Recent Changes in ENSO Teleconnection over the Western Pacific Impacts the Eastern China Precipitation Dipole

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ABSTRACT

The eastern China precipitation dipole (ECPD) features an out-of-phase relationship between boreal summer precipitation over the middle and lower reaches of the Yangtze River and the Hetao region to its northwest. The precipitation dipole is strongly influenced by ENSO teleconnections over the western tropical Pacific. Here it is shown that a pronounced weakening of both the rainfall variability over eastern China as well as the precipitation dipole structure occurred around the mid-1990s. The changes have been analyzed by considering two epochs: one during 1979–95 and the other during 1996–2009. The characteristic feature of the circulation anomaly during the first epoch is the well-known East Asia–Pacific/Pacific–Japan (EAP/PJ) pattern, a quasi-meridional teleconnection pattern emanating from the western tropical Pacific. On the other hand, during the latter epoch eastern China precipitation variability occurs as an integral part of the circulation anomalies over the western Pacific. In contrast to the more meridionally restricted anomalies during canonical ENSO episodes, the western Pacific circulation has a significantly larger meridional scale. Intriguingly correlation of the precipitation dipole with Pacific sea surface temperature flips in sign during the second epoch, with enhanced precipitation over southeastern China associated with La Niña–like variability, in contrast to the co-occurrence of enhanced precipitation over this region with El Niño during the first epoch. The results suggest that the dominance of Modoki or central Pacific El Niños, and the altered structure of ENSO teleconnections associated with these, may play a role in the weakened ECPD structure during the latter epoch.

1. Introduction

Agricultural and industrial production in eastern China are strongly affected by variations of the summer monsoon (Wang et al. 2000; Wang 2001; Wu and Wang 2002; Ding and Chan 2005; Ding et al. 2008; B. Wang et al. 2008; Wang et al. 2015). A case in point is the catastrophic floods over the Yangtze River valley (YRV) during the summer of 1998. This caused economic losses exceeding $20 billion (U.S. dollars) and more than 3000 fatalities (National Climate Center 1998; Huang et al. 1998). Another one is the flood that occurred during the summer of 2011 in the middle and lower reaches of the Yangtze River (Yang et al. 2013), after an extreme and persistent drought episode during late winter and spring (Jin et al. 2013). It also caused huge social influences.

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Two of the remarkable patterns in summer rainfall variability of eastern China are the meridional dipole (e.g., Chen et al. 2006; Ding et al. 2008, 2009; Zhu et al. 2011; Huang et al. 2007, 2011, 2012, 2016; Ye and Lu 2012; Sun and Wang 2015) and tripod (e.g., Wei and Zhang 1988; Hsu and Lin 2007; Huang et al. 2007, 2011, 2012, 2016; Ye and Lu 2012). The southern/northern pole of the dipole pattern is located in southeastern/northeastern China (e.g., Ding et al. 2008, 2009; Zhu et al. 2011; Huang et al. 2011, 2012, 2016; Sun and Wang 2015). The tripod pattern has a positive (negative) rainfall center over the YRV, and two negative (positive) rainfall centers over the northern and southern sides of the YRV, respectively (Wei and Zhang 1988; Hsu and Lin 2007; Huang et al. 2012). Jin et al. (2015) analyzed the interannual variations of regional summer rainfall over mainland China and found a quasi-meridional dipole in eastern China, hereafter the eastern China precipitation dipole (ECPD). The southern center of this pattern is located over the YRV (25°–35°N, 110°–122.5°E). To the northwest of this, over northern China or the Hetao region (NC; 35°–42°N, 105°–115°E), rainfall variations are out of phase with the YRV region (e.g., Wei and Zhang 1988; Weng et al. 1999; Hsu and Liu 2003; Hsu and Lin 2007; Huang et al. 2012; Jin et al. 2015). The poles of ECPD in YRV and NC resemble the northern and central poles of the aforementioned tripod pattern. Both regions are of tremendous socioeconomic and ecological importance. The YRV region is one of China’s most fertile, populous, and affluent zones. On the other hand, the ecological environment of the much drier NC is greatly sensitive to climate variations and change. The main purpose of this paper is to analyze in detail the meridional out-of-phase relation between rainfall over the YRV and NC and its interdecadal variations. Here we find evidence of an interdecadal change occurring around the mid-1990s. Section 4 compares changes in the local circulation before and after the mid-1990s. The associations with different types of ENSO events are explored in section 5. The conclusions and discussion are given in the last section.

2. Data

We obtained rainfall observations of 753 stations in China at daily resolution for the period 1979–2009 from the National Climate Center of China Meteorological Administration. From this we chose 31 and 25 stations (Fig. 1) located in the YRV and NC, respectively, for our analysis on the basis of Jin et al. (2015). The YRV and NC regions are defined by the spatial distribution of the higher loading of the 3rd and 18th components of rotated empirical orthogonal function decompositions of summer precipitation in China (Jin et al. 2015). For the period 2010–14, daily precipitation data are from the Meteorological Information Comprehensive Analysis and Process System, version 3.0 (MICAPS 3.0), which share the same observational gauges with those obtained from the China Meteorological Administration. Outgoing longwave radiation (OLR) data from 1979 to 2013 were obtained from the National Oceanic and Atmospheric Administration (NOAA) (Liebmann and Smith 1996). The rest of the datasets listed below are all from 1979 to 2014. Monthly mean NOAA Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997), which is available on a 2.5° × 2.5° grid are also used to complement our analyses. Environmental variables and sea surface temperature (SST) data used from the

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ECMWF interim reanalysis (ERA-Interim; Dee et al. 2011) and the Met Office Hadley Centre (Rayner et al. 2003), respectively.

To focus on interannual variations, we have high-pass-filtered periodicities of 9 years and longer using a 9-yr running mean filter. This filter method has been extensively used in climate analysis to exclude the decadal signal (e.g., Hsu and Lin 2007; Xie et al. 2009; Ye and Lu 2012; Huo et al. 2015). The definitions of EAP and PJ indices are from Huang (2004) and Wakabayashi and Kawamura (2004). The winter, spring, and summer here are defined as the time periods December–February (DJF), March–May (MAM), and June–August (JJA), respectively.

3. The out-of-phase relationship of summer rainfall

Figure 1a presents the spatial distribution of the climatology of mean JJA rainfall over the period 1979–2014 for China. There is a marked quasi-meridional gradient wherein rainfall amount decreases in amplitude from southeastern to northwestern China. An elongated maximum near the southern coast of China receives the most abundant rainfall with values exceeding 800 mm. Two secondary rainfall maxima with values exceeding 700 mm are located over the YRV, and to its west in the upper reaches of the Yangtze River. Markedly low rainfall is found over western Inner Mongolia, with minimum values around 100 mm. Thus, it can be seen that the spatial distribution of summer rainfall in China—especially in eastern China—is asymmetrical, which is consistent with previous studies (Zou et al. 2005; Zhai et al. 2010; Jin et al. 2015).

The distribution of the interannual standard deviation of JJA rainfall (Fig. 1b) resembles that of the climatological mean of summer precipitation (Fig. 1a). Maximum values greater than 250 mm are located over the YRV, whereas the minimum values less than 50 mm are over NC. It indicates that the YRV, with more summer precipitation and larger interannual variations, is more vulnerable to extreme flood events compared to NC (Jin et al. 2015).

The summer rainfall of the YRV is significantly negatively correlated with that over NC, with a correlation coefficient of $-0.62$ for the period 1979–2014. For the convenience of further analysis, we use the normalized difference between the YRV and NC as an index representing the inverse relation, and hereafter refer it to as the eastern China precipitation dipole index (ECPDI). We show the spatial structure of this inverse relation by correlating ECPDI with station rainfall over eastern China (Fig. 2a) and also with CMAP rainfall (Fig. 2b). The strong out-of-phase relation between the YRV and NC regions exhibit a meridional wavelike pattern resembling the meridional distribution of East Asian summer precipitation related to the EAP/PJ teleconnection (e.g., Huang 2004; Kosaka and Nakamura 2011), manifesting its tight relation with the EAP/PJ pattern. Although the two poles of ECPD resemble the northern and central poles of
the tripole rainfall pattern in eastern China, there is no significant correlation between the ECPD and the southern pole of the tripole pattern. It implies that the summer rainfall in the YRV and NC are closely related, but neither of which significantly correlates with the southern China rainfall in recent decades.

Although the inverse relation of summer rainfall over the YRV and NC are closely related, but neither of which significantly correlates with the southern China rainfall in recent decades.

Although the inverse relation of summer rainfall over the YRV with that over NC is robust for the entire 36-yr period (1979–2014), there are significant decadal variations in the two variables and their connectivity. For instance, there is a clear diminution of rainfall variation over both regions after the 1990s (Fig. 2c). In addition, we can also discern a weakening of the inverse relation around the same time (Fig. 2c). To depict the changes in the inverse relation over time, we have conducted a sliding correlation analysis using an 11-yr moving window (Fig. 3a). From this we see that the inverse correlation is significant at the 95% confidence level before the mid-1990s, with absolute correlation values of greater than 0.9. There is a sharp reduction of correlations after that continuing up to late 2009 or so. During this period the inverse relation gradually loses significance and becomes even slightly positive around late 2000. To further understand the reasons behind the decadal variations of the inverse relation we subdivide the period into two epochs from 1979 to 1995 and from 1996 to 2009 for detailed analysis.

The correlations between ECPDI and summer rainfall over eastern China are shown in Figs. 3b and 3c for the two epochs. For 1979–95, the correlations are very similar to
Fig. 3. (a) The 11-yr sliding correlation between $P_{YRV}$ and $P_{NC}$, the dashed line indicates a statistical confidence at the 95% confidence level; and correlations of ECPDI with station-observed rainfall for (b) 1979–95 and (c) 1996–2009. Correlation coefficients exceeding the 95% confidence level are stippled in (b) and (c).

Figs. 2b and 2c with significant positive correlation coefficients located in the YRV and significant negative ones in NC. However, for 1996–2009, all correlations are not significant despite positive values over the YRV and fewer negative ones over NC (Fig. 3c). The correlation coefficient between the YRV precipitation $P_{YRV}$ and NC precipitation $P_{NC}$ is $0.86$ for 1979–95 and $0.16$ for 1996–2009.

### 4. Related circulation background

To identify dynamic features of local circulation, we use OLR data as a proxy for convective activity. Further, we use NCEP–NCAR reanalyses data to calculate the atmospheric moisture flux and its convergence/divergence at both lower and upper levels for the two epochs: 1979–95 and 1996–2009.

Figure 4 shows the correlations between ECPDI and OLR. In the epoch 1979–95, three elongated zones of significant correlation extending from the western North Pacific (WNP) to northern China (Fig. 4a) form a meridional positive–negative–positive tripolar pattern. The negative correlations in the YRV–southern Korean Peninsula–Japan and the positive ones in NC show that the out-of-phase relationship of summer precipitation between the YRV and NC (Fig. 3b) is part of the larger EAP/PJ pattern (Kosaka and Nakamura 2006). It is also notable that remarkable positive correlations are located over the WNP (Fig. 4a), the source region for the EAP/PJ pattern (Huang and Li 1987; Nitta 1987). We further confirm the tight relation between ECPD and EAP/PJ by correlating their indices (Table 1). It is seen that the ECPDI significantly correlates with...
the EAP index for a 36-yr period, and for 1979–95 (Table 1).

However, during 1996–2009, the negative correlation coefficients between ECPDI and OLR cover most of eastern China, more or less in a uniform manner. Two negative centers are situated over southwestern and northeastern China, respectively. The correlation coefficients over the YRV and those over NC do not appear opposite in signs (Fig. 4b). Further, the correlations of ECPDI with EAP/PJ indices are very weak for 1996–2009 (Table 1). Thus, we can see that not only does the out-of-phase relation between the two regions vanish, but also the connection of eastern China rainfall with the EAP/PJ pattern disappears.

A further analysis on moisture transport is presented in Fig. 5. During 1979–95, water vapor flux vectors originating from the central WNP carry moisture to southeastern China. To its north, there is a cyclonic circulation anomaly transporting moisture from the Okhotsk Sea (Fig. 5a). This distribution of moisture flux related to ECPDI is similar to that of the EAP/PJ pattern (Kosaka and Nakamura 2006). The anomalous moisture convergence over the YRV and anomalous divergence over NC tend to bring above-normal rainfall to the YRV but below-normal rainfall to NC, which contributes to the out-of-phase relationship of summer precipitation between the YRV and NC. During 1996–2009, however, the moisture flux transport lacks the meridional wave train pattern during the previous epoch. Instead a much expanded meridional-scale structure is associated with large-scale moisture transport from the tropical central Pacific to southern China by way of the South China Sea (SCS) (Fig. 5b). During this epoch, the anomalous convergence zone is centered along the Yangtze River and brings abundant rainfall only over the Yangtze River basin.

5. Changing connectivity to ENSO variability

Figure 6a shows the 11-yr moving correlation of ECPDI with Pacific SST anomalies averaged over the 5°S–5°N latitudinal band. Note the evident inter-decadal shift in the spatial pattern of the correlation coefficients appearing around the mid-1990s. Before the mid-1990s, the correlation coefficients are negative to the west of 164°E, while those east of 164°E are positive (Fig. 6a). The notable positive correlations are mainly located over the eastern equatorial Pacific (Fig. 6a).

It is remarkable that the correlation pattern between ECPDI and equatorial Pacific SST anomalies flips in sign between the two epochs. After the mid-1990s, ECPDI is positively (negatively) correlated with SST to the west (east) of the 164°E region (Fig. 6a). Further, significant negative correlations are confined to the central equatorial Pacific between 170°E and 170°W (Fig. 6a).

Table 1. Correlation coefficients between ECPDI and EAP/PJ index (Huang 2004; Wakabayashi and Kawamura 2004). The critical values at 0.01, 0.05, and 0.1 significance levels are 0.42, 0.33, and 0.28 for 1979–2014; 0.61, 0.48, and 0.41 for 1979–95; and 0.66, 0.53, and 0.46 for 1996–2009, respectively. Values given in boldface are significant at the 99% confidence level.

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The change in the spatial structure of the rainfall–SST correlation is further analyzed by correlating ECPDI with the Niño-3 (5°S–5°N, 150°–90°W) and Niño-4 (5°S–5°N, 160°E–150°W) indices (Table 2). It is now clearly seen that in the first epoch the ECPDI significantly correlates only with the Niño-3 index, with a coefficient of 0.58. In contrast, during the second epoch, ECPDI loses the correlation with Niño-3, but gains significant correlation with the Niño-4 index.

To further understand the relation between ECPD and the evolution of tropical Pacific SST anomalies, we analyze the lead/lag correlation of the SST anomalies during the preceding winter, spring, and simultaneous summer with the ECPDI in the two epochs (Figs. 6b–g).

**Fig. 5.** Regressed fields of water vapor flux integrated from 1000 to 300 hPa (kg m⁻² s⁻¹; vectors) and its divergence (kg m⁻² s⁻¹; shading) in summertime with respect to the ECPDI for (a) 1979–95 and (b) 1996–2009. Regression coefficients (water vapor fluxes) that are significant at the 95% confidence level are stippled (shown).

**Fig. 6.** (a) The 11-yr sliding correlation between the ECPDI and the equatorial Pacific SST averaged over 5°S–5°N during the summer (JJA) for 1979–2014. The correlations between ECPDI and SST in (b),(c) preceding winter [DJF(−1)]; (d),(e) spring [MAM(0)]; and (f),(g) current summer [JJA(0)] for the periods (left) 1979–95 and (right) 1996–2009. Correlation coefficients significant at the 95% confidence level are cross hatched.
For the first epoch, the positive correlation coefficients in the central and eastern tropical Pacific gradually increase from the previous winter to summer (Figs. 6b,d,f), indicating a developing canonical El Niño episode. That is, the positive/negative phase of ECPD usually accompanied the developing El Niño/La Niña event during 1979–95. This result is consistent with a previous study (e.g., Huang and Wu 1989) that suggested the YRV receives abundant rainfall during the ENSO developing stage. For the latter epoch, no significant correlation is found in the tropical Pacific during the preceding winter and spring (Figs. 6c,e). In summer, however, the noticeable negative correlation values occur near the date line (Figs. 6a,g), suggesting the developing central Pacific La Niña condition. Thus, the central Pacific ENSO events might be responsible for the disappearance of ECPD.

The regression of velocity potential anomalies onto ECPDI are shown in Fig. 7 for the two epochs separately. During 1979–95, responding to the warm SST anomaly over the equatorial eastern Pacific (Fig. 6f), positive (negative) velocity potential anomalies accompanied by anomalous convergence (divergence) at the lower (upper) troposphere (Figs. 7a,c) are observed over the central and eastern equatorial Pacific. These are accompanied by a low-level westerly anomaly over the central tropical Pacific (Fig. 8a) caused by El Niño events (Wang et al. 2000). Additionally, anomalous descending motion around the Philippines (Fig. 9a) is associated with suppressed convection over the vicinity region of Philippines forcing the negative phase of the EAP/PJ teleconnection (Fig. 4a) (Guan and Yamagata 2003; Hsu and Lin 2007). The negative phase of EAP/PJ produces a wave train with an anomalous anticyclonic circulation around Taiwan and an anomalous cyclonic circulation over Japan at the lower troposphere (Fig. 8a). This pattern of circulation anomaly is largely responsible for the anomalous moisture convergence/divergence in YRV/NC (Fig. 5a), leading to the out-of-phase relationship of summer rainfall between the YRV and NC.

During 1996–2009, when a negative SST anomaly appears in the central tropical Pacific and positive SST anomalies in the western and eastern Pacific (Fig. 6c), the thermal gradient can induce surface pressure gradient among the central, western, and eastern Pacific (Ashok et al. 2007). The aforementioned configuration of the SST anomaly pattern is in favor of anomalous divergence at 850 hPa (Fig. 7b), anomalous convergence at 200 hPa (Fig. 7d), and descending motion (Fig. 9b) over the central equatorial Pacific. In addition, the anomalous low-level convergence (Fig. 7b), high-level divergence (Fig. 7d), and ascending motion (Figs. 9b,d) are located over the Maritime Continent region.
other words, the SST anomaly in the tropical Pacific (Fig. 6g) can induce a pair of overturning anomalous vertical circulations with a downward branch over the central Pacific and an upward branch over the Maritime Continent (Fig. 9b), which is consistent with the two anomalous Walker circulation cells over the tropical Pacific Ocean during the central tropical Pacific warming episodes (Ashok et al. 2007; Feng et al. 2010). The anomalous ascending motion over the western tropical Pacific can alter the meridional vertical circulation in East Asia, and the anomalous descent is located at approximately 20°N and produces a secondary circulation between 20° and 30°N in the mid- to lower troposphere (Fig. 9d). The secondary circulation is conducive to the anomalous ascent (Fig. 9d) and moisture convergence (Fig. 5b) over the YRV. But there are no significant vertical motion and moisture transport anomalies in NC. On the other hand, due to the Matsuno–Gill-type Rossby wave atmospheric response to the cooling in the central tropical Pacific (Fig. 6c), a zonally elongated anticyclonic circulation anomaly generates in the lower troposphere, occupying almost the entire tropical Pacific (Fig. 8b). To its northern flank, however, no counterpart anomalous cyclonic circulation is found, which indicates that there is no significant connection between the SST anomaly in the central tropical Pacific and the EAP/PJ teleconnection. It suggests that the out-of-phase relationship of summer precipitation between the YRV and NC becomes weakened in this period.

6. Conclusions and discussion

This study investigated the interdecadal change of the eastern China precipitation dipole (ECPD), a prominent feature of summer monsoon variability over eastern China. We show that a pronounced weakening of both the rainfall variability over eastern China as well as the precipitation dipole structure has occurred around the mid-1990s. Changes in large atmospheric circulation and sea surface temperature variations

![Graphs showing streamfunction and wind vectors with respect to the ECPDI for two time periods.](image)
associated with this interdecadal change were analyzed by considering two epochs: one during 1979–95 and the other during 1996–2009.

Our study confirms the results of previous studies that the ECPD is a manifestation of the EAP/PJ pattern. However, here we show that this connectivity is only valid during the 1979–95 period. The connectivity between eastern China precipitation and the EAP/PJ pattern completely disappeared during 1996–2009. Instead during the second epoch eastern China precipitation varied under the direct influence of meridionally expanded circulation anomalies over the western Pacific. Intriguingly correlation of the precipitation dipole with Pacific sea surface temperature flips in sign during the second epoch, with enhanced precipitation over southeastern China associated with La Niña–like variability with a center closer to the central Pacific. It is interesting that the meridionally broader atmospheric circulation features noted during the second epoch bear close similarity to the altered structure of ENSO teleconnections (Kug et al. 2009; Chen and Tam 2010; Feng et al. 2011) associated with central Pacific/Modoki flavors of ENSO. Therefore, it is likely that the weakening of ECPD may be a direct consequence of the predominance of Modoki-type ENSO events during the second epoch.

In this study we have focused only on ENSO in explaining interdecadal variations in the ECPD. However, other potential factors such as the Pacific decadal oscillation (PDO; Mantua et al. 1997) and North Atlantic Oscillation (NAO; Hurrell 1995) need to be investigated. Previous studies have suggested that the PDO and NAO have influences on the interdecadal shift of the East Asian climate (e.g., L. Wang et al. 2008; Wu et al. 2009; Jin et al. 2013; Feng et al. 2014). Could these factors affect the ECPD? In addition, we have noticed that during the second epoch the ECPDI correlates also with the north tropical Atlantic (NTA) SST anomaly (Fig. 6c). Previous research has revealed mechanisms through which the NTA SST can exert influence on the climate of the WNP and East Asia (Ham et al. 2013; Huo et al. 2015). It is of interest to see if the NTA SST anomaly impacts the ECPD through such teleconnections. Finally, it is interesting that the ECPD has gradually recovered since the late 2000s. Will the ECPD continue to strengthen hereafter? If so, what is the reason for this? All these questions are the goals of additional studies.

It is also noted that 1994 is identified as a Modoki/central Pacific El Niño year (Ashok et al. 2007; Kug et al. 2009), but we find the ECPD phenomenon is obvious in this year. Additionally, 1994 is a well-known Indian Ocean dipole year (Saji et al. 1999; Guan and Yamagata 2003; Saji and Yamagata 2003). It has been found that the Indian Ocean dipole can also exert influence on the eastern China summer precipitation (e.g., Guan and Yamagata 2003). There is a need for further research into whether the Modoki/central Pacific El Niño and Indian Ocean dipole have combined effects on the ECPD.
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