Clim Dyn (2015) 44:2979–2987
DOI 10.1007/s00382-015-2605-2

# Direct impact of El Niño on East Asian summer precipitation in the observation

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Received: 19 January 2014 / Accepted: 10 April 2015 / Published online: 23 April 2015


**Abstract** This study investigates the direct impact of El Niño in the tropical Pacific on the East Asian summer precipitation. Generalized equilibrium feedback assessment is used to isolate this direct impact from interrelated ocean forcings in the observations. Results indicate that the El Niño can directly influence the summer precipitation in East China significantly. The precipitation response presents a tri-pole pattern, with anomalous wet in the Southeast and the Northeast China and anomalous dry in the northern China. Amplitude of the precipitation response is around 20 % of the total precipitation for 1 °C El Niño forcing in most area of the East China, with maximal response up to 30 %/°C. The tri-pole precipitation response is attributed to an El Niño-induced cyclonic anomaly in the Northeast Asia and an anticyclonic anomaly in the western North Pacific (WNP). The anomalous cyclone deepens the East Asian trough southwestward, favoring an air ascending in front of the trough in the Southeast and the Northeast China, and an air descending at the rear of the trough in the northern China. The anomalous anticyclone in the WNP strengthens the WNP Subtropical High northeastward, providing adequate water vapor to the Southeast China. The anomalous cyclone and anomalous anticyclone work together to generate the tri-pole precipitation response pattern in the East China. Further investigation suggests that these two key anomalous circulations are part of a northwestward propagating Rossby wave, which is excited by the El Niño warming-induced convection over the subtropical west-central Pacific. This study can serve as a reference for the prediction of the East Asian precipitation in both the developing and decaying summer of El Niño.

**Keywords** Direct impact · El Niño · East Asian summer precipitation · GEFA

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## 1 Introduction

El Niño[1] is the most prominent climate variability in the tropical Pacific. This variability is characterized by an anomalous sea surface temperature (SST) in the eastern-central equatorial Pacific, with a warming (cooling) corresponding to the El Niño (La Niña). Previous studies have shown that El Niño can affect East Asian summer monsoon rainfall via atmospheric teleconnections (e.g. Fu 1987; Huang and Wu 1989; Liu and Ding 1992; Zhang et al. 1996). Due to its phase-locking nature, El Niño SST anomaly tends to peak in boreal winter. In boreal summer, therefore, the El Niño SST anomaly is usually undergoing its developing or decaying stage in the tropical Pacific. Thus, early studies of the impact of El Niño on the summer monsoon in the East Asia have discussed the impact mostly for the developing stage or decaying stage of the El Niño.

1 In this paper, unless otherwise specified, El Nino refers to both the warm El Nino and the cold La Nina.

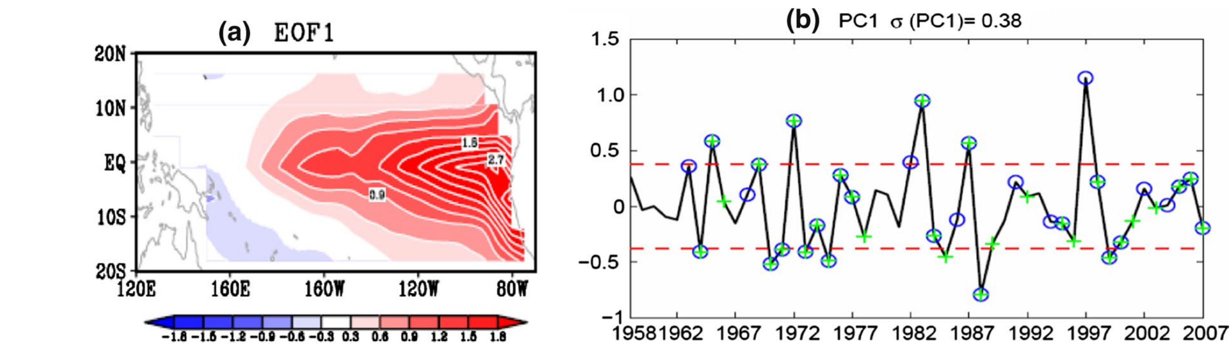


**Fig. 1** **a** The first EOF pattern of summer SST in the tropical Pacific. *Solid* (*dashed*) *line* for positive (*negative*) value with contour interval 0.3 °C. The magnitude of the spatial pattern is normalized with a standard deviation of 1 °C. **b** The corresponding PC with the standard deviation 0.38 (marked by the *red dash lines*) and 56 % of explained variance. The *blue circle* denotes the developing summer of the El Niño/La Niña year, and the *green* cross indicates the next year of the developing summer. The El Niño (La Niña) year is selected in terms of Nino3.4 SST anomalies greater (less) than 0.5 °C for a minimum of five consecutive over-lapping seasons

Huang and Wu (1989) reported a tri-pole summer rainfall response in the developing stage of El Niño, with flooding over the Yangtze-Huaihe River Valley and drought in the southern and northern China. This tri-pole rainfall anomaly pattern is reversed during the decaying stage in the ensuing summer after the peak of El Niño. They proposed that in the developing (decaying) stage of El Niño, the cold (warm) SST in the western tropical Pacific suppresses (actives) the atmospheric convection over the South Asia, resulting in a southward (northward) shift of the Subtropical High in the western North Pacific (WNP) and, in turn, an abundant (deficient) rainfall in the Yangtze-Huaihe River Valley. This view of the direct impact of El Niño on the East Asia summer monsoon has been challenged by recent studies, especially for the decaying stage (e.g. Wang et al. 2000; Wang and Zhang 2002; Yang et al. 2007; Wu et al. 2012). These latter studies suggest that, during the decaying stage of El Niño, the climate impact of El Niño is produced indirectly by the bridging effect from other ocean regions. Wang and Zhang (2002) proposed that the anomalous Philippine Sea anticyclone (PSAC) from mature El Niño to the ensuing summer plays an important role in conveying the impacts of the El Niño to the East Asian summer monsoon. The long persistence of the PSAC is due to the local air–sea interaction between the cold SST anomaly and the descending branch of the atmospheric Rossby waves in the WNP (after a preceding warm El Niño). Yang et al. (2007) argued that tropical Pacific El Niño induces a basin-wide warming over the tropical Indian Ocean, which prolongs the El Niño influence into the following summer. This warming over the Indian Ocean impacts the East Asian summer monsoon either through the atmospheric Rossby wave train over the Asian continent (Yang et al. 2010), or through the atmospheric Kelvin wave into the western subtropical North Pacific to affect the PSAC and the subsequent atmospheric teleconnection towards the East Asia (Xie et al. 2009). Wu et al. (2012) proposed that the winter El Niño can also affect the rainfall of the following summer in the East Asia via the atmospheric teleconnection through the North Atlantic.

To better understand the physical mechanism, we classify the climate impacts of El Niño in two types: the direct El Niño impact and indirect El Niño impact. The direct impact is caused by atmospheric teleconnections forced directly by the SST anomaly in the equatorial Pacific—the source region of El Niño, while the indirect impact is caused by atmospheric teleconnections forced indirectly by a SST anomaly outside the equatorial Pacific. In this terminology, our discussions above suggest that early studies tend to think of the El Niño impact on the East Asian summer monsoon in terms of the direct impact, while more recent studies tend to highlight the indirect impact in the decaying summer. Physically, the response of East Asian summer monsoon to El Niño depends only on the SST anomaly occurring in the same summer, because of the rapid adjustment of the atmosphere to the SST forcing (less than a month; e.g. Liu and Alexander 2007). In boreal summer, the El Niño SST anomaly in the central-eastern tropical Pacific is a dominant mode of SST variability in the tropical Pacific, as the leading EOF of the summer tropical Pacific SST anomaly (Fig. 1a). In spite of its peaking time in winter, the El Niño SST anomaly can still be present in the tropical Pacific in summer, either the developing summer, or the decaying summer of the El Niño (Fig. 1b), such as the long persistent La Niña in 1984 up to the next winter (decaying summer), the rapid transition of El Niño-La Niña-El Niño from 1963 to 1965 (developing summer), and the continuation of La Niña from 1998 to 2000 (developing/decaying summer). These summer El Niño SST anomalies would exert its contribution to the East Asian summer precipitation both in the developing and the decaying stages of the El Niño. However, this direct impact is difficult to be

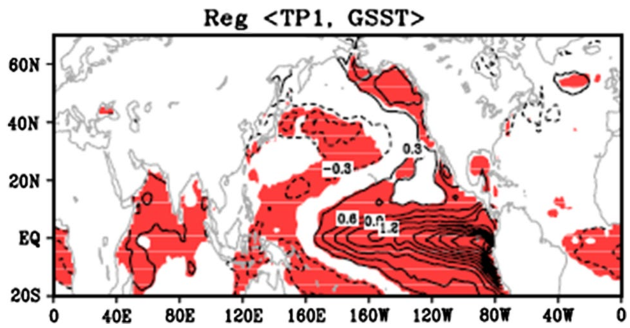


**Fig. 2** Regression of the summer SST anomaly in the tropical-northern hemisphere onto the summer El Niño time series (Fig. 1b). *Solid* (*dashed*) *line* for positive (negative) value with contour interval 0.3. The *shaded* indicates the 90 % significance by two-tail student $t$ test

isolated in the observations, because of the potential interference of atmospheric responses to summer SST anomalies in other oceans. Regression map of the global SST anomaly on the summer El Niño (Fig. 2) suggests that there are significant SST anomalies coexisiting outside the tropical Pacific, notably in the North Pacific,the tropical Indian Ocean and the tropical Atlantic. These SST anomalies in, say, the WNP and tropical India Ocean can further force the atmosphere to interfere with the direct El Niño impact on the East Asia summer precipitation. Therefore, the key to assess the direct impact of the El Niño on East Asian summer precipitation in the observation is to isolate the direct impact from the impacts from other oceanic regions.

It has been a challenge to isolate a specific ocean influence from the interrelated ocean forcings in observations. One common practice has been to filter out the effect of the dominant external forcing using regression and then to assess the climate impact in the residual variability. However, this approach is effective only if there is a single dominant external forcing and the forcing is known a priori. With complex interactions among the climate impact from multiple oceanic forcings, the simple residual approach is difficult to apply. Liu et al. (2008) proposed a multivariate statistic method-the GEFA, to identify the impact of a specific oceanic forcing. Without any pre-filtering, GEFA can separate the impacts of different SST forcings. This method has been applied successfully to distinguish the atmospheric geopotential height response to various global SST variability modes (Wen et al. 2010), and to evaluate the contribution of the regional SST variability modes on the US precipitation variability (Zhong et al. 2011).

In this paper, we use GEFA to isolate the direct El Niño impact on the summer precipitation in the East Asia, and further investigate its physical mechanism. We find that El Niño has a significant direct impact on the summer precipitation in eastern China with a tri-pole response pattern. This tri-pole precipitation anomaly is mainly attributed to an El Niño-induced cyclonic anomaly in the East Asia and an anticyclonic anomaly in the WNP. This direct impact largely corresponds to, and offers a physical explanation to, the El Niño impact in the developing stage (Huang and Wu 1989). The rest of the paper is arranged as follows. Section 2 describes the data and method. Section 3 studies the direct impact of the El Niño on the East Asian summer precipitation. Section 4 summarizes the main conclusions.

## 2 Data and method

The precipitation data is the monthly mean precipitation data from 160 stations covering the mainland China from 1958 to 2007 from the Chinese Meteorological Data Center. Given the sparse observation in the west, the analysis will be confined to eastern China, where 140 stations are evenly distributed east of 100E. To reduce noise, the rainfall data is then reconstructed using the first 10 empirical orthogonal function (EOF) modes, which retains about 60 percent of the total variance. Major conclusions remain unchanged when the original rainfall data are used. Precipitation variability is represented by the precipitation percentage, which is defined as the ratio of the precipitation anomaly to its seasonal climatological mean at each station. Other data used include the monthly mean SST, air temperature at 850 hPa, the omega at 500 hPa, wind at 850 and 500 hPa, and geopotential height at 850, 500 and 200 hPa from 1958 to 2007 from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalney et al. 1996). Monthly anomalies are calculated after removing their seasonal cycles and detrended with a third-order polynomial filter. Low resolution data with 7.5-degree longitude by 7.5-degree latitude gird is used for the global atmosphere response analysis, while the original 2.5-degree longitude by 2.5-degree latitude grid is used for the East Asia analysis.

GEFA is used to isolate the direct impact of the El Niño on the East Asian summer precipitation. Details of GEFA has been discussed in Liu et al. (2008) and Liu and Wen (2008). Here, for convenience of the readers, we only describe it briefly. Given the atmospheric variability $X(t)$ and SST variability $Y(t)$, averaged on the monthly time scale or longer, the atmosphere variability can be decomposed into two parts: one forced by the SST $B \times Y(t)$, and the other by stochastic climate noise $N(t)$ associated with the atmospheric internal variability, such that

$$X(t) = B \times Y(t) + N(t). \tag{2.1}$$

The SST field $Y(t)$ includes multiple ocean forcings. Matrix$B$ is the feedback matrix with $b_{ij}$ representing atmospheric response at the $i$th point$x_i(t)$ to the individual SST forcing at $j$th point $y_j(t)$. On climate time scale, the atmospheric internal variability can be taken as a while noise, and therefore the SST variability of earlier times can't be forced by later stochastic atmospheric variability ($C_{yn}(\tau) = 0$). Using the leading ocean $Y(t-\tau)$ to covary the both side of the Eq. (2.1), we derive the feedback matrix $B$ as

$$B(\tau) = C_{xy}(\tau)C_{yy}^{-1}(\tau), \tag{2.2}$$

where $C_{uv}(\tau) = U(t-\tau) \times V^{'}(t)/T$ represents a lagged covariance matrix ($U$, $V$ is the general variable with superscript prime indicating the transpose), $T$ is the sample size and $\tau$ is SST lead time. The lead time $\tau$ is longer than the atmospheric response time and the persistent time of atmospheric internal variability, here taken as 1 month.

For seasonal GEFA estimation, three consecutive months per year are used to represent a season to increase the sample size and thus reduce the estimation errors (Liu et al. 2012a, b; Wen et al. 2013). Unlike the three-month mean, there's non overlapping month in this way with one month shifting when using the leading ocean to covary the atmospheric Eq. (2.1) for GEFA estimation. For example, for the summer feedback (June–July–August, JJA), the matrix $B_s$ is calculated as the ratio of the covariance of SST in May–June–July (MJJ) (3 months per year) with the atmosphere of June–July–August (JJA) (for each corresponding time it's 1 month shifting behind the SST data) to the auto-covariance of SST in MJJ with the SST of JJA. The SST forcings are derived as the regional EOFs of five non-overlapping subbasins (Wen et al. 2010): the tropical Pacific (TP; 20S–20N, 120E–60W), tropical India (TI; 20S–20N, 35E–120E), tropical Atlantic (TA; 20S–20N, 65W–15E), North Pacific (NP; 20N–60N, 120E–60W) and North Atlantic (NA; 20N–60N, 100W–20E). The first three leading EOF modes of the summer SST in these suboceans are combined into a grand set of EOF modes to represent the set of ocean forcings. Figure 1 shows the first leading EOF mode of summer SST in the tropical Pacific, with the explained variance of 0.56. It resembles the El Niño mode with the loading in the central-eastern equatorial Pacific. In this paper, we will focus on the atmospheric response to this El Niño mode. The significance of $B$ is examined using a Monte Carlo approach, in which the year of atmosphere variable is scrambled randomly 1000 times, while the order of the three consecutive months within a year is retained.

## 3 Impact of El Niño on East Asian summer precipitation

The direct impact of El Niño on the summer precipitation in the East China (Fig. 3) exhibits a tri-pole pattern. It is characterized by a pair of wet regions in the southeastern and northeastern China, sandwiching a dry region in northern China. This response pattern resembles the composite result and the correlation pattern between the summer rainfall in China and the summer El Niño index at the El Niño developing stage, especially in northern China and the Northeast China (Wen et al. 2015; Huang and Wu 1989). That's because that in the developing stage of the El Niño, the El Niño-induced SST anomalies in other oceans haven't been established yet, and the global SST anomaly is dominated by that in the tropical Pacific. Thus, the precipitation response is forced almost solely by the El Niño SST anomaly in the central-eastern Pacific (Specific discussion is given in an accompanying paper (Wen et al. 2015). Figure 3 further shows that the response amplitude in the precipitation percentage is around 20 % in most areas of the East China, passing the 90 % confidence level. This suggests that a summer El Niño SST anomaly of 1 °C lead to a 20 % change of the total summer precipitation over most areas of the East China. The largest increase of precipitation of 30 % occur in the lower reach of the Yangtze River (~32N) and the Songhua River (~47N), while the largest decrease of precipitation of −30 % occurs in the northern China. Overall, Fig. 3 suggests a significant direct impact of the El Niño on the summer precipitation in the East

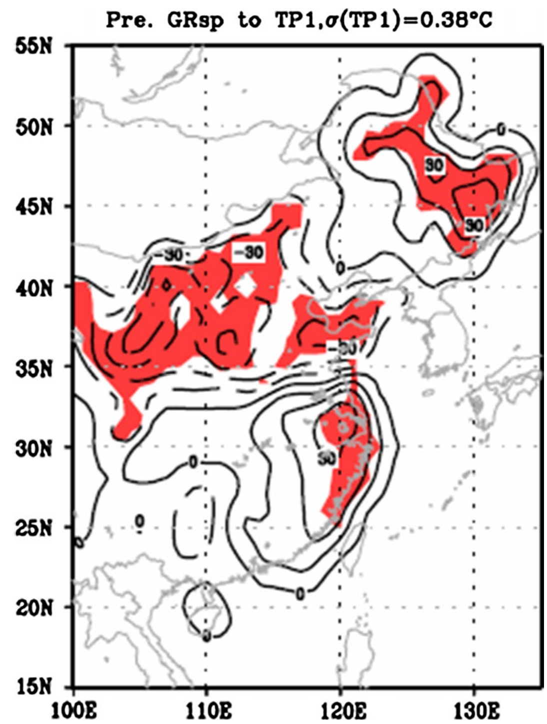


**Fig. 3** GEFA response of the East Asian summer precipitation to summer El Niño forcing in Fig. 1. *Solid* (*dashed*) *line* for positive (negative) values with contour interval 10 %/ °C. The *shaded* indicates the 90 % significance

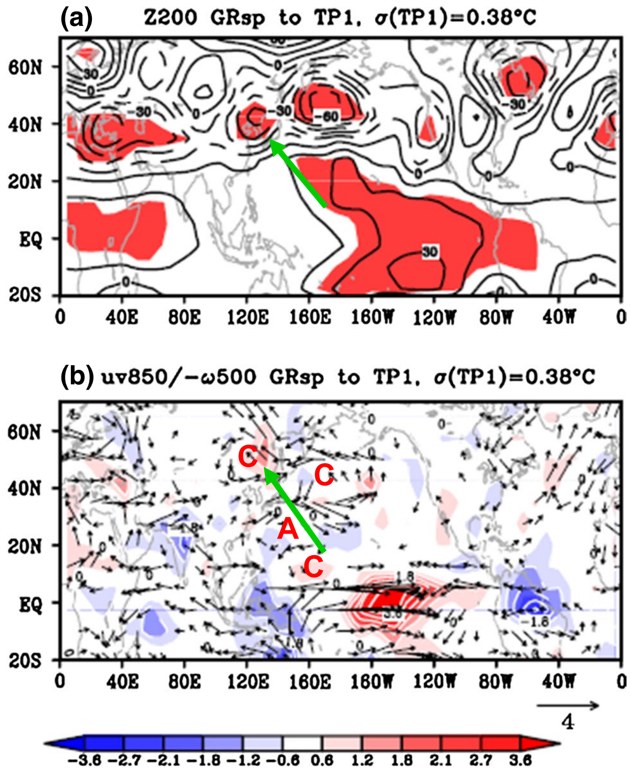


**Fig. 4** **a** GEFA response of the 200 hPa geopotential height in the tropical-northern hemisphere to the summer El Niño forcing. *Solid* (*dashed*) *line* for positive (negative) values with contour interval (CI = 10 m/°C). And, the shaded indicates the 90 % significance. **b** Same as **a**, but for the 850 hPa wind (vector, unit: m/s) and 500 hPa omega with opposite sign ($-\omega$) (*shaded*) response. The small response value of the wind (magnitude <0.5 m/s °C) is omitted in the figure. The *red letters* "C" and "A" mark the centers of cyclone and anticyclone, respectively. The *red* (*blue*) shaded with *solid* (*dash*) *white line* (CI = 0.6 Pa/s °C) indicates the air ascending (descending)

China. In the following, we will further investigate the physical mechanism of the tri-pole response.

### 3.1 Global impact

To better understand the direct impact of El Niño on the East Asian summer precipitation, we first examine the summer global atmospheric response to summer El Niño forcing. Figure 4 shows the three-dimensional structure of the atmospheric response: the low-level wind, mid-level vertical velocity and upper-level geopotential height. The response of the 200 hPa geopotential height in the tropics is characterized by a pair of Rossby waves straddling across the central equatorial Pacific with an amplitude ~20 m/°C, resembling the Matsuno–Gill pattern (Matsuno 1966; Gill 1980). In contrast to the response in winter when the equatorial Rossby wave response is confined in the central-eastern equatorial Pacific (east of ~140E) (Fig. 5 in Liu et al. 2012b), the pair of Rossby waves in the summer extend the two "wings" far westward into the northwestern and southwestern subtropical Pacific. In northern hemisphere extratropics, a wave train response is seen to circulate around the globe with an amplitude about −30 m/°C. In contrast to the winter response of Pacific North America (PNA) teleconnection, in which the response signal is centered in the eastern North Pacific-North America, the dominant signal in summer is centered in the East Asia-Northwest Pacific region, with a maximal low pressure anomaly of −60 m/°C.

At the lower level, the 850 hPa wind response (vectors in Fig. 4b) is dominated by a westerly in the equatorial Pacific converging towards the central-eastern Pacific, with the maximum westerly in the central Pacific. To the east of the maximum westerly anomaly, the westerly converges to the central-eastern Pacific, where the warm SST drives an ascending flow, as seen in 500 hPa omega response field (the red shading in the east-central equatorial Pacific in Fig. 4b). This ascending flow forms the ascending branch of the anomalous Walker circulation. To the west of the westerly maximum is the accompanying descending branch over the marine continent of Sumatra and the North Australia (the blue shading in Fig. 4b). Overall, the characteristic of the atmospheric GEFA response to summer El Niño is in line with our physical understanding for ENSO phenomenon, giving some confidence of the assessed atmospheric response over the globe and in the East Asia as discussed below.

### 3.2 Impact on the East Asia

The atmospheric response in the East Asia is characterized by two major features: an anomalous cyclone in the Northeast Asia from lower to upper atmosphere and an anomalous anticyclone in the WNP in the lower atmosphere. The former favors ascending air and in turn increased precipitation, while the latter supplies additional moisture for precipitation, over the Southeast China. As shown in Fig. 5a, the prominent feature of the 500 hPa geopotential height response is a low pressure anomaly centered at the Northeast Asia, with a magnitude of up to −24 m/°C. This anomalous cyclone in the summer, which has been noticed in some previous studies (Wang and Zhang 2002; Wu et al. 2003), deepens the East Asian trough southwestward up to 27N of South China, as marked by the thick solid trough line in Fig. 5a. Downstream (upstream) of the trough line, anomalous southwesterly (northwesterly) wind prevails (vectors in Fig. 5a). Climatological mean air temperature decreases from the Southwest to the Northeast Asia (red lines in Fig. 5a). Thus, downstream of the trough, the anomalous southwesterly wind transports warm and moist air from the south into the East China Sea and Japan Sea; upstream of the trough, the anomalous northwesterly wind

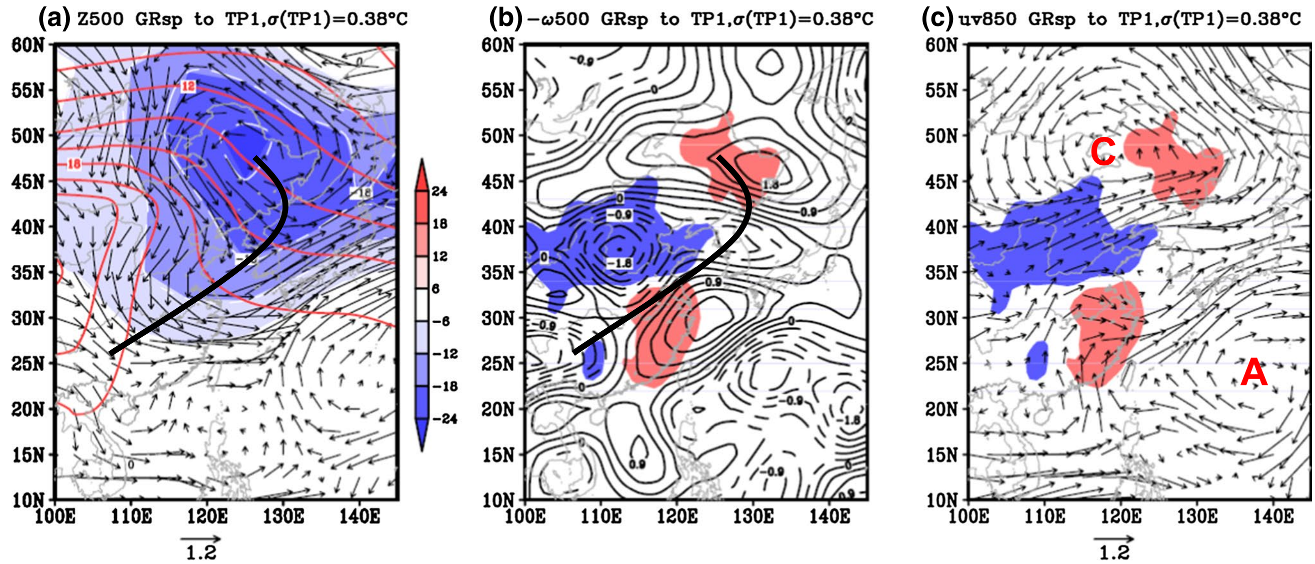


**Fig. 5** Atmospheric GEFA response in East Asia to summer El Niño forcing: **a** 500 hPa geopotential height (*shaded*, the *blue shaded* with white *dash line* for negative value, CI = 6 m/ °C) and wind (vector, unit: m/s) response with superposed climatological mean air temperature at 850 hPa (*red* contours, CI = 2 °C); **b** 500 hPa omega with opposite sign ($-\omega$) response [contours, *solid* (*dash*) *line* for the positive (negative) value, CI = 0.3 Pa/s °C]; **c** 850 hPa wind response (vector, unit: m/s). In **b** and **c**, the *shaded* indicate the precipitation response as Fig. 3 with the *red* (*blue*) for the anomalous precipitation greater (less) than +(−)10 %/ °C. In **a** and **b**, the *thick black solid line* segments mark the 500 hPa strengthened East Asian trough

advects cold and dry air from Lake Baikal into southern China. In term of the $\omega$ equation (Holton 1972; Tao et al. 2012), the warm advection in front of the trough facilitates ascending, and the cold advection at the rear of the trough favors descending. Consistent with the synoptic principle, GEFA response of $-\omega$ at 500 hPa (Fig. 5b) shows a strong ascending area stretching from the Southeast China through the East China Sea to the South of Japan, a strong ascending center in the Northeast China, and a descending center over the northern China, corresponding to the tri-pole rainfall response of the increased precipitation in the Southeast China, and in the Northeast China (red shading in Fig. 5b), and the decreased precipitation in northern China (blue shading in Fig. 5b). These coincidences of the increased rainfall in front of the trough and decreased rainfall at the rear of the trough suggests a critical role of the El Niño-induced anomalous cyclone in the tri-pole precipitation response in the East Asia (Fig. 3).

In addition to the anomalous cyclone, El Niño also generates an anticyclone in the WNP, which is important in enhancing the moisture supply to the Southeast China for increased precipitation. The 850 hPa wind response (Fig. 5c) shows an anticyclonic anomaly in the WNP (15–35N, 115–160E), coherent with the descending motion there (Fig. 5b). This WNP anticyclone is different from the PSAC (10–20N, 120–150E) produced indirectly by El Niño. The PSAC shifts the western Pacific Subtropical High southward and enhances the El Nino-following summer precipitation in the Yangtze valley (Wang and Zhang 2002). In contrast, the El Niño-induced anomalous anticyclone strengthens the WNP Subtropical High northeastward. As a result, moisture can only be transported to the Southeast China by the southwesterly wind along the northwestern flank of the North Pacific Subtropical High, increasing rainfall there (red shading in Fig. 5c). For the Northeast China, plenty of water vapor is from the vicinity of Japan Sea which is advected by the southeasterly anomalies over the southeast tip of the anomalous cyclone in the Northeast Asia. Further investigation about the relative importance of the vapor transport items in the humidity equation (e.g. the wind anomaly-induced moisture flux, the humidity anomaly-induced moisture transport and transient eddy vapor transport) indicates the dynamic contribution induced by the wind anomaly plays a dominant role in the total vapor transport over the East Asian region (Figures are not shown). In summary, in the Southeast and Northeast China, a stronger ascending and enhanced moisture supply from the south intensifies the rainfall, whereas in northern China, enhanced descending of cold and dry air from the north lead to the rainfall reduction. Therefore, the summer tri-pole precipitation anomaly pattern in the East China (Fig. 3) is the result of the joint effects of the anomalous cyclone in the Northeast Asia and the anomalous anticyclone in the WNP induced by the El Niño.

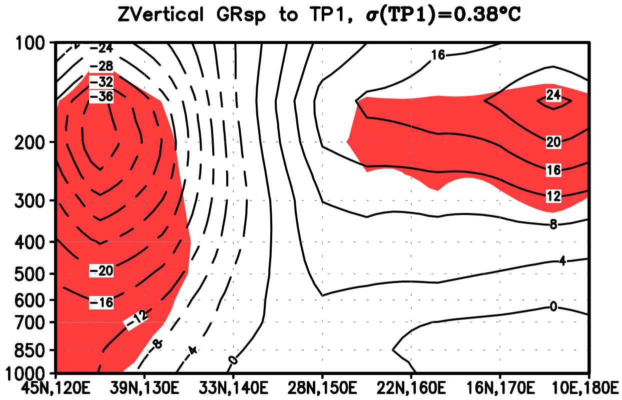


**Fig. 6** The vertical profile GEFA response of the geopotential height to summer El Niño forcing along the line from (45N, 120E) to (10N, 180E) (as indicated by the *green line* in Fig. 4). *Solid* (*dash*) *line* for the positive (*negative*) value with contour interval CI = 4 m/ °C. The *shaded* indicates the 90 % significance

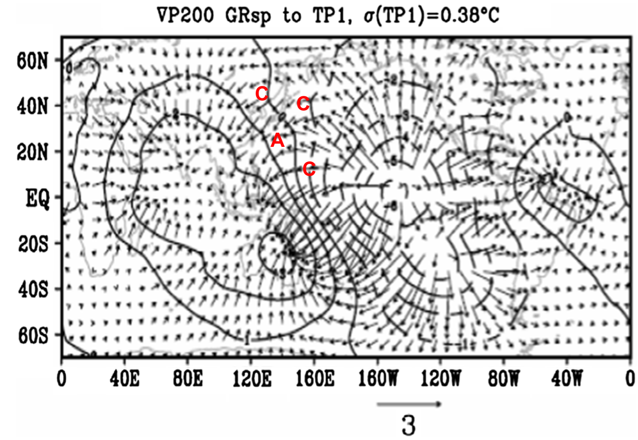


**Fig. 7** The GEFA response of the 200 hPa velocity potential (contours, *solid* (dash) line for positive (negative) value, CI = $10^{-6}$ $m^2/s$ °C) and its divergent wind (vector) to summer El Niño forcing. The red letters "C" and "A" mark the low-level cyclonic and anticyclonic response, as indicated in Fig. 4b

### 3.3 How does El Niño teleconnect to the East Asia?

The final question is how the El Niño from the central-eastern Pacific directly generates the anomalous cyclone in the Northeast Asia and the anomalous anticyclone in the WNP? Figure 4 provides some clues. In the upper level, the equatorial Rossby wave generated by the El Niño warming extends its northern wing northwestward into the west-central subtropical Pacific, with its edge reaching south of Japan (Fig. 4a). The southwesterly wind anomaly at the northwestern tip of the anomalous anticyclone can strengthen southern flank of the upper level westerly jet, favoring genesis of the anomalous cyclone in the East Asia and Northwest Pacific. The corresponding lower level exhibits a wave-train response (denoted as 'C–A–C' in Fig. 4b) that propagates northwestward from the western tropical Pacific ('C' at 160E, 10N) through the WNP ('A' at 140E, 25N) to the Northwest Pacific-East Asia ('C' at 120E, 50N). This wave train response resembles the Pacific-Japan (PJ) pattern (Nitta 1986, 1987). This upper-lower level configuration exhibits a baroclinic response in the western-central tropical Pacific, changing to an equivalent barotropic response in the WNP and the Northwest Pacific-East Asia, as shown in Fig. 6—the vertical profile of the geopotential height response along the wave train (thick green line in Fig. 4). The change of vertical structure here is consistent with previous works on the tropical heating excited Rossby wave to the extratropics (e.g. Lee et al. 2009; Wang et al. 2010). The baroclinic response in the tropical western-central Pacific facilitates the local convection activity as indicated by 500 hPa ascending velocity in Fig. 4b, and further excites the northward propagating Rossby wave, resulting in the anomalous anticyclone in WNP, and the anomalous cyclone in Northeast Asia. In contrast to the winter PNA in the Northeast Pacific-North America, the northward Rossby wave in summer is mostly confined in the Northwest Pacific-East Asia section.

Unlike the original PJ Rossby wave (Nitta 1986, 1987) that is assumed to be excited by SST anomaly and, in turn, convection over the Philippine Sea, the northwestward wave-train due to the direct El Niño impact is forced by the SST warming in the central-eastern tropical Pacific. Figure 7 shows the response of the velocity potential at 200 hPa, which reflects the intensity of the upper level convergence and divergence induced by the El Niño warming. The response in the North Pacific and East Asia is part of the anomalous convergence and divergence induced directly by El Niño warming in the central-eastern tropic Pacific. In Fig. 7, the upper level airflow is mainly controlled by a divergence center at central-eastern tropical Pacific and a convergence center at Sumatra and North Australia, corresponding respectively to the strong ascending and descending in the tropics shown in Fig. 4b. The strong outflow from the central Pacific flows westward and converges to the Sumatra and North Australia; the outflow also flows towards the extratropics in both hemispheres. In the Northern Hemisphere, the outflow from the central Pacific first flows northwestward across the WNP, then turns southwestward towards the East Asia coast, and finally converges over the Sumatra. This is correspondent to the low-level convergence in the tropical western-central Pacific, the divergence in the WNP and convergence in the Northwest Pacific and East Asia. Figure 7 further confirms that those two key circulations, the anomalous anticyclone in WNP and the anomalous cyclone in East Asia, are induced by the El Niño warming in the central-eastern tropical Pacific. This conclusion is consistent with the work of Grimm and Silva (1995), which showed that the PJ wave train can be excited by tropical heating almost anywhere in the Pacific.

## 4 Summary

In this study, we isolate the direct impact of El Niño from the summer tropical Pacific SST on the East Asian summer precipitation using the generalized equilibrium feedback assessment (GEFA). It is shown that El Niño has a significant direct impact on the summer precipitation in the East China. The direct impact on rainfall exibits a tri-pole pattern, wet in the southeastern and northeastern China and dry in northern China. The amplitude of precipitation response to 1 °C El Niño SST anomaly forcing in the tropical Pacific is about 20 % of the total summer precipitation over most areas of the East China, with maximal response up to 30 %/°C.

Two key El Niño-induced anomalous circulations explain the precipitation tri-pole response. The El Niño-induced cyclonic anomaly in the Northeast Asia deepens the East Asian trough southwestward, resulting in the air ascending in front of the trough over the Southeast and Northeast China, and the air descending at rear of the trough over the northern China. Meanwhile, the El Niño-induced anticyclone anomaly in the WNP intensifies the WNP Subtropical High northeastward, providing abundant water vapor to the Southeast China. This anticyclone is quite different from the El Niño-indirect-induced PSAC, which displaces the Subtropical High southward favoring the anomalous summer precipitation in the Yangtze valley (Wang and Zhang 2002). Therefore, in the Southeast and Northeast China, a strong ascending and enhanced moisture supply from the south intensifies the rainfall there, while in the northern China, the enhanced descending of cold and dry air from the north reduces the rainfall.

Further investigation indicates that the above two key circulation anomaly over the WNP-East Asia is achieved through the El Niño warming-induced convection over the subtropical west-central Pacific. The El Niño warming in the eastern-central Pacific generates the anomalous anticyclone in the upper level—the northern lobe of the crossing-equatorial Rossy wave, extending far northwestward into the northwestern subtropical Pacific, and the anomalous cyclone in the western subtropical Pacific at the low level. The baroclinic structure response facilitates the activities of the atmospheric convection over this region, and further excites the northwestward propagating Rossby wave, resulting in the anomalous anticyclone in WNP and the anomalous cyclone in East Asia. Unlike the PJ wave excited by the SST anomaly over the Philippine Sea, here the northwestward wave-train response is originally forced by the SST warming in the central-eastern tropical Pacific. It means that the direct response of the atmosphere to the summer El Niño SST anomaly in the eastern-central tropical Pacific connects the El Niño with the East Asian summer precipitation anomaly. For more specific study about the mechanism, it's beyond the scope of this paper.

This study provides an objective estimation of the El Niño direct impact on the East Asian summer precipitation. As long as the El Niño SST anomaly occurs in summer, it has its own contribution to the summer precipitation in the East Asia, as discussed above. In the developing summer of the El Niño, the El Niño SST anomaly in the tropical Pacific triggers little SST anomalies in other oceans. The summer precipitation anomaly in the East Asia from El Niño is mainly caused by the direct impact through the SST anomaly in the eastern-central tropical Pacific. A more detailed study of this aspect will be given in an accompanying paper (Wen et al. 2015). In the decaying summer of the El Niño, the direct impact of El Niño barely contributes to the ensemble mean summer precipitation anomaly, because of the neutralization of the positive and negative phase. However, for annual seasonal forecast, the direct impact of El Niño on the summer precipitation in the East Asia should still be considered at its decaying stage for two reasons: frequently occurring El Niño SST anomaly either in the warm or cold phase, with the considerable amplitude at the decaying stage (as indicated in Fig. 1b), and the large response of the summer precipitation over the East Asia to the summer El Niño forcing. Therefore, this study can serve as a reference for the summer precipitation prediction in East China both in the developing or decaying summer of the El Niño year.

**Acknowledgments** This work is supported by Chinese NSF41475089, 2012CB955200, and NSF41005048.

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