Relationship between interannual changes of summer rainfall over Yangtze River Valley and South China Sea–Philippine Sea: Possible impact of tropical zonal sea surface temperature gradient

Yao Ha1,2,3 | Zhong Zhong1,2,3 | Yun Zhang1 | Jinfeng Ding1,3 | Xiangrong Yang1,4

1College of Meteorology and Oceanography, National University of Defense Technology, Nanjing, China
2Key Laboratory of Meteorological Disaster of Ministry of Education, and Collaborative Innovation Center on Forecast and Evaluation of Meteorological Disasters, Nanjing University of Information Sciences and Technology, Nanjing, China
3Jiangsu Collaborative Innovation Center for Climate Change and School of Atmospheric Sciences, Nanjing University, Nanjing, China
475839 Army of PLA, Guangzhou, China

Abstract

This study investigates the interannual variability of rainfall over the middle and lower reaches of Yangtze River Valley (MLYRV) and that over the South China Sea and Philippine Sea (SCS-PS) during boreal summer (June–August) from 1979 to 2012. Results exhibit out-of-phase rainfall variations between the MLYRV and the SCS-PS, which is closely related to the tropical zonal sea surface temperature gradient (ZSG) between the northern Indian Ocean (NIO; 5°–25°N, 60°–100°E) and the equatorial central and eastern Pacific (CEP; 5°S–5°N, 180°–130°W). The ZSG can explain as much as 40% of the total variance of the summer rainfall over the MLYRV and the SCS-PS, which is closely related to the tropical zonal sea surface temperature gradient (ZSG) between the northern Indian Ocean (NIO; 5°–25°N, 60°–100°E) and the equatorial central and eastern Pacific (CEP; 5°S–5°N, 180°–130°W). The ZSG can explain as much as 40% of the total variance of the summer rainfall over the MLYRV and the SCS-PS in the past 40 years. This is much higher than that explained by the NIO SST (24%), CEP SST (14%) and Niño-3.4 index (16%) alone. A positive ZSG between the warm NIO and the cold CEP tends to increase rainfall over the MLYRV and decrease rainfall over the SCS-PS, whereas a negative ZSG between the cold NIO and the warm CEP is generally favourable for rainfall over the SCS-PS and unfavourable for rainfall over MLYRV. The close connection between the ZSG and the out-of-phase interannual rainfall variation over the two regions can be explained by influences of the ZSG on atmospheric circulation over East Asia. A positive ZSG induces anomalous easterlies and Walker-like circulation in the tropics, which results in an anomalous subsidence and boundary layer divergence over the SCS-PS. As a result, summer rainfall decreases in this region. Meanwhile, moisture transport increases due to the anomalously strong southwesterlies along the northwestern flank of the intensified western Pacific subtropical high, providing more precipitable water to the MLYRV region. In contrast, a negative ZSG induces surface westerlies and favourable environmental condition...
for rainfall over the SCS-PS. Dry descending flow induced by local anomalous Hadley circulation develops over the MLYRV around 30°N, which is unfavourable for rainfall over the MLYRV. The mechanism further examined using numerical experiments. These thermal-dynamical processes induced by the ZSG work together to cause the out-of-phase interannual changes of rainfall between the MLYRV and the SCS-PS, suggesting that the ZSG is highly indicative of the interannual change of summer rainfall in the two regions.

KEYWORDS
equatorial central and eastern Pacific, interannual variability, middle-lower reaches of Yangtze River Valley, northern Indian Ocean, out-of-phase relationship, Philippine Sea, South China Sea, summer rainfall, tropical zonal SST gradient

1 INTRODUCTION

East Asian summer monsoon is characterized by complicated changes in rainfall with large spatial–temporal variations. The tremendous interannual variation of summer monsoon over East Asia makes an accurate prediction of summer rainfall change over China a crucial but challenging issue. The interannual variability of the East Asian summer monsoon rainfall is affected by multiple factors, and one of the most critical sources of predictability is sea surface temperature (SST) anomalies, which are most closely associated with the El Niño-Southern Oscillation (ENSO) (Yang et al., 2012; Zuo et al., 2013; Yim et al., 2014; Wang et al., 2015; He et al., 2016). Previous studies have revealed that rainfall increases over the middle-lower reaches of Yangtze River Valley (MLYRV) and decrease over South China, the South China Sea (SCS) and the western Philippine Sea (PS) often occur simultaneously. These rainfall changes are considered to be related to anomalous western Pacific anticyclonic circulation forced by SST warming over the tropical Indian Ocean (e.g., Li et al., 2008; Xie et al., 2009; Hu and Duan, 2015) and cooling over the equatorial central Pacific Ocean (e.g., Wang et al., 2013; Xiang et al., 2013). The anomalous anticyclonic circulation together with the water vapour transport over the western Pacific is closely associated with the intensification of the western Pacific subtropical high during the El Niño decaying summer (Chang et al., 2000a; Zhou and Yu, 2005; Li et al., 2014). Wang et al. (2013) proposed that the variability of the western Pacific subtropical high is significantly affected by SST anomalies (SSTAs) over the tropical central Pacific and Indian Ocean due to a positive atmosphere–ocean feedback mechanism, which provides a source of predictabilities of climate and rainfall over East Asia. In addition, the seasonal movement of the South Asian high is favourable for rainfall intensification and maintenance over the MLYRV. On the long-term time scale, the South Asian high exhibits an evident southeastward-shifting trend, coinciding with the increasing trend of rainfall in the MLYRV and the decreasing trend of rainfall in northern China in the past 20 years (Yoo et al., 2006; Yang et al., 2007; Zhou et al., 2009; Huang et al., 2012). This change is related to the sustaining SST warming in the tropical Indian Ocean (Wei et al., 2012; 2014; Ha et al., 2016).

Summer rainfall over the SCS-PS is affected by multiscale systems such as the SCS monsoon trough, the western North Pacific (WNP) monsoon trough, tropical cyclone, tropical depressions, easterly waves, etc., while interactions between these systems also play an important role (Chen and Weng, 1996; Chen et al., 2011; 2017a; 2017b). Particularly, activities of tropical cyclones over the SCS and PS can bring frequent rainstorm along coastal regions of East and Southeast Asia, posing great threat to human life and property (Ha et al., 2014; Ha and Zhong, 2015). Previous studies revealed that ENSO contributes directly and indirectly to the interannual variability of rainfall over the SCS-PS region in boreal summer (Wang et al., 2000; 2003; Wu et al., 2003; 2009a; 2010c). Due to the modulation of ENSO on atmospheric circulation, convection, and SST in the SCS-PS, the anomalous anticyclone over the WNP in El Niño decaying summer is associated with reduced local rainfall in the SCS-PS and positive rainfall anomalies from East China to southern part of Japan (Wu et al., 2009a; He and Wu, 2014). Moreover, the SST anomalies over the tropical Indian Ocean can also affect the climate variability in the SCS-PS (Wu and Wang, 2000; Yang et al., 2007; Li et al., 2008; Xie et al., 2009; Wu et al., 2010b; 2014). El Niño-induced warming in the tropical Indian Ocean basin plays a crucial role by forcing equatorial Kelvin waves, which lead to anticyclonic shear and boundary layer divergence that can extend to the western Pacific. The resultant Ekman divergence over north of the equatorial region suppresses
convection over the WNP (Xie et al., 2009; Wu et al., 2010a). Anomalous convection and associated heating over the SCS-PS can further affect climate over East and Southeast Asia. Therefore, examining the interannual variability of SCS-PS rainfall and its influencing factors can improve our understanding of the impact of tropical Indo-Pacific oceans on East and Southeast Asian climate.

It is noted that the climate variability and rainfall change in the MLYRV and SCS-PS are subjected to both remote forcing of anomalous SST over the tropical Indian Ocean and tropical Pacific Ocean that is closely related to ENSO (Zhou et al., 2010; Hu et al., 2014; Hu and Wu, 2015). Meanwhile, the anomalous state of climate in the SCS-PS can impact East Asian climate (Xie et al., 2003). Huang and Sun (1994) pointed out that the thermal conditions and convective activities over the areas from the Philippines to the Indo-China Peninsula can impact the summer rainfall variation in the Yangtza-Huaihe River valley, which is connected to the meridional movement of the western Pacific subtropical high. Therefore, the meridional pattern of anomalous atmospheric circulation over East Asia actually connects the tropical and subtropical regions. For example, the severe flood in the Yangtze River valley in 1998 is closely related to the warm event over the SCS-PS from 1997 to 1998 (Wang et al., 2002). Based on the fact that weather and climate over the MLYRV and the SCS-PS are linked by the anomalous western Pacific anticyclonic circulation associated with the ENSO cycle and anomalous SST, the relationship between interannual changes of summer rainfall in the above two regions need be further identified. In the present study, the following scientific questions will be addressed based on observational analyses and numerical experiments. Which factor and/or physical process dominate the interannual rainfall relationship between the two regions, and whether the influential factor can be a better indicator to illustrating the interannual variability of summer rainfall over East Asia? We will focus on the relationship between the interannual variabilities of rainfall over the MLYRV and the SCS-PS in boreal summer (June–August), and explore the mechanism behind these changes in terms of the anomalous large-scale circulation on the interannual time scale.

This paper is organized as follows. Dataset and methodology used in this study are described in Section 2. Detailed characteristics of observed out-of-phase interannual changes in summer rainfall between the MLYRV and the SCS-PS are presented in Section 3. Influences of atmospheric circulation anomalies on the interannual rainfall changes over the two regions are analysed in Section 4. Section 5 discusses influences of the tropical zonal SST gradient (ZSG) between the different oceans on summer rainfall in the MLYRV and the SCS-PS. Finally, Section 6 presents conclusions and discussion.

2 | DATASET AND METHODOLOGY

We use three rainfall datasets to detect features of interannual changes in summer rainfall over the MLYRV and the SCS-PS. The first rainfall dataset is derived from version 2 of the Global Precipitation Climatology Project (GPCP) with a spatial resolution of $2.5^\circ \times 2.5^\circ$ and covers the period from 1979 to 2012 (Adler et al., 2003). This product provides global precipitation from an integration of various satellite datasets over land and ocean, and a gauge analysis over land. The second rainfall dataset is the monthly Climate Prediction Center Merged Analysis of Precipitation (CMAP) with a spatial resolution of $2.5^\circ \times 2.5^\circ$ (Xie and Arkin, 1997). Rainfall amounts in this dataset are obtained from several kinds of satellite estimates, gauge data and the National Centers for Environmental Prediction–National Center for Atmospheric Research Reanalysis Project Dataset (NCEP-NCAR) Reanalysis Precipitation values. The third rainfall dataset is the monthly National Oceanic and Atmospheric Administration Precipitation Reconstruction (NOAA-PREC) with a spatial resolution of $2.5^\circ \times 2.5^\circ$ (Chen et al., 2002). In this dataset, the land portion is defined by optimum interpolation of gauge observations at over 17,000 stations, while the oceanic portion is produced by Empirical Orthogonal Function (EOF) reconstruction of historical gauge observations over islands and coastal areas.

Other reanalysis data used in this study include monthly atmospheric temperature, geopotential height, relative humidity, horizontal and vertical velocity fields on global $2.5^\circ \times 2.5^\circ$ grids at 17 standard pressure levels. These data are extracted from the NCEP-NCAR Reanalysis Project Dataset (Kalnay et al., 1996). The SST data used here is from the monthly Extended Reconstructed Sea Surface Temperature dataset (ERSST; Smith et al., 2008) with a spatial resolution of $2^\circ \times 2^\circ$. The atmospheric moisture flux is vertically integrated from 1,000 hPa to 300 hPa, which is expressed as $\int_{500 \text{ hPa}}^{1000 \text{ hPa}} q dp$, where $q$ is the specific humidity, $g$ is the gravitational acceleration, $p$ is pressure, and $\mathbf{v}$ is the horizontal wind vector. The monthly Niño-3.4 index is obtained from the NOAA. The reanalysis dataset is also available at the Earth System Research Laboratory of NOAA (http://www.esrl.noaa.gov/psd/data/reanalysis/reanalysis.shtml).

The EOF analysis is applied in this study to illustrate the dominant modes of rainfall anomalies over East Asia during summers from 1979 to 2012. Additionally, composite analysis and linear regression analysis are applied to the reanalysis data, and the climatological mean state in summers of 1979–2012 is taken as the reference state. All anomalies are relative to the reference state. Significance test is conducted using the Student’s $t$-test.
To explore rainfall variation and atmospheric response to different SST boundary conditions, especially to the tropical ZSG, the NCAR Community Atmospheric Model version 3.1 (CAM3; Collins et al., 2006), which is the atmospheric component of the community Earth System Model (CESM) is employed in this study. This atmospheric general circulation model (AGCM) can be used both as the atmospheric component of the CESM, and as a stand-alone model, which is able to estimate the equilibrium response of atmospheric circulation to external forcings (Taschetto and Ambrizzi, 2010; Feng et al., 2013; He and Wu, 2014; Chen et al., 2017a; 2017b). The horizontal resolution of the AGCM is T42 (approximately 2.8° × 2.8°), with 26 hybrid vertical levels. More details of this model can be referred to http://www.cccsm.ucar.edu/models/atm-cam/docs/description. The horizontal resolution of the AGCM is employed in this study. This atmospheric general circulation model (AGCM) can be used both as the atmospheric component of the community Earth System Model (CESM), and as a stand-alone model, which is able to estimate the equilibrium response of atmospheric circulation to external forcings (Taschetto and Ambrizzi, 2010; Feng et al., 2013; He and Wu, 2014; Chen et al., 2017a; 2017b). The horizontal resolution of the AGCM is T42 (approximately 2.8° × 2.8°), with 26 hybrid vertical levels. More details of this model can be referred to http://www.cccsm.ucar.edu/models/atm-cam/docs/description.

3 | OUT-OF-PHASE INTERANNUAL CHANGES IN RAINFALL BETWEEN THE MLYRV AND THE SCS-PS

Figure 1 shows the spatial pattern and principal component of the first EOF mode of summer rainfall over East Asia. The first EOF mode explains 23.1% of the total variance, which is significant in statistics distinguished from the rest of the eigenvectors. This mode presents a southeast–northwest oriented tripolar pattern, which is characterized by enhanced rainfall over the MLYRV and Japan, and suppressed rainfall from the northern and central SCS to the PS (Figure 1a). The principal component (PC) of the first EOF mode exhibits an obvious interannual variability (Figure 1b) with two dominant peaks of approximately 2- and 7-year on the interannual time scales (Figure 1c). The statistical significance of the above two peaks in a power spectrum are tested according to the power spectrum of mean red noise, and the result indicates that the interannual signal associated with the dipolar rainfall change over East Asia is robust above the 95% confidence level. This dominant interannual mode is proved to be closely related to the ENSO cycle, which has been extensively studied in previous works (e.g., Chang et al., 2000a; Wang and Li, 2004; Wang et al., 2008; Zhu et al., 2014). Besides, note that the interannual variability of the first leading mode shows a clear out-of-phase relationship between the MLYRV and the SCS-PS region, that is, the increased rainfall over the MLYRV is accompanied by concurrent decreased rainfall over the SCS-PS, and vice versa. The second EOF mode accounts for 11.4% variance out of the total variance, which is significant in statistical test. The PC of the second EOF mode presents a decadal variability with significant periods on the interannual-to-interdecadal time scales (not shown).

To further determine the features of the out-of-phase interannual changes in rainfall over the MLYRV and SCS-PS, we calculate simultaneous correlation between the SCS-PS rainfall series (area-weighted mean over 10°–16°N, 115°–140°E) and rainfall field over East Asia (Figure 2a). Positive coefficients are located over the central SCS and PS, while negative correlations are found over the MLYRV. This pattern is reasonable and the correlation coefficients over the above two regions are significant at/above 95% confidence level. On the other hand, the correlation analysis between the summer rainfall series over the MLYRV (area-weighted mean over 28°–34°N, 106°–120°E) and the regional rainfall exhibits significant negative correlation over the central SCS-PS (Figure 2b), which is opposite to that shown in Figure 2a. The above result indicates that summer rainfall changes in the two regions are highly negatively correlated, and the negative correlation reaches −0.62 at/above the 99% confidence level (Figure 2c and Table 1). Meanwhile, it is noted that PC1 of the first EOF mode is highly correlated with the MLYRV (SCS-PS) summer rainfall with the correlation coefficients generally larger than 0.57 (−0.92) at the 99% confidence level (Table 1). The high correlation between PC1 and rainfall series in the two

![Figure 1](wileyonlinelibrary.com)
regions implies that PC1 is able to represent this out-of-phase feature of summer rainfall between the MLYRV and SCS-PS. The analysis above is based on the GPCP rainfall data. The correlation of summer rainfall over the two regions is still evident using other two independent rainfall datasets of the CMAP data and the NOAA Reconstruction data (not shown), suggesting that the out-of-phase interannual changes in summer rainfall over the MLYRV and SCS-PS are robust. We identify a positive (negative) year when the normalized PC1 is more (less) than 0.6 (−0.6). Based on the thresholds of standard deviation from the PC1, nine positive and 10 negative phase years are determined and listed in Table 2.

To reveal the spatial distribution of the out-of-phase interannual rainfall changes, Figure 3 shows anomalies of rainfall rate and vertically integrated moisture flux and its divergence in the composite years. In the positive years, the anomalous rainfall presents a dipolar pattern with increased (decreased) rainfall in the MLYRV (SCS-PS). Extensive negative anomalies of rainfall are found in the central SCS and the PS at around (17°N, 119°E) and (15°N, 133°E), respectively (Figure 3a). The vertically integrated moisture flux manifests an anticyclonic pattern over East Asia, which leads to plenty of moisture transported from the SCS to the MLYRV by the southwesterly and thus is favourable for the development of rainfall in the MLYRV (Wang and Wang, 2013). The integrated moisture divergence shows a dipole pattern similar to that of anomaly in rainfall with the moisture convergence (divergence) located over the MLYRV (SCS-PS) (Figure 3b). In the negative years, rainfall anomalies and associated moisture are very similar to that in the positive years for both spatial distribution and quantity, but with the opposite phase. Positive (negative) anomalies of rainfall are found over the SCS-PS (MLYRV; Figure 3c), and a cyclonic pattern of the vertically integrated moisture flux indicates atmospheric water vapour convergence (divergence) in the SCS-PS (MLYRV; Figure 3d).

4 | ATMOSPHERIC CIRCULATION ANOMALIES

Figure 4 shows the environmental fields simultaneously regressed upon PC1 of the first EOF rainfall mode. The

| Year       | Positive years | 1980, 1983, 1988, 1993, 1996, 1998, 2007, 2008, 2010 |
|------------|----------------|--------------------------------------------------|
| Negative years | 1981, 1984, 1985, 1990, 1994, 1997, 2001, 2002, 2009, 2012 |
atmospheric circulation in the upper level in association with the increased (decreased) MLYRV (SCS-PS) rainfall is characterized by significant anomalous westerlies (easterlies) around 30°N (20°N), corresponding to anomalous anticyclone (cyclone) in the MLYRV (SCS-PS) (Figure 4a). This distribution implies that the southward shift of the westerly

**FIGURE 3** Anomalies of (a), (c) summer rainfall (shadings; mm/day) and (b), (d) vertically integrated moisture flux (vectors; g m⁻¹ s⁻¹) from 1,000 to 300 hPa and its divergence (shadings; 10⁻⁷ g m⁻² s⁻¹) in the composite positive (upper panels) and negative (lower panels) years. Shaded areas in (a)–(d) and bold wind vectors in (b), (d) indicate the differences are statistically significant above the 90% confidence level [Colour figure can be viewed at wileyonlinelibrary.com]

**FIGURE 4** Simultaneous regression upon the PC1 for (a) 200-hPa zonal wind (vectors; m/s), (b) 850-hPa zonal wind (vectors; m/s), (c) 500-hPa vertical $p$ velocity (contours; 10⁻³ pa/s), (d) relative humidity (contours; %) averaged from 850- to 500-hPa and (e) 500-hPa geopotential height (shadings; gpm) in the period 1979–2012. Bold wind vectors in (a), (b) and shaded areas in (c)–(e) indicate the regression coefficients are statistically significant above the 95% confidence level. “AC” and “C” indicate anticyclonic and cyclonic circulations, respectively [Colour figure can be viewed at wileyonlinelibrary.com]
Jet stream and the intensified South Asian high occur in the positive years. The intensification of the westerlies to the south of the jet stream can induce strong zonal divergence in the upper level and lead to heavy local rainfall, which explains the increased convective activity found in the MLYRV (Zhou and Yu, 2005; Yu and Zhou, 2007). Simultaneous regressions of the lower-to-middle level wind field, vertical velocity and humidity upon PC1 present an anticyclonic circulation with descending motion occurring over the SCS-PS (Figures 4b–d). These conditions lead to decreased summer rainfall in the SCS-PS. Meanwhile, significant moisture transport by the southwesterly further intensifies over the eastern China, leading to moisture convergence over the MLYRV. The anomalously wet environmental condition in the MLYRV enhances rainfall in this region. The regression of PC1 onto 500 hPa geopotential height shows a monopole pattern with significant positive anomaly centre located over the northern SCS (Figure 4e), indicating that the western Pacific subtropical high becomes stronger and shifts westward. These anomalies are directly associated with the out-of-phase interannual changes in summer rainfall between the MLYRV and the SCS-PS.

To better understand the effect of large-scale circulation anomalies on changes in summer rainfall over the two regions, composite analysis is applied to atmospheric circulations in the positive and negative years. The composite anomalies of environmental conditions are generally consistent with the regression results mentioned above. In the positive years, anomalous anticyclonic and cyclonic circulations in the upper level are found at around 25° and 15°N, respectively (Figure 5a). Meanwhile, the anticyclonic circulation in the lower level over the WNP intensifies (Figure 5b). The thermodynamic condition in the mid-level can influence the intensity of summer rainfall via affecting vertical motion and moisture condition (Li et al., 2013). The western Pacific region is under control of the anomalous anticyclone in the

**FIGURE 5** Anomalies of (a), (e) 200-hPa wind (vectors; m/s), (b), (f) 850-hPa wind (vectors; m/s), (c), (g) 500-hPa vertical p velocity (contours; pa/s), and (d), (h) relative humidity (contours; %) averaged from 850- to 500-hPa in the composite positive (left panels) and negative (right panels) years. Bold wind vectors in (a), (b), (e), (f) and shaded areas in (c), (d), (g), (h) indicate the differences are statistically significant above the 90% confidence level. “AC” and “C” indicate anomalous anticyclonic and cyclonic circulations, respectively [Colour figure can be viewed at wileyonlinelibrary.com]
lower level, where abnormally dry condition and descending flow suppress convection, leading to less than normal precipitation. In contrast, ascending motion prevails over the MLYRV region. Combined with abundant moisture supply in this region, the ascending motion leads to active convective activities and large amounts of rainfall (Figure 5c,d). In the negative years, the spatial pattern of atmospheric circulation anomaly and environmental variables is opposite to that in the positive years. Anomalous cyclonic and anticyclonic centres in the upper level are found to the north of Taiwan and over the Philippines, respectively (Figure 5e), and an anomalous cyclonic circulation is located over East Asia (Figure 5f). Spatial distribution of vertical motion and moisture in the negative years presents a southeast–northwest oriented tripolar pattern with enhanced precipitation over the SCS-PS and decreased precipitation over the MLYRV and Japan (Figure 5g,h). This meridional wave train pattern is similar to the East Asian-Pacific and/or Pacific-Japan teleconnection patterns (Nitta, 1987; Li et al., 2014).

Figure 6a shows the difference in 500 hPa geopotential height between the positive and negative years. It can be seen that positive geopotential height anomaly centre in the mid-level is located over the northern SCS. The western Pacific subtropical high extends further westward compared to that in the negative years and the climatological mean, which is corresponding to the intensified anticyclone and anomalous descending motion over the SCS-PS shown in Figure 5. The intensified subtropical high is unfavourable for rainfall in the SCS-PS region, but it can promote rainfall over the MLYRV due to the enhanced southwesterly and increased moisture transport along the northwestern flank of the subtropical high.

South Asian high is a persistent anticyclonic system in the upper troposphere and lower stratosphere of the Northern Hemisphere during boreal summer, and its intensity and location can affect East Asian summer monsoon rainfall (Zhang et al., 2002; Wei et al., 2014; 2017). Figure 6b shows the difference in 200 hPa geopotential height between positive and negative years. It is found that the South Asian high becomes stronger in the positive years with the positive anomaly centre located over southwestern China (Figure 6b). This pattern indicates that the South Asian high shifts more southeastward in positive years. An anomalous anticyclone forms to the east of the coastal region of China at around 25°N, whereas a cyclonic anomaly develops over the SCS-PS in the upper level (Figure 5a). Divergence and convergence anomalies well correspond to well with the more rainfall in the MLYRV and less rainfall in the SCS-PS, respectively. When the South Asian high shifts northwestward in the negative years, the pattern of atmospheric circulation anomaly is opposite as shown in Figures 5e–h, which result in the less (more) rainfall over the MLYRV (SCS-PS). The anomalous circulation pattern associated with the meridional shifts of South Asian high and western Pacific subtropical high can cause inverse conditions of moisture supplement and vertical flows in the MLYRV and the SCS-PS, which can explain the out-of-phase rainfall pattern over the two regions.

5 | INFLUENCES OF THE TROPICAL ZONAL SST GRADIENT ON SUMMER RAINFALL

Many previous studies have shown that the tropical SST forcing over the tropical Indian Ocean and equatorial Pacific can trigger anomalous atmospheric circulation over East Asia, which subsequently leads to interannual changes in rainfall in this region (e.g., Chang et al., 2000a; 2000b; Zhou et al., 2009; Wu et al., 2009a). To identify the driving force that controls the out-of-phase interannual changes in rainfall over the MLYRV and the SCS-PS, we calculate the simultaneous correlation between PC1 and SST filed from 1979 to 2012 (Figure 7a). Results show significant positive correlation over the northern Indian Ocean (NIO) and the WNP.
around 120°E, with significant negative correlation over the equatorial central and eastern Pacific (CEP). Figure 7b illustrates the composite difference in summer SST between the positive and negative years. Obvious SST warming is found over the NIO north of 5°S and the WNP. Meanwhile, the SST anomaly displays an extensive cooling over the CEP. Note that the distribution of correlation coefficient highly resembles that of the SST difference between the positive and negative years, implying that changes in SST over the NIO, WNP and CEP may have important influences on interannual variations in environmental fields and rainfall over the two target regions in the past 40 years.

Based on the spatial pattern of significant SSTA as shown in Figure 7, we identify two key regions where the SST changes are most highly correlated with PC1 that is associated with rainfall over the MLYRV and the SCS-PS (the solid boxes in Figure 7a,b). Figure 8a shows the SST averaged from 1979 to 2012 over the two regions, which are referred to as the NIO (5°–25°N, 60°–100°E) and the CEP (5°S–5°N, 180°–130°W) regions. The simultaneous correlation coefficients of PC1 with SST over the NIO and CEP during the period 1979–2012 are listed in Table 3. It can be seen that the SST indices over the NIO and CEP are well correlated with the simultaneous PC1 with significant correlation coefficients of 0.45 and −0.38, respectively (Table 3).

However, simply considering the NIO or CEP alone seems to be not sufficient to explain the PC1 variability associated with the out-of-phase interannual changes in the two regions. The joint effect of the two oceans is examined to investigate whether the variability of summer rainfall can be captured. Taking into account the opposite conditions of SSTA over the NIO and the CEP, a ZSG index is constructed based on SST difference between the NIO and CEP. A positive ZSG index is related to warmer SST over the NIO and cooler SST in the CEP, while a negative ZSG index

---

**FIGURE 7** (a) Simultaneous correlation between SST and PC1. Calculations cover the period 1979–2012, and dark (light) shaded areas indicate where the correlation coefficients are statistically significant above the 99% (95%) confidence level. (b) Differences of SST in the summer between the composite positive and negative years. Shaded areas indicate the differences are statistically significant above the 90% confidence level. The boxes indicate the areas of the NIO (5°–25°N, 60°–100°E) and CEP (5°S–5°N, 180°–130°W), where time series of SST are calculated [Colour figure can be viewed at wileyonlinelibrary.com]

**TABLE 3** Simultaneous correlations of PC1 with the NIO SST, CEP SST, and ZSG indices in JJA from 1979 to 2012. The second column lists the correlations computed based on 7-year running mean time series. Solid (hollow) triangles indicate significance at the 99% (95%) confidence level. The effective degrees of freedom are used when estimating the significance for the running mean time series

|                  | Interannual | 7-year period |
|------------------|-------------|---------------|
| PC1 with NIO SST | 0.49 ▲      | 0.45 △        |
| PC1 with CEP SST | −0.38 △     | −0.56 ▲       |
| PC1 with ZSG     | 0.63 ▲      | 0.65 ▲        |

---

**FIGURE 8** Time series of (a) NIO SST and CEP SST and (b) normalized ZSG and PC1 for the period 1979–2012. Linear trends of SST have been removed [Colour figure can be viewed at wileyonlinelibrary.com]
summer rainfall, and the coefficients above 0.63 exceed the 99.9% confidence level (Table 3). The correlation between ZSG and PC1 is much higher than that between the NIO-/CEP SST indices and PC1, suggesting that the simultaneous ZSG is highly indicative of the relationship of interannual changes in summer rainfall between the MLYRV and the SCS-PS. The ZSG index can be identified as an important factor that controls the interannual variability of summer rainfall over the MLYRV and SCS-PS.

It is well known that the SSTAs associated with the ENSO cycle are considered to be key predictors for East Asian summer rainfall change (Wang et al., 2000; He et al., 2016). The ENSO-related positive rainfall anomalies first occur in southern China in the developing phase of ENSO, and then migrate northeastward to the MLYRV and Japan in the spring and summer during the decaying phase. These rainfall changes are associated with the variation of low-level atmospheric circulation in the WNP (Wu et al., 2003; Wu et al., 2009b; Wang and Wang, 2013). In this section, we detect the lead–lag relationship between the ENSO-related indices and the out-of-phase summer rainfall changes and investigate the influence of remote SSTA on the rainfall changes over East Asia. Figure 9a shows the seasonal lead–lag correlations of the summer PC1 with the Niño-3.4, CEP SST, NIO SST, and ZSG indices during the period 1979–2012. Here, the preceding, current and subsequent years are denoted by −1, 0 and 1, respectively. The Niño-3.4 index in the preceding autumn (SON[−1]) and winter (D[−1]JF[0]) is significantly positively correlated with the current summer rainfall above the 95% confidence level ($r = 0.42$) during the whole period. However, the simultaneous correlation in JJA(0) reverses to significant negative one ($r = -0.40$) at the 95% confidence level. The same feature can be seen in the relationship between the CEP SST and PC1. Meanwhile, the positive correlation between the NIO SST and PC1 exhibits two peak values located in the preceding autumn (SON[−1]; $r = 0.39$) and current summer (JJA[0]; $r = 0.49$), respectively. Furthermore, it is worth noting that PC1 is most significantly correlated with simultaneous ZSG index with the correlation coefficient of 0.63 that exceeds the 99.9% confidence level. The relationship between the ZSG index and PC1 is much more robust than those between Niño-3.4, CEP and NIO SST indices and PC1 in the simultaneous summer and the preceding cold seasons. This means that the variation of the summer ZSG can explain as much as 40% of the total variance of the interannual variability in the rainfall changes over the MLYRV and the SCS-PS. This number is much higher than that explained by the NIO SST (24%), CEP SST (14%) and Niño-3.4 index (16%), suggesting that the ZSG index is highly indicative of the relationship between interannual changes of summer rainfall in the two regions. Figure 9b shows the simultaneous spatial correlation between summer rainfall and the ZSG index. Positive correlations are found from the MLYRV to southern Japan, and negative correlations are centred over the SCS and eastern PS. Here, our results indicate that the ZSG between the NIO SST and CEP SST can play a more substantial role in shaping the interannual variabilities of summer rainfall over the MLYRV and the SCS-PS.

To reveal the physical processes for the linkage between the ZSG index and the out-of-phase interannual changes in anomalous rainfall and atmospheric circulation over the MLYRV and SCS-PS, we investigate the differences in 200-hPa velocity potential and divergent wind between the positive and negative years (Figure 10a). It can be seen that the anomaly of velocity potential displays a planet-scale zonal wave train over the tropical Indo-Pacific Ocean. An upper-level divergent centre is located at the equatorial western Indian Ocean around 70°E. The anomalous zonal winds converge and form an upper-level meridional mass sink from 150°E to the tropical eastern Pacific. East Asia is influenced by the upper level westerly and southwesterly. These patterns are associated with the La Niña forcing.
Figure 10b shows the longitude-height section of differences in velocity and relative humidity averaged from the equator to 20°N between the positive and negative years. Corresponding to the ZSG caused by the NIO warming and CEP cooling (Figure 10c), anomalous ascending motions are observed from 50°E to 80°E over the tropical Indian Ocean, transporting the mid-tropospheric moist air mass upwards. Meanwhile, a strong descending branch of the anomalous zonal circulation dominates from east of 110°E to the eastern Pacific (Figure 10b), which corresponds to the upper-level mass sinking band shown in Figure 10a. The anomalous subsiding flow with dry air over the SCS-PS reinforces the western Pacific subtropical high (Figure 6a). It is noted that the cold SSTA over the CEP usually occurs in the decaying phase of El Niño (McPhaden and Zhang, 2009; Ohba and Ueda, 2009), and the ENSO-induced tropical India Ocean warming can induce a low-level anomalous anticyclone over the western Pacific in the subsequent summer (Xie et al., 2009). Hu et al. (2012; 2013) demonstrated that ENSO can affect rainfall and high temperature extremes in China through an anomalous WNP anticyclone and descent-related adiabatic warming. These conditions suppress synoptic-scale disturbances and convection development, and further lead to the decrease in summer rainfall over SCS-PS in the positive years.

Local meridional or Hadley circulation plays a crucial role connecting the tropics and subtropics (Chang et al., 2000b; Wei et al., 2012), and thus affects the meridional
rainfall distribution. Figure 10d shows the latitude-height cross section of differences in summer velocity and relative humidity averaged from 100°E to 125°E between the positive and negative years. A remarkable descending motion dominates the region from 5°N to 25°N, which is located over the SCS-PS. Note that this is the common descending branch of the anomalous Walker and Hadley circulations (Figure 10b), which contributes to the dry condition and westward extension of the western Pacific subtropical high over the SCS-PS. This condition of meridional circulation leads to decreased SCS-PS rainfall in the positive years. Meanwhile, anomalous ascending flow occurs over the MLYRV around 30°N, where the atmosphere is moist in the lower-to-middle troposphere (Figure 10d). Such kinds of conditions are favourable for rainfall over the MLYRV, but lead to decreased rainfall over the SCS-PS.

To confirm the influence of the tropical ZSG between the NIO and CEP on summer rainfall variation over the MLYRV and the SCS-PS, several numerical experiments using the CAM3 model are conducted in the present section. The control experiment is based on a 20-year integration with climatological SST prescribed in the global oceanic domain (CTL run). The results of control run are used as a reference state. The observed climatological rainfall and atmospheric circulation conditions during summer are well represented by the control run. The sensitivity experiments are integrated for 22 years, and results from the latter 20 years are used to construct a 20-member ensemble mean to reduce uncertainties arising from different initial conditions. The amplitude of imposed positive/negative SSTA forcings are same as the differences of SST between the composite positive and negative years shown in Figure 7b over the areas 5°–25°N, 60°–100°E (NIO) and 5°S–5°N, 180°–130°W (CEP). To emphasize the impact of tropical ZSG on summer atmospheric circulation and rainfall over East Asia, we designed six sensitivity experiments, which are divided into three groups. In the ZSG-forced group, the warm (cold) NIO SSTA and cold (warm) CEP SSTA in JJA are imposed, and climatological SST is imposed in other regions (+ZSG/-ZSG run). In the NIO-forced group, the warm/cold SSTA forcing is prescribed only in the NIO region, and climatological SST is imposed in other regions (+NIO/-NIO run). While in the CEP-forced group, the warm/cold SSTA forcing is specified only in the CEP region, and climatological SST is imposed in other regions (+CEP/-CEP run).

Figures 11 and 12 show the differences in rainfall and circulation between the positive/negative SSTA sensitivity experiments and the control run (left/middle panels), and those between the positive SSTA sensitivity experiments and the control run (right panels). (a)–(c) ZSG-forced runs, (d)–(f) NIO-forced runs, and (g)–(i) CEP-forced runs. “+” and “−” indicate positive and negative SSTA, respectively [Colour figure can be viewed at wileyonlinelibrary.com]
and the negative ones (right panels) during summertime. In general, the ZSG-forced runs can capture well the circulation and rainfall features as seen in the observational results in the former section. Particularly, in the positive ZSG-forced case (positive NIO SSTA and negative CEP SSTA), below-normal rainfall is found over the northern SCS and the PS, and above-normal one appears over southeastern China and southern Japan (Figure 11a). Anomalous easterlies prevail over the equatorial western Pacific, and an anomalous low-level anticyclonic circulation together with an intensified western Pacific subtropical high can be seen over the WNP (Figure 12a). In the negative ZSG-forced run (negative NIO SSTA and positive CEP SSTA), rainfall anomalies show the opposite phase relative to those in the positive ZSG-forced run, with above-normal (below-normal) rainfall over the northern SCS-PS (MLYRV; Figure 11b). Anomalous westerlies in the lower level extend from the eastern Indian Ocean to the equatorial central Pacific, and an anomalous cyclonic circulation is located over East Asia (Figure 12b). The difference between the positive ZSG-forced run and the negative one presents a clearer dipolar pattern with increased (decreased) rainfall in the MLYRV (SCS-PS; Figure 11c). Meanwhile, the western Pacific region is under control of the anomalous anticyclone in the lower level (Figure 12c).

For the NIO/CEP SSTA forcing cases, the positive NIO-forced run and the negative CEP-forced run manufacture consistent low-level easterlies anomalies over the equatorial western Pacific (Figure 12d,h). Negative rainfall anomalies are found over the SCS-PS and the northwestern SCS in the negative CEP-forced run and the positive NIO-forced run, respectively (Figure 11d,h). The negative NIO-forced run and the positive CEP-forced run exhibit the similar the circulation and rainfall features with anomalous westerlies in the tropics and above-normal rainfall over the SCS-PS (Figure 11e,g and 12e,g). It is noted that the NIO-forced or CEP-forced cases cannot display the out-of-phase rainfall pattern between the MLYRV and the SCS-PS, meanwhile the magnitude of anomalies in rainfall and atmospheric circulation in the NIO-forced runs and CEP-forced runs tend to be smaller than those in the ZSG-forced runs (Figures 11f,i and 12f,i).
The results of numerical experiments further indicate that the tropical ZSG is highly indicative of the relationship between interannual changes of summer rainfall over the MLYRV and the SCS-PS. Corresponding to the changes of thermodynamic conditions over East Asia, which are associated with anomalous Walker and Hadley circulations forced by the tropical ZSG between the NIO and CEP, summer rainfall over the MLYRV and SCS-PS exhibits out-of-phase interannual changes in the past 40 years.

6 | CONCLUSIONS AND DISCUSSION

Statistical analyses and numerical experiments are performed in the present study to investigate interannual variability of rainfall over the MLYRV and the SCS-PS in boreal summer (June–August) during 1979–2012. Correlation analysis reveals out-of-phase interannual variations of rainfall over the MLYRV and SCS-PS. Both observational analyses and numerical experiments with specified SST forcing to demonstrate that the interannual changes in the above two regions are closely related to tropical zonal SST gradient (ZSG) between the northern Indian Ocean (NIO; 5°–25°N, 60°–100°E) and the equatorial central and eastern Pacific (CEP; 5°S–5°N, 180°–130°W). A positive ZSG between the warm NIO and the cold CEP tends to increase rainfall over the MLYRV but decrease rainfall in the SCS-PS, whereas a negative ZSG between the cold NIO and the warm CEP is generally favourable (unfavourable) for rainfall over the SCS-PS (MLYRV). The relationship between the ZSG and the out-of-phase rainfall changes in the two regions is prominent with a temporal correlation coefficient of 0.63 that is much higher than coefficients of the NIO SST, CEP SST or Niño index with rainfall changes. The variation of the simultaneous ZSG can explain as much as 40% of the total variance of the summer rainfall variabilities over the MLYRV and the SCS-PS, suggesting that the ZSG is highly indicative of the interannual changes of summer rainfall in the two regions. The index can be identified as an important factor controlling the interannual variability of summer rainfall over the MLYRV and the SCS-PS.

Figure 13 presents the schematic illustration of the processes that are involved in the ZSG and interannual rainfall changes over the two regions. Induced by the positive (negative) ZSG between the warm (cold) NIO and the cold (warm) CEP SSTAs, the anomalous easterlies (westerlies) between the NIO and CEP lead to anomalous Walker-like circulation. Anomalous ascending (descending) motions are triggered over the NIO from

**FIGURE 13** Schematic diagram illustrating anomalous atmospheric circulation in lower level (850-hPa) and upper level (200-hPa) in the (a) composite positive and (b) negative years. Red and green shaded areas indicate warm and cold SST anomalies, respectively. Dark and light arrows indicate anomalous zonal and meridional circulation branches, respectively. “AC” and “C” indicate anomalous anticyclonic and cyclonic circulations, respectively. “Con” and “Div” indicate convergence and divergence in the upper level, respectively [Colour figure can be viewed at wileyonlinelibrary.com]
50°E to 80°E. The descending (ascending) branch of the anomalous Walker circulation becomes dominant from the Maritime Continent to the CEP region, leading to abnormally dry (moist) lower-level air over the SCS-PS. As a result, rainfall decreases (increases) in this region. Corresponding to the anomalous easterlies (westerlies) over the tropics, the western Pacific subtropical high and the South Asian high become stronger (weaker) and extend westward and eastward (eastward and westward), respectively, in the positive (negative) years. Atmospheric moisture transport increases (decreases) due to the anomalous southwesterly along the northwestern flank of the subtropical high, providing more (less) precipitable water to the MLYRV region. Meanwhile, corresponding to the descending (ascending) motion dominated over the SCS-PS, moist (dry) ascending (descending) flow develops over the MLYRV around 30°N, which is induced by the local anomalous meridional circulation. The thermal-dynamical processes induced by changes of the ZSG work together to cause the out-of-phase interannual changes of rainfall between the MLYRV and the SCS-PS. It is indicated that the ZSG between the NIO and the CEP plays a more substantial role in shaping the interannual variabilities of summer rainfall over the MLYRV and the SCS-PS.

This paper mainly focuses on the study of interannual summer rainfall changes over the MLYRV and the SCS-PS, which are found to be significantly related to the ZSG change between the NIO and the CEP. The anomalies in large-scale circulation associated with rainfall changes are also induced by the ZSG. Future in-depth studies of the mechanisms and quantitative investigation of the impact of ZSG on the atmospheric environment and rainfall over East Asia will be conducted in detail. In addition, it is noted that the correlation between the ZSG and summer rainfall over the MLYRV/SCS-PS shows inconsistent features in different decades. That is, remarkably positive correlation between PC1 and the ZSG is found in the recent period of 1987–2012, with the coefficient reaching up to 0.77 above the 99.99% confidence level. However, the correlation is insignificant in the former period of 1979–1987 \( (r = 0.28) \). This contrast implies that the interannual relationship between the ZSG and summer rainfall over the MLYRV and the SCS-PS may have inter-decadal variation on the long-term time scale. In the future, we will enlarge the study period to further investigate the cause of the interdecadal variability of this interannual relationship.

REFERENCES

Adler, R.F., Huffman, G.J., Chang, A., Ferraro, R., Xie, P.P., Janowiak, J., Rudolf, B., Schneider, U., Curtis, S., Bolvin, D., Gruber, A., Susskind, J., Arkin, P. and Nelkin, E. (2003) The version-2 global precipitation climatology project (GPCP) monthly precipitation analysis (1979-present). Journal of Hydrometeorology, 4, 1147–1167.

Chang, C.P., Zhang, Y. and Li, T. (2000a) Intertropical and interdecadal variations of the east Asian summer monsoon and tropical Pacific SSTs, part I: roles of the subtropical ridge. Journal of Climate, 13, 4310–4325.

Chang, C.P., Zhang, Y. and Li, T. (2000b) Intertropical and interdecadal variations of the East Asian summer monsoon and tropical Pacific SSTs, part II: meridional structure of the monsoon. Journal of Climate, 13, 4326–4340.

Chen, J., Wen, Z., Wu, R., Wang, X., He, C. and Chen, Z. (2017a) An interdecadal change in the intensity of interannual variability in summer rainfall over southern China around early 1990s. Climate Dynamics, 48, 191–207.

Chen, M., Xie, P., Janowiak, J.E. and Arkin, P.A. (2002) Global land precipitation: a 50-yr monthly analysis based on gauge observations. Journal of Hydrometeorology, 3, 249–266.

Chen, T.-C., Huang, W.-R. and Yen, M.-C. (2011) Interannual variation of the late spring–early summer monsoon rainfall in the northeastern part of the South China Sea. Journal of Climate, 24, 4295–4313. https://doi.org/10.1175/2011JCLI3930.1.

Chen, T.-C., Tsay, J.-D. and Matsumoto, J. (2017b) Intertropical variation of the summer rainfall center in the South China Sea. Journal of Climate, 30, 7909–7931. https://doi.org/10.1175/JCLI-D-16-0889.1.

Chen, T.-C. and Weng, S.-P. (1996) Some effects of the intraseasonal oscillation on the equatorial waves over the western tropical Pacific–South China Sea region during the northern summer. Monthly Weather Review, 124, 751–756. https://doi.org/10.1175/1520-0493(1996)124<0751:SEOTIO>2.0.CO;2.

Collins, W.D., Rasch, P.J., Boville, B.A., Hack, J.J., McCaa, J.R., Williamson, D.L., Briegleb, B.P., Bitz, C.M., Lin, S.J. and Zhang, M. (2006) The formulation and atmospheric simulation of the community atmosphere model version 3 (CAM3). Journal of Climate, 19, 2144–2161.

Feng, J., Li, J. and Xu, H. (2013) Increased summer rainfall in Northwest Australia linked to southern Indian Ocean climate variability. Journal of Geophysical Research – Atmospheres, 118, 467–480. https://doi.org/10.1029/2012JD018323.

Ha, Y. and Zhong, Z. (2015) Decadal change of tropical cyclone activity over South China Sea around 2002/2003. Journal of Climate, 28, 5935–5951.

Ha, Y., Zhong, Z., Chen, H. and Hu, Y. (2016) Out-of-phase decadal changes in boreal summer rainfall between Yellow-Huaihe River Valley and southern China around 2002/2003. Climate Dynamics, 47, 137–158. https://doi.org/10.1007/s00382-015-2828-2.

Ha, Y., Zhong, Z., Sun, Y. and Lu, W. (2014) Decadal change of South China Sea tropical cyclone activity in mid-1990s and its possible linkage with intraseasonal variability. Journal of Geophysical Research, 119, 5331–5344. https://doi.org/10.1002/2013JD021286.

He, C., Wu, B., Li, C., Lin, A., Gu, D., Zheng, B. and Zhou, T. (2016) How much of the interannual variability of East Asian summer rainfall lies in large-scale circulation associated with rainfall over the MLYRV around 30°N, which is induced by the local anomalous meridional circulation. The thermal-

ORCID

Zhong Zhong https://orcid.org/0000-0002-6145-2698
rainfall is forced by SST? Climate Dynamics, 47, 555–565. https://doi.org/10.1007/s00382-015-2855-z.

He, Z. and Wu, R. (2014) Indo-Pacific remote forcing in summer rainfall variability over the South China Sea. Climate Dynamics, 42, 2323–2337. https://doi.org/10.1007/s00382-014-2123-7.

Hu, J. and Duan, A. (2015) Relative contributions of the Tibetan Plateau thermal forcing and the Indian Ocean Sea surface temperature basin mode to the interannual variability of the east Asian summer monsoon. Climate Dynamics, 45, 2697–2711. https://doi.org/10.1007/s00382-015-2503-7.

Hu, K., Huang, G., Qu, X. and Huang, R. (2012) The impact of Indian Ocean variability on high temperature extremes across the southern Yangtze River valley in late summer. Advances in Atmospheric Sciences, 29, 91–100.

Hu, K., Huang, G. and Wu, R. (2013) A strengthened influence of ENSO on August high temperature extremes over the southern Yangtze River valley since the late 1980s. Journal of Climate, 26, 2205–2221.

Hu, W. and Wu, R. (2015) Relationship between South China Sea precipitation variability and tropical indo-Pacific SST anomalies in IPCC CMIP5 models during spring-to-summer transition. Advances in Atmospheric Sciences, 32, 1303–1318. https://doi.org/10.1007/s00376-015-4250-4.

Hu, W., Wu, R. and Liu, Y. (2014) Relation of the South China Sea precipitation variability to tropical indo-Pacific SST anomalies during spring-to-summer transition. Journal of Climate, 27, 5451–5467. https://doi.org/10.1175/JCLI-D-14-00089.1.

Huang, R., Chen, J., Wang, L. and Lin, Z. (2012) Characteristics, processes and causes of the spatio-temporal variabilities of the East Asian monsoon system. Advances in Atmospheric Sciences, 29, 910–942.

Huang, R., and Sun, F. (1994) Impact of the thermal state and the convective activities in the tropical western warm pool on the summer climate anomalies in East Asia (in Chinese). Scientia. Atmos. Sinica., 18(2), 141–151.

Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Saha, S., White, G., Woollen, J., Zhu, Y., Leetmaa, A., Reynolds, R., Chelliah, M., Ebisuzaki, W., Higgins, W., Janjic, Y., Mo, K.C., Ropelewski, C., Wang, J., Jenne, R. and Joseph, D. (1996) The NCEP/NCAR 40-year reanalysis project. Bulletin of the American Meteorological Society, 77, 437–497.

Li, S., Lu, J., Huang, G. and Hu, K. (2008) Tropical Indian Ocean basin warming and East Asian summer monsoon: a multiple AGCM study. Journal of Climate, 21, 6080–6088.

Li, X., Zhou, W., Chen, D., Li, C. and Song, J. (2014) Water vapor transport and moisture budget over eastern China: remote forcing from the two types of El Niño. Journal of Climate, 27, 8778–8792. https://doi.org/10.1175/jcli-d-14-00649.1.

Li, X., Zhou, W., Li, C. and Song, J. (2013) Comparison of the annual cycles of moisture supply over southwest and Southeast China. Journal of Climate, 26, 10139–10158.

Liu, G., Wu, R., Sun, S. and Wang, H. (2015) Synergetic contribution of precipitation anomalies over northwestern India and the South China Sea to high temperature over the Yangtze River Valley. Advances in Atmospheric Sciences, 32, 1255–1265. https://doi.org/10.1007/s00376-015-4280-y.

McPhaden, M.J. and Zhang, X. (2009) Asymmetry in zonal phase propagation of ENSO Sea surface temperature anomalies. Geophysical Research Letters, 36, L13703. https://doi.org/10.1029/2009GL038774.

Nitta, T. (1987) Convective activities in the tropical western Pacific and their impact on the northern hemisphere summer circulation. Journal of the Meteorological Society of Japan, 65, 373–390.

Ohba, M. and Ueda, H. (2009) Role of nonlinear atmospheric response to SST on the asymmetric transition process of ENSO. Journal of Climate, 22, 177–192.

Smith, T.M., Reynolds, R.W., Peterson, T.C. and Lawrimore, J. (2008) Improvements to NOAA’s historical merged land-ocean surface temperature analysis (1880-2006). Journal of Climate, 21, 2283–2296.

Taschetto, A.S. and Ambrizzi, T. (2010) Can Indian Ocean SST anomalies influence south American rainfall? Climate Dynamics, 38, 1615–1628.

Wang, B., Lee, J.-Y. and Xiang, B. (2015) Asian summer monsoon rainfall predictability: a predictable mode analysis. Climate Dynamics, 44, 61–74. https://doi.org/10.1007/s00382-014-2218-1.

Wang, B. and Li, T. (2004) East Asian monsoon and ENSO interaction. In: Chang, C.-P. (Ed.) East Asian Monsoon. Singapore: World Scientific, pp. 177–212.

Wang, B., Wu, R. and Fu, X. (2000) Pacific–East Asian teleconnection: how does ENSO affect East Asian climate? Journal of Climate, 13, 1517–1536. https://doi.org/10.1175/1520-0442(2000)013<1517:PEATHD>2.0.CO;2.

Wang, B., Wu, R. and Li, T. (2003) Atmosphere–warm ocean interaction and its impacts on Asian–Australian monsoon variation. Journal of Climate, 16, 1195–1211. https://doi.org/10.1175/1520-0442(2003)016<1195:AOIIAI>2.0.CO;2.

Wang, B., Wu, Z., Li, J., Liu, J., Chang, C.-P., Ding, Y. and Wu, G. (2008) How to measure the strength of the East Asian summer monsoon. Journal of Climate, 21, 4449–4463.

Wang, B., Xiang, B. and Lee, J.-Y. (2013) Subtropical high predictability establishes a promising way for monsoon and tropical storm predictions. Proceedings of the National Academy of Sciences, 110, 2718–2722. https://doi.org/10.1073/pnas.1214626110.

Wang, C. and Wang, X. (2013) Classifying El Niño Modoki I and II by different impacts on rainfall in southern China and typhoon tracks. Journal of Climate, 26, 1322–1338.

Wang, D., Xie, Q., Du, Y., Wang, W. and Chen, J. (2002) The 1997–1998 warm event in the South China Sea. Chinese Science Bulletin, 47, 1221–1227. https://doi.org/10.1007/BF02907614.

Wei, W., Zhang, R. and Wen, M. (2012) Meridional variation of south Asian high and its relationship with the summer precipitation over China (in Chinese). Journal of Applied Meteorological Science, 23, 650–659. https://doi.org/10.3969/j.issn.1001-7513.2012.06.002.

Wei, W., Zhang, R., Wen, M., Rong, X. and Li, T. (2014) Impact of Indian summer monsoon on the South Asian high and its influence on summer rainfall over China. Climate Dynamics, 43, 1257–1269. https://doi.org/10.1007/s00382-013-1938-y.

Wei, W., Zhang, R., Wen, M. and Yang, S. (2017) Relationship between the Asian westerly jet stream and summer rainfall over Central Asia and North China: roles of the Indian Monsoon and the South Asian High. Journal of Climate, 30, 537–552. https://doi.org/10.1175/JCLI-D-15-0814.1.

Wu, B., Li, T. and Zhou, T. (2010a) Asymmetry of atmospheric circulation anomalies over the western North Pacific between El Niño and La Niña. Journal of Climate, 23, 4807–4822. https://doi.org/10.1175/2010JCLI3222.1.
Wu, B., Li, T. and Zhou, T. (2010b) Relative contributions of the Indian Ocean and local SST anomalies to the maintenance of the Western North Pacific anomalous anticyclone during the El Niño decaying summer. Journal of Climate, 23, 2974–2986. https://doi.org/10.1175/2010jcli3300.1.

Wu, B., Zhou, T. and Li, T. (2009a) Seasonally evolving dominant interannual variability modes of East Asia Climate. Journal of Climate, 22, 2992–3005. https://doi.org/10.1175/2008jcli2710.1.

Wu, R., Hu, Z.-Z. and Kirtman, B.P. (2003) Evolution of ENSO-related rainfall anomalies in East Asia. Journal of Climate, 16, 3742–3758. https://doi.org/10.1175/1520-0442(2003)016<3742:eoeeraa>2.0.co;2.

Wu, R., Huang, G., Du, Z. and Hu, K. (2014) Cross-season relation of the South China Sea precipitation variability between winter and summer. Climate Dynamics, 43, 193–207. https://doi.org/10.1007/s00382-013-1820-y.

Wu, R. and Wang, B. (2000) Interannual variability of summer monsoon onset over the western North Pacific and the underlying processes. Journal of Climate, 13, 2483–2501.

Wu, R., Wen, Z., Yang, S. and Li, Y. (2010c) An interdecadal change in southern China summer rainfall around 1992/93. Journal of Climate, 23, 2389–2403.

Wu, Z., Wang, B., Li, J. and Jin, F.-F. (2009b) An empirical seasonal prediction model of the East Asian summer monsoon using ENSO and NAO. Journal of Geophysical Research, 114, D18120. https://doi.org/10.1029/2009jd011733.

Xiang, B., Wang, B., Yu, W. and Xu, S. (2013) How can anomalous western North Pacific subtropical high intensify in late summer? Geophysical Research Letters, 40, 2349–2354. https://doi.org/10.1002/2012gl05431.

Xie, P. and Arkin, P.A. (1997) Global precipitation: a 17-year monthly analysis based on gauge observations, satellite estimates, and numerical model outputs. Bulletin of the American Meteorological Society, 78, 2539–2558.

Xie, S.P.; Hu, K., Hafner, J., Tokinaga, H.O., Du, Y., Huang, G., and Sampe, T. (2009) Indian Ocean capacitor effect on Indo-Western Pacific climate during the summer following El Niño. Journal of Climate, 22, 730–747. https://doi.org/10.1175/2008jcli2544.1.

Xie, S.P., Xie, Q., Wang, D. and Liu, W.T. (2003) Summer upwelling in the South China Sea and its role in regional climate variations. Journal of Geophysical Research, 108, 3261. https://doi.org/10.1029/2003Jc001867.

Yang, D., Tang, Y., Zhang, Y. and Yang, X. (2012) Information-based potential predictability of the Asian summer monsoon in a coupled model. Journal of Geophysical Research, 117, D03119. https://doi.org/10.1029/2011jd016775.

Yang, J., Liu, Q., Xie, S.-P., Liu, Z. and Wu, L. (2007) Impact of the Indian Ocean SST basin mode on the Asian summer monsoon.