How well do existing indices measure the strength of the East Asian winter monsoon? *Adv. Atmos. Sci.*, **27**, 855–870, <https://doi.org/10.1007/s00376-009-9094-3>.
- , and J. Feng, 2011: Two major modes of the wintertime precipitation over China (in Chinese). *Chin. J. Atmos. Sci.*, **35**, 1105–1116.
- , W. Chen, W. Zhou, and R. Huang, 2009: Interannual variations of East Asian trough axis at 500 hPa and its association with the East Asian winter monsoon pathway. *J. Climate*, **22**, 600–614, <https://doi.org/10.1175/2008JCLI2295.1>.
- Webster, P. J., and S. Yang, 1992: Monsoon and ENSO: Selectively interactive systems. *Quart. J. Roy. Meteor. Soc.*, **118**, 877–926, <https://doi.org/10.1002/qj.49711850705>.
- Wu, R., and B. Wang, 2000: Interannual variability of summer monsoon onset over the western North Pacific and the underlying processes. *J. Climate*, **13**, 2483–2501, [https://doi.org/10.1175/1520-0442\(2000\)013<2483:IVOSMO>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<2483:IVOSMO>2.0.CO;2).
- , and —, 2002: A contrast of the East Asian summer monsoon–ENSO relationship between 1962–77 and 1978–93. *J. Climate*, **15**, 3266–3279, [https://doi.org/10.1175/1520-0442\(2002\)015<3266:ACOTEA>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<3266:ACOTEA>2.0.CO;2).
- , and B. P. Kirtman, 2007: Roles of the Indian Ocean in the Australian summer monsoon–ENSO relationship. *J. Climate*, **20**, 4768–4788, <https://doi.org/10.1175/JCLI4281.1>.
- , Z.-Z. Hu, and B. P. Kirtman, 2003: Evolution of ENSO-related rainfall anomalies in East Asia. *J. Climate*, **16**, 3742–3758, [https://doi.org/10.1175/1520-0442\(2003\)016<3742:EOERA1>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<3742:EOERA1>2.0.CO;2).
- Xie, S.-P., 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>.
- Yang, Q., 2009: Impact of the Indian Ocean subtropical dipole on the precipitation of east China during winter monsoons. *J. Geophys. Res.*, **114**, D14110, <https://doi.org/10.1029/2008JD011173>.

- Yim, S.-Y., B. Wang, J. Liu, and Z. Wu, 2014: A comparison of regional monsoon variability using monsoon indices. *Climate Dyn.*, **43**, 1423–1437, <https://doi.org/10.1007/s00382-013-1956-9>.
- Zhai, P., X. Zhang, H. Wan, and X. Pan, 2005: Trends in total precipitation and frequency of daily precipitation extremes over China. *J. Climate*, **18**, 1096–1108, <https://doi.org/10.1175/JCLI-3318.1>.
- Zhang, C., and H. Zhang, 2010: Potential impacts of East Asian winter monsoon on climate variability and predictability in the Australian summer monsoon region. *Theor. Appl. Climatol.*, **101**, 161–177, <https://doi.org/10.1007/s00704-009-0246-2>.
- Zhang, R., A. Sumi, and M. Kimoto, 1996: Impact of El Niño on the East Asian monsoon: A diagnostic study of the '86/87 and '91/92 events. *J. Meteor. Soc. Japan*, **74**, 49–62, https://doi.org/10.2151/jmsj1965.74.1_49.
- Zhou, L.-T., and R. Wu, 2010: Respective impacts of the East Asian winter monsoon and ENSO on winter rainfall in China. *J. Geophys. Res.*, **115**, D02107, <https://doi.org/10.1029/2009JD012502>.
- Zhou, T., B. Wu, and B. Wang, 2009: How well do atmospheric general circulation models capture the interannual variability of the Asian-Australian monsoon? *J. Climate*, **22**, 1159–1173, <https://doi.org/10.1175/2008JCLI2245.1>.
- Zhou, W., J. C. L. Chan, W. Chen, J. Ling, J. G. Pinto, and Y. Shao, 2009: Synoptic-scale controls of persistent low temperature and icy weather over southern China in January 2008. *Mon. Wea. Rev.*, **137**, 3978–3991, <https://doi.org/10.1175/2009MWR2952.1>.