Large-Scale Pattern of the Diurnal Temperature Range Changes over East Asia and Australia in Boreal Winter: A Perspective of Atmospheric Circulation

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ABSTRACT

The present study applies the empirical orthogonal function (EOF) method to investigate the large-scale pattern and the plausible dynamic processes of the boreal winter diurnal temperature range (DTR) changes in the East Asia (EA)–Australia (AUS) region based on the CRU Time Series version 4.00 (TS4.00) and NCEP–NCAR reanalysis datasets. Results show that the DTR changes during 1948–2015 are dominated by two distinct modes. The first mode, characterized by a same-sign variation over most regions of EA–AUS, represents a declining trend of DTR. The second mode, featuring an opposite-sign variation, represents the interannual variations in DTR. The two modes are both closely associated with the changes in cloud cover (CLT) caused by atmospheric circulation anomalies in EA–AUS. For the trend mode, anomalous southerly and northerly winds over EA and AUS, respectively, bring warm and wet air from low latitudes to EA–AUS, inducing an increase in CLT and thereby reducing DTR in most areas of EA–AUS. The changes of circulation are mainly due to the thermodynamic responses of atmosphere to the nonuniform warming in EA–AUS. In addition, the second mode of DTR is largely forced by the ENSO variability. The weakened Walker circulation associated with warm ENSO events triggers a pair of anomalous low-level anticyclones (south and north of the equator) over the western Pacific. The AUS region is under control of the southern anticyclone, thereby reducing the CLT and increasing the DTR in AUS as a result of anomalous descending motion. In contrast, the EA region is controlled by anomalous southerlies to the west of the northern anticyclone. The northward transports of moistures from the warm ocean increase the CLT, reducing DTR in EA.

1. Introduction

The observed global average temperatures have been increasing since the midtwentieth century, and the rate of change has been amplified by anthropogenic forcing

(Meehl et al. 2004; IPCC 2013; Q. Li et al. 2017). Impacts of climate change include changes not only in mean temperatures but also in the variability and extreme values of the diurnal temperature range (DTR) (e.g., Zhai and Pan 2003) and precipitation (e.g., Zhai et al. 2005; Guo et al. 2014, 2016; Li et al. 2016; Guo et al. 2017). It has long been known that the observed global warming is generally characterized by a greater increase in minimum temperatures than maximum temperature, which induces a decline in the DTR represented by the difference between the maximum temperature and

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minimum temperature during the past decades (e.g., Karl et al. 1993; Easterling et al. 1997; Zhai and Ren 1999; Durre and Wallace 2001; Vose et al. 2005; Lauritsen and Rogers 2012). The decline in the DTR has been identified as an index of global climate change (Braganza et al. 2004). In addition, the changes in the DTR may also affect the crop yield (Lobell 2007) and the net exchange of carbon dioxide in the ecosystems strongly depends upon diurnal temperature changes (e.g., Yi et al. 2010). It is therefore of great importance to elucidate the DTR changes under global warming. Many previous studies have sought to identify the controlling factors of DTR changes (e.g., Stenchikov and Robock 1995; Dai et al. 1999, 2006; Stone and Weaver 2002, 2003; L. Zhou et al. 2007; Liu et al. 2016). Cloud cover and precipitation are widely recognized to play an important role in the variations of DTR due to their significant influences on the surface energy and hydrological balance (e.g., Karl et al. 1993; Dai et al. 1999; Xia 2013). The clouds can cool Earth's surface by reflecting downward solar shortwave radiation in daytime and emitting the downwelling longwave radiation back to surface in nighttime, thereby exerting a significant damping effect on DTR. The precipitation can influence the DTR not only by inducing the increase in the cloud cover, but also by strengthening the evaporative cooling attributed to more soil moisture in rainy days. Therefore, the negative correlation between DTR and precipitation variations largely results from the close association of precipitation with cloud cover and soil moisture (e.g., Dai et al. 1997; Zhou et al. 2009a). Up to 80% of the variance of DTR could be explained by cloud and precipitation records (Dai et al. 1999). In addition, other factors can contribute to the DTR variability, including changes in greenhouse gases (GHGs), aerosols, soil moisture, and surface properties (e.g., Hansen et al. 1995; Mitchell et al. 1995; Stenchikov and Robock 1995; Collatz et al. 2000; Stone and Weaver 2003; Zhou et al. 2004; L. Zhou et al. 2007; Dai et al. 2006; Huang et al. 2006; Liu et al. 2016; Xu et al. 2017).

Although some climate factors, such as cloud cover, precipitation, and anthropogenic aerosols, are regarded as the causes for the variations in DTR (e.g., Dai et al. 1999; Zhou et al. 2009a; Xia 2013; Liu et al. 2016; Z. Li et al. 2017), limited studies have been carried out to analyze the atmospheric circulation related to the changes in DTR (e.g., Wu 2010; Ionita et al. 2012). Since cloud cover and precipitation are both significantly influenced by atmospheric circulation, the changes in atmospheric circulation may play a nonnegligible role in the variations of DTR. Previous studies have demonstrated that the climate anomaly in East Asia (EA) shows some potential relationship to that in Australia

(AUS) as a result of the coupling of atmospheric circulation between EA and AUS during boreal winter (e.g., Chen et al. 2000; Wang et al. 2009a; Zhang and Zhang 2010; Gong et al. 2015). Although there have been some studies on the DTR change in EA (e.g., Zhai and Ren 1999; Xia 2013; Shen et al. 2014) and AUS (e.g., Plummer et al. 1995) separately, the possible covariations of DTR between EA and AUS have not been investigated yet. Therefore, an interesting question is raised: Are there coherent variations of DTR between EA and AUS during boreal winter? If so, what roles do cloud cover and precipitation play in the variations of DTR in EA and AUS and what are the dynamic mechanisms associated with the changes of DTR in these regions?

In this study, the empirical orthogonal function (EOF) method is at first employed to identify the large-scale pattern of the DTR change in the EA–AUS region. The roles of atmospheric circulation and related dynamic mechanism responsible for the DTR changes are further investigated. The rest of the paper is organized as follows. Section 2 describes data and methods used in this study. The large-scale pattern of boreal winter DTR variations and its relationship to the cloud cover and precipitation in EA–AUS are presented in section 3. In section 4, we discuss the dynamic mechanisms of atmospheric circulation responsible for the variations of wintertime DTR in EA–AUS. A summary of the key findings and discussion are given in section 5.

2. Data and methods

a. Data

The observational monthly DTR, total cloud cover (CLT), and precipitation (Pre) used in this study are provided by the Climatic Research Unit (CRU) of the University of East Anglia [Time Series version 4.00 (TS4.00)], with a horizontal resolution of $0.5^\circ \times 0.5^\circ$ covering the period from 1901 to 2015 (Harris et al. 2013; Harris and Jones 2017). To investigate the atmospheric processes associated with DTR variations, the present study uses the observational proxies of atmospheric circulation and air temperature data from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al. 1996) covering the period from January 1948 to the present with a horizontal resolution of $2.5^\circ \times 2.5^\circ$. To verify the results from the NCEP–NCAR reanalysis, a parallel analysis based on JRA-55 (Kobayashi et al. 2015; Harada et al. 2016) has to be performed, although this reanalysis covers a shorter time period (1958–2015) at a horizontal resolution of

$1.25^\circ \times 1.25^\circ$. The CRU dataset has been bilinearly interpolated to a resolution of $2.5^\circ \times 2.5^\circ$ (latitude $\times$ longitude) for convenience of analyzing the relationship to the atmospheric changes based on the reanalysis data. The sea surface temperature (SST) data from the monthly mean Extended Reconstructed Sea Surface Temperature, version 3 (ERSST.v3) dataset (Xue et al. 2003; Smith et al. 2008) were employed to investigate the possible linkage between SST and DTR variations; the data have a $2^\circ \times 2^\circ$ horizontal resolution and cover the period from January 1854 to the present. The present analysis is based on the time period from 1948 to 2015, during which the above variables are available.

b. Methods

The EOF analysis has been a widely used statistical method in the climate research since it was first introduced to meteorology (e.g., Lorenz 1956; von Storch and Zwiers 1999). In particular, it is a useful tool to extract the dominant patterns with coherent temporal variations in a geophysical field (e.g., Hannachi et al. 2007; Monahan et al. 2009). In this study, the EOF method was applied to reveal the dominant patterns of DTR variations over EA–AUS. The Niño-3.4 index defined as SST anomaly averaged over the equatorial east-central Pacific region (5°S – 5°N , 170° – 120°W) was adopted to represent El Niño–Southern Oscillation (ENSO) (Rasmusson and Carpenter 1982). In this study, the boreal winter was considered and the winter means were constructed by averaging monthly mean data of December, January, and February (DJF). Here, our convention is that the winter of 1948 refers to the 1947/48 winter. Regression and correlation analyses were used to investigate the linkage of CLT and Pre as well as the associated atmospheric processes to the leading EOF patterns of DTR variations. The significance of a 95% confidence level ($\alpha = 0.05$ level) of the results was evaluated with the two-tailed Student's t test.

3. The leading modes of the winter DTR variations in EA–AUS

The EOF analysis is applied to winter DTR in the region of 40°S – 50°N , 90° – 155°E to reveal the characteristics of the DTR variations in the EA–AUS region. The EOF analysis was carried out by constructing an area-weighted covariance matrix. In an area-weighted covariance matrix, a $\cos\theta$ weighting, where θ refers to latitude, was applied to the analyzed variable, taking into account of the convergence of the meridians with latitude. In the following discussion, the first two leading modes are analyzed, both of which account for about 45% of the total variances of DTR in EA–AUS. These

leading modes are well separated from the rest of the modes, according to the criteria proposed by North et al. (1982).

a. EOF1 of the DTR: Trend mode

The spatial pattern of the first EOF mode (EOF1) is characterized by the same-sign loading over almost all the regions of EA–AUS and an opposite loading in part of southwestern Australia (Fig. 1a). The EOF1 explains 27.3% of the total variance of the DTR in the EA–AUS region. The temporal evolution of the corresponding principal component (PC1) exhibits a sharp drop during the whole time period of 1948–2015 (Fig. 1c). The power spectrum of PC1 shows that the PC1 is dominated by extreme low-frequency variations (Fig. 1e). The linear trend of PC1 reaches -0.43 standard deviation per decade. This pronounced trend of PC1 implies that the EOF1 actually reflects the trend of DTR in the EA–AUS region. To verify our hypothesis, the linear trend of wintertime DTR is displayed in Fig. 2. It can be clearly seen that the winter DTR shows a significantly decreasing trend in most areas of EA–AUS, as opposed to an increasing trend in part of southwestern Australia. The spatial distribution of the DTR trend is almost the same as the EOF1 pattern. This confirms that the trend is the dominant characteristic of DTR variability in EA–AUS, which agrees with previous studies that revealed a decreasing DTR trend over most of the globe (e.g., Karl et al. 1993; Easterling et al. 1997; Vose et al. 2005).

Also, the DTR has been widely recognized to be significantly associated with CLT and Pre because of their damping effects on the solar radiation (e.g., Dai et al. 1999; Zhou et al. 2009a; Xia 2010). An increase in cloud cover reduces the downward solar radiation, contributing to surface cooling during the daytime, and enhances downward longwave radiation, contributing to surface warming during the nighttime, which results in a net effect of decrease in DTR (Dai et al. 1999). To illustrate the possible relationship between EOF1 and CLT, we display in Fig. 3a the correlation map that is obtained by the temporal correlation coefficient between PC1 time series and the CLT time series at each grid point. It shows that the correlation between PC1 and CLT is significantly negative in most areas of the EA–AUS region, and small region with positive correlation is located in southwestern AUS and north of 45°N in EA. This spatial pattern of correlations is quite similar to that of EOF1 (Figs. 1a and 3a), suggesting that the CLT indeed has a good association with DTR in EA–AUS. Moreover, the correlation between PC1 and Pre in EA–AUS is somewhat similar to that of CLT, but with a smaller magnitude in EA (Fig. 3b). This suggests a weaker influence of Pre on DTR associated with EOF1 in EA.

FIG. 1. (a) The spatial pattern of EOF1 in boreal winter DTR in the EA–AUS region, (c) the corresponding PC of EOF1, and (e) the spectrum of PC1. (b),(d),(f) As in (a),(c),(e), but for EOF2. The percentage of the explained variance is given at the top right of (a)–(d).

We estimate the spatial distribution of linear trend for CLT and Pre to further investigate the relationship between long-term DTR and CLT and Pre variations. The spatial pattern of the linear trend of CLT is quite similar to that of the correlation map between CLT and PC1, as shown in Fig. 3a, with positive trends in most parts of EA–AUS, and negative trend in southwestern AUS and north of 45°N in EA. The trend of Pre is much weaker and insignificant in most areas of EA (Fig. 3d) compared with that in CLT. The significant negative correlations between CLT and DTR in most areas of EA–AUS suggest that the cloud cover is a crucial factor in the changes of DTR in EA–AUS. This confirms the previous findings that the damping effect of cloud cover on solar radiation plays an important role in the trend of DTR (Dai et al. 1997, 1999; Xia 2013; Shen et al. 2014).

It is noted that although the trend of CLT in most areas of eastern China is positive, the amplitude of CLT trend is relatively weak, which implies that, in addition to CLT, there may be other factors that lead to the decline in DTR. L. Zhou et al. (2007) reported that the DTR was reduced maximally in dry areas of the western Sahel during a period of intensive drought. They proposed that the decrease in DTR is largely attributed to the reduction in vegetation cover or the reduction in soil emissivity. Liu et al. (2016) pointed out that the anthropogenic aerosol plays an important role in the trend of DTR decrease in eastern China due to its negative radiation forcing in daytime. In consideration of the distinct climatic background and properties of surface, the factors responsible for the decline in DTR over different regions may be complicated. Therefore,

FIG. 2. Linear trends in boreal winter DTR over the EA–AUS region from 1948 to 2015. Dotted areas indicate the significance at the 95% confidence level.

although the CLT is a very important factor contributing to the decrease of DTR in EA–AUS, other factors, such as anthropogenic aerosol and land processes, may also influence the DTR to a certain extent.

b. EOF2 of the DTR: Interannual variability mode

As illustrated in Fig. 1b, the spatial pattern of EOF2 displays an opposite loading between EA and AUS. Overall, the EOF2 accounts for 17.6% of the total variance of DTR in the EA–AUS domain. The corresponding time series (PC2) shows pronounced fluctuation on interannual time scales with a spectrum peak around 2–3 yr (Figs. 1d,f). During the positive phases of PC2, the DTR has significant negative anomaly in most areas of EA and positive anomaly in the whole AUS region, and vice versa. Therefore, the EOF2 actually represents the out-of-phase variations of DTR over EA–AUS on interannual time scale; that is, during the positive phases of PC2, there are negative DTR anomalies in the EA and positive DTR anomalies in the AUS and vice versa.

To illustrate the possible connection between the EOF2 and the CLT and Pre, the correlation map between PC2 and CLT in EA–AUS is presented in Fig. 4a. It shows that there are significant positive correlations between PC2 and CLT in most areas of the EA and significant negative correlations in the whole AUS. This spatial distribution of correlation coefficients is quite

similar to the EOF2 pattern, but with opposite sign (Fig. 1b). Since the PC2 reflects the time evolution of out-of-phase pattern of DTR variations, the opposite polarity between the correlation map and the EOF2 pattern actually indicate the negative effect of CLT on DTR in the EA–AUS region. This result suggests that the variations of EOF2 are tightly related to the variations of CLT in EA–AUS. The significant negative correlation between DTR and CLT over EA–AUS confirms the important role of CLT in the variations of DTR through its damping effect on solar radiation (Dai et al. 1997, 1999; Xia 2013; Shen et al. 2014). The spatial distribution of correlation coefficients between PC2 and Pre is similar to that of CLT, but the magnitudes of correlation coefficients are relatively smaller in EA–AUS (Fig. 4b). Since the variations in the CLT and Pre are not fully independent, to further reveal the importance of CLT on the DTR the variability of Pre is removed from CLT in each grid separately through a linear regression method (Gong et al. 2017) as follows:

$$\text{CLT}_{i,\text{res}} = \text{CLT}_i - r_i \text{Pre}_i,$$

where r_i is the regression coefficient of CLT_i with respect to Pre_i at the grid point i and the $\text{CLT}_{i,\text{res}}$ is the precipitation-independent CLT at the grid point i . The $\text{CLT}_{i,\text{res}}$ is then used to calculate the correlation with DTR. It can be seen clearly that, although the variability of Pre is linearly removed from the CLT in EA–AUS, the correlations between PC2 and the residual CLT are still significant in most areas of EA–AUS (Fig. 4c). It suggests that cloud may be the primary factor for the interannual variations of winter DTR over EA–AUS, although Pre may also be important.

4. Physical processes for the leading variations of wintertime DTR in EA–AUS

The above results indicate that the dominant patterns of the winter DTR variations in EA–AUS are closely tied to the changes in CLT. In this section, we investigate the atmospheric dynamic processes associated with the variation of the cloud and DTR based on the two leading modes. Regression analyses are employed to analyze the change of atmospheric circulation corresponding to the change of the specified indices (e.g., PCs and Niño-3.4). This change is regarded as the anomaly in the following section.

a. Role of nonuniform warming in the trends of DTR

The atmospheric circulation plays an important role in the formation and maintenance of CLT and Pre. For example, in the tropics, the organized deep convective

FIG. 3. (a) Correlation maps between the PC1 and CLT over EA–AUS for 1948–2015. (b) As in (a), but for Pre. (c),(d) As in (a),(b), but for the linear trends of CLT and Pre, respectively. Dotted areas indicate the significance at the 95% confidence level.

clouds form in the ascending branches of the Hadley and Walker circulations, while the low clouds cover the ocean in anticyclonic areas (e.g., Bony et al. 2015). The changes of atmospheric circulation related to the EOF1 of DTR variations are first investigated. Because the EOF1 actually represents the trend of DTR in the EA–AUS region, we mainly analyze the trend of atmospheric circulation over 1948–2015. Figure 5a shows the trends of winter sea level pressure (SLP) and 1000-hPa winds in EA–AUS during 1948–2015.

The SLP shows a pronounced negative trend in most part of China and a positive trend in Japan and the South China Sea (SCS). The reduced contrast of sea–land pressure differences in EA induces the southerly wind

responses along coastal China, weakening the East Asian winter monsoon (EAWM). This result is consistent with the previous findings that the EAWM has been weakening during the past several decades (e.g., W. Zhou et al. 2007; Wang et al. 2009b; Li and Yang 2017). The southerly anomalies reduce the transport of cold and dry air from higher latitudes over EA, which is beneficial to the increase of atmospheric moisture. Meanwhile, the changes in winds feature cyclonic anomalies over EA that are conducive to the ascending motion of atmosphere. These processes contribute to an increase in the cloud amount but a reduction in the DTR in EA (Figs. 2 and 3c).

Meanwhile, there are positive SLP trends mostly along the coastal AUS, whereas the trend of SLP is

FIG. 4. (a) Correlation maps between PC2 and CLT over EA–AUS for 1948–2015. (b),(c) As in (a), but for Pre and precipitation-independent CLT (CLT_{res}), respectively. Dotted areas indicate the significance at the 95% confidence level.

negative in central AUS (Fig. 5a). The resultant change in winds shows an anomalous cyclone response in the central AUS, accompanied by northerly anomalies in the central-eastern AUS and southerly anomalies in the southwestern AUS (Fig. 5a). The northerly wind anomalies bring more warm and wet air from low latitudes to eastern AUS. The anomalous cyclonic winds in central AUS induce anomalous ascending motion, thereby facilitating the increase in cloud and decrease in the DTR in eastern-central AUS. Meanwhile, weak

southerly anomalies are present over southwestern AUS and they transport the relatively cold air from mid-latitudes to southwestern AUS, leading to a small region of decrease in clouds (Fig. 3c), hence leading to increase in DTR in the southwestern AUS (Fig. 2).

Since the changes of atmospheric circulation are closely related to the changes of thermal condition of atmosphere itself, the trends of 1000-hPa air temperature (T_{1000}) during 1948–2015 in the EA–AUS region are examined. Although the global average temperature

FIG. 5. (a) Linear trends in SLP (shading) and 1000-hPa winds (vectors) in the EA–AUS region from 1948 to 2015. (b) As in (a), but for T_{1000} . The vectors shown in (a) indicate the trends of 1000-hPa winds statistically significant at the 95% confidence level. Dotted areas in (a) and (b) indicate the significance of the SLP and T_{1000} at the 95% confidence level, respectively.

is increasing over the past several decades due to increased GHGs, the trends of temperature on different regions are not uniform because of the different properties of surface (IPCC 2013). The trends of T1000 indeed show distinct differences in EA–AUS (Fig. 5b). Over EA, there is large warming north of 30°N in China, but small warming in the western North Pacific and even cooling in Japan and southern China. Consequently, the land–sea thermal gradient is weakened in EA. This reduces the land–sea pressure difference, thereby generating anomalous southerly wind responses along coastal China (Fig. 5a). Over AUS, there is cooling along the west coast of AUS and weak warming in central AUS (Fig. 5b). This thermal condition induces an anomalous cyclone response in central AUS, accompanied by northerly anomalies in central-eastern AUS and southerly anomalies in southwestern AUS (Fig. 5a). It is noted that the trend of wind shows an anomalous anticyclone along the east coast of AUS centered around 160°E, which also contributes to the northerly anomalies in eastern AUS. However, the center of the anomalous anticyclone does not match well with the cooling center of temperature trend located east of 160°E. It implies that although the local thermodynamic response of atmosphere to nonuniform surface warming appears to be an important reason for the changes of circulation, other processes may also contribute to the circulation changes in AUS.

The above results suggest that anomalous water vapor transportation and vertical motions associated with anomalous circulation may significantly influence the trends of cloud cover, thereby affecting the trends of DTR in EA–AUS. Meanwhile, the trends of atmospheric circulation are mainly attributed to the atmospheric thermodynamic responses to the nonuniform trend of temperature in EA–AUS, especially in EA. It is noted that the nonuniform warming in the EA–AUS may be contributed by the anthropogenic forcing. Although the radiation forcing associated with the increasing GHG emission is relatively uniform on globe, the response of regional temperature could be different as a result of the distinct climatic background and surface properties (IPCC 2013). Therefore, the decline in DTR in the EA–AUS region may also be influenced by the anthropogenic forcing. Previous studies found that the climate models can simulate a trend of decrease in the DTR, but with a much smaller magnitude than that found in observations (e.g., Karoly et al. 2003; Braganza et al. 2004; Zhou et al. 2009b, 2010). Therefore, the present results may serve to provide insight into plausible reasons for the weak trend of DTR in the climate models; that is, the climate models may not well reproduce the observational nonuniform warming.

b. Role of ENSO in the interannual variability of DTR

To investigate the factors governing the variations of EOF2, the atmospheric circulation and SST anomalies associated with EOF2 are examined. Figure 6a reveals the regressed 850-hPa wind and SST anomalies in tropical Indo-Pacific Ocean on the normalized time series of PC2. During the positive phases of PC2, an anomalous lower-tropospheric anticyclone is located in the western North Pacific, which is accompanied by anomalous southerly winds along the northwest of the anticyclone (Fig. 6a). The southerly anomalies over the EA not only weaken the EAWM, but also bring warm and wet air from low latitudes to EA and induce the increase in CLT and Pre, thereby reducing the DTR in EA (Figs. 4a,b). AUS is controlled by a large anomalous anticyclone over the east Indian Ocean to the entire AUS region. The anomalous descending motion associated with the anomalous anticyclone can significantly reduce the CLT and Pre in the western-central AUS (Fig. 4). Meanwhile, the southerly anomalies in the eastern part of the anomalous anticyclone bring the relatively cold and dry air to eastern AUS and reduce the CLT and Pre there. Therefore, both processes lead to the increase in DTR in AUS.

As shown in Fig. 6a, the SST anomaly pattern associated with PC2 manifests an evident warming in the tropical eastern-central Pacific and Indian Ocean during the positive phases of PC2. This pattern of SST anomalies is quite similar to that of ENSO that is a dominant interannual mode of climate variability. Figure 6b shows the regression map of SLP, 200-hPa velocity potential, and divergent wind anomalies relative to PC2. There are significant positive and negative SLP anomalies in the western and eastern Pacific, respectively. This pattern is also similar to the classic Southern Oscillation (e.g., Ji et al. 2015). In the upper troposphere, an anomalous divergence center is located in the eastern-central Pacific and convergence centers are mainly located in western Pacific and North Pacific (Fig. 6b). The locations of the dipole structure of convergence and divergence centers in the tropical Pacific well match the dipole structure of SLP anomaly pattern, which indicates that the Walker circulation is weakened in association with the significant warming in the tropical eastern-central Pacific, with a complementary contribution from the tropical Indian Ocean warming. Therefore, the anomalous descending motion associated with the weakening Walker circulation forms a pair of lower-tropospheric anomalous anticyclones over the tropical western Pacific (Fig. 6a). The center of the anomalous anticyclone on the north side of the equatorial western Pacific shifts eastward slightly compared to

FIG. 6. (a) Regression maps of (a) SST (shading) and 850-hPa winds (vectors) against the normalized PC2 for 1948–2015. (b) As in (a), but for SLP (shading), 200-hPa velocity potential (contours; $0.1 \times 10^6 \text{ m}^2 \text{ s}^{-1}$), and divergent winds (vectors). Dotted areas indicate the significance at the 95% confidence level.

that in the Southern Hemisphere because of the strong positive local air–sea feedback in the western North Pacific (Wang et al. 2000). These results suggest that the out-of-phase interannual variations of the winter DTR between EA and AUS are mainly attributed to the opposite effects of the atmospheric circulation caused by the SST anomalies in tropical Pacific. It is noted that ENSO represents the natural variability in the climate system, and therefore this opposite-sign pattern of the DTR changes in the EA–AUS region may be attributed to the natural change and be independent of the anthropogenic forcings.

The above analysis indicates that the EOF2 may be largely influenced by ENSO. To further examine the influence of ENSO, 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) was employed. Figure 7 displays the normalized time series of the Niño-3.4 index and PC2. There are consistent variations between PC2 and the Niño-3.4 index and the correlation coefficient between the two time series reaches 0.37 during 1948–2015, exceeding the 99% confidence level. This confirms that the interannual variability of DTR in EA–AUS represented by EOF2 is largely influenced by ENSO.

To delineate the influence of ENSO on the circulation anomalies over EA–AUS in detail, the regressed SST,

FIG. 7. The normalized Niño-3.4 index (blue line) and PC2 (red line). The correlation coefficient r between Niño-3.4 index and PC2 is shown in the top-left corner of the figure.

850-hPa wind, SLP, 200-hPa velocity potential, and corresponding divergent wind anomalies onto normalized Niño-3.4 index over tropical Indo-Pacific are presented in Fig. 8. It can be seen that the spatial pattern of ENSO-related SST and circulation anomalies are similar to those of PC2 (Fig. 6). During the positive phases of the Niño-3.4 index, the pattern of SST anomalies associated with Niño-3.4 index shows evident warming in the tropical eastern-central Pacific (Fig. 8a). The resultant weakened Walker circulation (Fig. 8b) triggers a pair of the lower-level anticyclone north and south of the equatorial western Pacific (Fig. 8a). Southerly anomalies in the west side of the anomalous anticyclone near the Philippine Sea bring warm and wet air from low latitudes to EA and induce the increase in CLT and Pre, thereby reducing the DTR in EA (Fig. 8a). The AUS region is also controlled by a large anomalous anticyclone, extending from the central Indian Ocean to the South Pacific. The anomalous descending motion associated with the anomalous anticyclone can significantly reduce the CLT and Pre in AUS (Fig. 8a).

To further elucidate the influences of ENSO on the interannual variations of DTR, CLT, and Pre, the correlation maps between the Niño-3.4 index and DTR, CLT, and Pre in EA–AUS are presented in Fig. 9. It shows that the correlation between the Niño-3.4 index and DTR is basically opposite over EA and AUS (Fig. 9a). Meanwhile, the correlations between the Niño-3.4 index and CLT and Pre are broadly opposite in EA and AUS. The results indicate that ENSO induces opposite cloud and precipitation changes over EA and AUS, which in turn contribute to opposite variations of DTR between EA and AUS on the interannual time scale. It is noted that correlations are weak in the Maritime Continent (Fig. 9). This phenomenon may be attributed to the sparse observational stations and relatively low data quality in this region.

5. Summary and discussion

This study attempts to unveil the large-scale patterns of the winter DTR changes in EA–AUS and to investigate the physical processes responsible for these variations of DTR based on the observational and

FIG. 8. As in Fig. 6, but against the Niño-3.4 index.

reanalysis data over the past 68 years (1948–2015). The EOF analysis reveals two leading modes with distinct winter DTR variations in EA–AUS. These two modes have different time scales and dynamic mechanisms.

The first mode is characterized by a same-sign variation in most areas of EA–AUS. The corresponding PC1 shows a decreasing trend during 1948–2015, which suggests that the EOF1 represents the trends of DTR in EA–AUS. The trends of DTR are mainly related to the changes of CLT in EA–AUS. The influence of Pre is relatively weak in EA–AUS, especially in EA. The changes of CLT in EA–AUS are closely tied to the changes of atmospheric circulation. The pronounced southerly trends over EA induce the northward transport of moisture from the low latitudes to EA. Meanwhile, the cyclonic wind anomalies in EA are also conducive to the ascending motion, thereby increasing the CLT and reducing the DTR in most areas of EA during the past decades. In addition, the change of wind shows an anomalous cyclone in central AUS accompanied by the northerly wind anomalies in eastern AUS and southerly wind anomalies in southwestern AUS. The anomalous cyclone induces a significant increase in CLT with the ascending motion, thus decreasing the DTR in central AUS during the past decades. The northerly wind anomalies bring the warm wet air from low latitudes to eastern AUS, leading to the increase in clouds and decrease in DTR in eastern AUS. The weak southerly anomalies transport the relative cold air from midlatitudes to southwestern AUS, inducing a decrease in clouds and hence an increase in DTR in the southwestern AUS during the past decades.

FIG. 9. (a) Correlation maps between the winter Niño-3.4 index and DTR over the EA–AUS region for 1948–2015. (b),(c) As in (a), but for CLT and Pre, respectively. Dotted areas indicate the significance at the 95% confidence level.

Furthermore, the changes of atmospheric circulation are largely due to the thermodynamic responses of atmosphere to the nonuniform warming of temperature in EA–AUS. The strongest warming in most areas of China and relatively weak warming in western North Pacific and cooling in Japan reduce the land–sea thermal contrast, thereby generating a southerly response in EA. The warming in AUS is also not uniform, with cooling in most of the coast of AUS but weak warming in central AUS. This thermal condition may trigger an anomalous cyclone response in central AUS, accompanied by the northerlies and southerlies anomalies in eastern-central and southwestern AUS, thereby increasing CLT in eastern-central AUS and decreasing CLT in southwestern AUS. Since the response of regional temperature to the radiative forcing associated with the increasing GHGs emission could be different because of the distinct climatic background and surface properties (IPCC 2013), the decline in DTR caused by the non-uniform warming in the EA–AUS region may result from anthropogenic forcing.

The second leading mode presents an opposite variation in DTR between EA and AUS, and shows fluctuations on interannual time scales. This mode is closely tied to CLT and Pre change, and the former plays a more important role in the variations of DTR. The second mode of DTR variations is found to be largely forced by the ENSO variability. Meanwhile, the remarkably opposite variations of DTR in EA and AUS are primarily due to the opposite effect of ENSO on the CLT and Pre in EA and AUS. During the positive phases of ENSO, the weakened Walker circulation associated with the warming in tropical eastern-central Pacific triggers a pair of anomalous low-level anticyclone over western Pacific. The CLT decreases in AUS owing to the descending motion related to the anomalous anticyclone which covers the entire AUS, thereby leading to increase the DTR in AUS. In contrast, the anomalous southerlies along the west of the anomalous anticyclone around the Philippine Sea transport warm and wet air to EA and lead to the increase in CLT, thus reducing the DTR in EA. Since ENSO is the natural variability in the climate system, this opposite-sign pattern of the DTR changes in the EA–AUS region may be largely attributed to the natural change.

Meanwhile, it is also noted that although the EOF2 is largely influenced by the ENSO, there are some years when the PC2 and Niño-3.4 index do not match well and are even opposite (Fig. 7). This indicates that some other factors may also contribute to the interannual DTR variations in the EA–AUS region. For instance, the Indian Ocean basinwide SST anomaly, generally regarded as a response to ENSO (e.g., Xie et al. 2009), can in turn affect the ENSO, including its amplitude,

evolution, and climate impacts (e.g., Wu and Kirtman 2004; Kug and Kang 2006; Terray et al. 2016). In particular, the Indian Ocean warming (cooling) at the peak of El Niño (La Niña) in boreal winter was found to be able to modulate the effect of El Niño on air temperature over the African, Asian, and Australian landmasses via cloud forcing (Herold and Santoso 2017). Previous modeling studies (e.g., Terray et al. 2016) showed that without feedback from the Indian Ocean, the period of ENSO would be longer compared to the irregular ENSO period ranging from 2 to 7 yr. The short period (2–3 yr) ENSO signals as seen in the power spectrum of EOF2 (Fig. 1f), therefore, potentially highlight the role of the Indian Ocean. The detailed dynamic mechanisms regarding the influence of the Indian Ocean on DTR change need to be further investigated in future studies.

We perform a parallel analysis using JRA-55 for the period 1958–2015. The dominant modes of DTR variations in the EA–AUS region in the period of 1958–2015 show almost the same patterns as those in the period of 1948–2015. The trends of winter SLP, 1000-hPa winds, and 1000-hPa air temperature during 1958–2015 based on JRA-55 display major features similar to those during 1948–2015 based on the NCEP–NCAR reanalysis (Fig. S1 in the supplemental material). Corresponding to the EOF2, the El Niño–like SST pattern, a pair of lower-level anticyclones north and south of the equatorial western Pacific, the weakened Walker circulation represented by the dipole structure of SLP anomaly, and the upper-level convergence and divergence anomaly centers in the tropical Pacific are well reproduced in the JRA-55 dataset (Fig. S2 in the supplemental material).

The present study focuses on the DTR changes in EA–AUS region during boreal winter when the ENSO signal is the strongest. A question arises as to the status of DTR changes in the other seasons and the underlying plausible factors. Our preliminary analysis indicates that the leading patterns of DTR changes in the other seasons are not the same as those in boreal winter (figure not shown). This discrepancy may be related to the climatic background and the surface properties. For example, the SST anomaly patterns are quite different in different seasons. During boreal summer, the ENSO SST patterns have decayed and the SST anomalies are neutral in the eastern-central Pacific, whereas the Indian Ocean has the strongest warming during this season and significantly impacts the climate of surrounding regions (e.g., Yang et al. 2007; Xie et al. 2009). Hence, the change of atmospheric circulation forced by the SST anomalies is also significantly different compared with that in the boreal winter. More systematic investigation is needed to unravel the main features of DTR changes in other seasons and the factors and associated processes.

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