Weakened Impact of the Indian Early Summer Monsoon on North China Rainfall around the Late 1970s: Role of Basic-State Change

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ABSTRACT

Previous studies have found a link between north China and Indian rainfall during summer, with significantly increased rainfall in north China related to a stronger Indian summer monsoon. This link is weakened after the late 1970s, generally attributed to the reduced magnitude of interannual variability in the Indian summer rainfall. This study reveals a similar change in this rainfall link in early summer after the late 1970s. Related to a heavier Indian early summer rainfall, rainfall in north China enhances significantly before the late 1970s but not thereafter. The change in rainfall teleconnection is caused by the weakened impact on north China rainfall of a midlatitude wave train along the Asian jet in the upper troposphere. After the late 1970s, the portion of the wave train over East Asia displaces eastward, leading to an eastward shift in the associated ascending motion and, subsequently, enhanced rainfall from north China to the Yellow Sea. Moreover, the change in the midlatitude wave train is attributed to the change in the basic state over East Asia (i.e., a northward shift of the East Asian upper-tropospheric westerly jet after the late 1970s). The latter reduces stationary Rossby wavenumber and increases wavelength of the midlatitude wave train, leading to an eastward shift of the wave train over East Asia. Therefore, in this study a mechanism is proposed for the change in early summer, different from the previous mechanism for the entire summer period.

1. Introduction

The two main components of the Asian summer monsoon, the Indian summer monsoon and East Asian summer monsoon, are closely related (Tao and Chen 1987; Lau et al. 2000; Wang et al. 2001; Ding and Liu 2008; Day et al. 2015). Related to the Indian summer monsoon, the most significant interannual rainfall signal appears in north China (Guo and Wang 1988; Liang 1988; Kripalani and Kulkarni 2001; Liu and Ding 2008; Saeed et al. 2011; Lin et al. 2015; Preethi et al. 2017; Wu 2017). Specifically, the stronger Indian summer monsoon corresponds to above-normal rainfall in north China. In addition, increased rainfall over the Yangtze River valley (Zhang 2001) and decreased rainfall in Japan (Kripalani and Kulkarni 2001; Krishnan and Sugi 2001; Wang et al. 2001; Wu 2002; Yun et al. 2014; Preethi et al. 2017; Wu 2017) were also identified during the mei-yu–baiu season in the years with the stronger Indian summer monsoon. Therefore, to understand the relationship between the Indian and East Asian summer monsoons it is vital to reveal not only features of the two monsoon systems, but also the interaction among the Asian summer monsoon.

The Indian summer monsoon is linked to the summer rainfall in north China through a midlatitude wave train along the Asian upper-tropospheric westerly jet (Kripalani et al. 1997; Lu et al. 2002; Wu 2002; Enomoto et al. 2003; Enomoto 2004; Ding and Wang 2005; Greatbatch et al. 2013). The Indian summer rainfall-induced heat release triggers a circumglobal teleconnection (CGT) pattern in midlatitudes of the Northern Hemisphere (Ding and Wang 2005; Lin...
2. Data, indices, and methodology

The precipitation data used in this study include the following: 1) the 160-station observed rainfall data in mainland China provided by the National Climate Center of the China Meteorological Administration; 2) the land precipitation data from the Climatic Research Unit (CRU), University of East Anglia, United Kingdom, with a horizontal resolution of 0.5° × 0.5° (Mitchell and Jones 2005); and 3) the global rainfall derived by the Global Precipitation Climatology Project (GPCP) (Huffman et al. 1997; Adler et al. 2003). The monthly atmospheric data used in this study are derived from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis dataset (Kalnay et al. 1996). All the data used cover the period 1951–2014, except for the GPCP rainfall data during 1979–2014.

Three monthly indices used in this study include the all Indian rainfall index (AIRI), the north China rainfall index (NCRI), and the CGT index (CGTI). The monthly AIRI is derived from the Indian Institute of Tropical Meteorology (Parthasarathy et al. 1994). The NCRI is calculated as 13-station mean rainfall within the north China region (35°–40°N, 110°–120°E) based on the 160-station observed rainfall data. The CGTI is defined as anomalies of geopotential height at 200 hPa (H200) averaged over west-central Asia (35°–40°N, 60°–70°E) in the same manner as Ding and Wang (2005). In this study, the three indices in June are used without specific description. A window of 15 yr is employed to calculate the sliding correlations between the indices. The same analysis is performed with a sliding window of 21 yr and the results are similar. Therefore, only the results with the 15-yr window are presented in this study.

To explore dynamics for the connection of Indian and north China rainfall, the zonal and meridional components of a wave-activity flux for stationary Rossby waves \( \mathbf{W} \) are employed following Takaya and Nakamura (2001), which is defined as

\[
\mathbf{W} = \frac{1}{2|V|} \left[ \frac{1}{4\pi} \left( \frac{\partial^2}{\partial x^2} - \frac{\partial^2}{\partial y^2} \right) + \frac{1}{4\pi} \left( \frac{\partial^2}{\partial y^2} - \frac{\partial^2}{\partial x^2} \right) \right],
\]

where \( \mathbf{V} \) is the velocity field.
where $|\nabla|$ is the magnitude of the horizontal vector wind $(u, v)$ and $\psi$ is the streamfunction; variables with an overbar represent their climatological mean; and variables with subscript and prime notations signify their partial derivatives and anomalies, respectively.

A barotropic model is used to investigate the effect of Indian rainfall in early summer on the CGT. The linearized barotropic vorticity equation is

$$\frac{\partial \zeta'}{\partial t} = -\nabla \cdot (f + \zeta) - \nabla \cdot (f + \zeta) \nabla \cdot \nabla' \zeta' - \kappa \zeta' - \varepsilon \nabla^4 \zeta',$$

where the overbar represents the zonal-mean variables and the prime is the deviation from the zonal-mean state. Here $\nabla \psi$ and $\nabla \psi_x$ are the rotational and divergent wind components, respectively; $f$ is the Coriolis parameter; and $\zeta$ is the relative vorticity. The zonal mean flow is set to their climatological values at 200 hPa in early summer calculated from the NCEP–NCAR reanalysis data. The biharmonic diffusion coefficient $\varepsilon$ is set to $2.34 \times 10^{16}$ m$^4$ s$^{-1}$, and the damping coefficient $\kappa = 5$ day$^{-1}$ is used in this model. The vorticity equation is solved using the spectrum transform technique with a triangular truncation at wavenumber 21.

### 3. Results

#### a. Change in the relationship between the Indian early summer monsoon and north China rainfall after the late 1970s

To investigate the relationship between the India monsoon and north China rainfall in early summer, two indices in June are employed in this study: AIRI and NCRI. The definitions of the two indices can be found in section 2 and their normalized time series are presented in Fig. 1a. The AIRI is significantly correlated with the NCRI for the period 1951–2014 with a correlation coefficient of 0.36, significant at the 99% confidence level.

The India–north China rainfall relationship, however, exhibits a significant change around the late 1970s (Fig. 1b). The positive correlation between the AIRI and NCRI, with a sliding window of 15 yr, is strong in 1960s and 1970s, and becomes weak since 1980. The significant positive correlation occurs during the period of 1962–79. To investigate the decadal change in this early summer relationship between Indian and north China rainfall around the late 1970s, an equal 18-yr period of 1980–97 is chosen. The AIRI is significantly correlated with the NCRI during the first epoch with a correlation coefficient of 0.56, significant at the 95% confidence level, but not during the second epoch with a
correlation coefficient of 0.39. In the following study, the first period is referred to as the pre-1980 epoch and the second period as the post-1980 epoch. The main reason for choosing the end-point year of the second period in the mid-1990s is discussed in section 3b. In addition, the two periods are close to those of 1962–77 and 1978–93 used by Wu and Wang (2002) and Wu (2002) when they discussed the weakened effect of the Indian summer (June–August) monsoon on north China rainfall after the late 1970s. The similar periods facilitate comparison between the result in the present study and that revealed by Wu and Wang (2002) and Wu (2002), which is presented in section 3b.

The weakened rainfall signal in north China related to the Indian early summer rainfall is revealed in rainfall correlation with the AIRI during the pre- and post-1980 epochs (Fig. 2). Related to a positive AIRI, rainfall increases significantly in north China during the first epoch (Fig. 2a) and slightly during the post-1980 epoch (Fig. 2b), based on the 160-station observational rainfall data in mainland China. A similar change is identified using the CRU land rainfall data over East Asia (Figs. 2c,d). In addition, increased rainfall is noticed over the region between the Yellow and the Yangtze Rivers during the second epoch (Figs. 2b,d). This rainfall anomaly to the south of north China, indeed, is a westward extension of a strong, positive rainfall anomaly centered over the Yellow Sea related to the AIRI during the second epoch (Fig. 2e), concurrent with decreased rainfall over the subtropical western North Pacific. The results indicate an eastward shift of the AIRI-related rainfall anomalies over East Asia after the late 1970s from north China (Figs. 2a,c) to the Yellow Sea (Fig. 2e).

The decadal change is also confirmed by correlation of rainfall in the Indian subcontinent with the NCRI during the two epochs (Fig. 3). Rainfall increases significantly in the northern Indian subcontinent during the first epoch (Fig. 3a). However, it only slightly increases along the east coast of the Indian subcontinent during the second epoch (Fig. 3b), which is confirmed by the same analysis using the GPCP rainfall data (figure not shown). In summary, the early summer relationship between the Indian and north China rainfall is weakened after the late 1970s. Related to a heavier Indian rainfall in June, the above-normal rainfall region shifts eastward from

Fig. 2. Correlation coefficient of (a),(b) station rainfall in mainland China, (c),(d) CRU land rainfall, and (e) GPCP rainfall over East Asia with the AIRI in June for the period 1962–79 in (a),(c) and 1980–97 in (b),(d),(e). Regions with blue or orange shading are significant at the 95% confidence level by using the Student’s t test, and contour interval is 0.2, with the contours of ±0.2 and 0 omitted. The blue rectangles depict north China, and the red dots in (a) are the 13 stations within north China.
north China during the pre-1980 epoch to the Yellow Sea during the post-1980 epoch. In section 3b, circulation anomalies responsible for this decadal change are investigated.

b. Circulation anomalies associated with the early summer decadal change

To reveal circulation anomalies associated with the Indian and north China rainfall in early summer, Figs. 4a,b show H200 anomalies in June regressed against the AIRI and NCRI for the period 1951–2014, respectively. Related to the enhanced Indian rainfall, a midlatitude wave train is identified over the Asian continent in the upper troposphere with two anticyclonic anomalies over southwestern Asia and East Asia (Fig. 4a). The wave train is formed because of eastward propagation of the stationary wave-activity flux along the Asian westerly jet. A similar pattern is found associated with the enhanced rainfall in north China (Fig. 4b). This wave train pattern resembles the previously referred “Silk Road” pattern (Lu et al. 2002; Enomoto et al. 2003) or the CGT pattern (Fig. 4c) proposed by Ding and Wang (2005). Here the CGT pattern (Fig. 4c) is obtained by regressing H200 anomalies against the CGTI (Fig. 5a) that is defined in section 2.

To understand the role of CGT in the weakened relationship between the Indian and north China rainfall after the late 1970s, the 15-yr sliding correlation coefficient between the AIRI and NCRI is recalculated after removing the effect of the CGT. The CGT’s effect is removed by subtracting the component linearly regressed upon the CGTI with a sliding window of 15 yr. For example, in 1962, we first remove the CGTI-related linear component from the AIRI and NCRI, respectively, during the 15-yr period 1955–69 with the central year of 1962. Then, after removing the effect of the CGT, the correlation coefficient in 1962 is calculated between AIRI and NCRI. The correlation after removing the effect of the CGT remains around zero (Fig. 5). Compared to the interdecadal change in the AIRI–NCRI relationship before removing the effect of the CGT, the result implies that the weakened relationship between the Indian and north China rainfall after the late 1970s is attributed to the role of the CGT.

The CGT can weaken the early summer relationship between the Indian and north China rainfall after the late 1970s by weakening its connection with either the Indian rainfall or the north China rainfall. We first calculate the correlation coefficients of the CGTI with the AIRI, with a sliding window of 15 yr (Fig. 5b). The correlation remains significant at the 95% confidence level during the whole period of 1951–2014, and is 0.71 during the pre-1980 epoch and 0.66 during the post-1980 epoch. The stable, significant relation of the CGT pattern with the Indian rainfall in early summer is further confirmed by the correlation between the CGTI and rainfall in the Indian subcontinent during the pre-(Fig. 6a) and post-1980 epochs (Fig. 6b), respectively. Rainfall increases significantly in central-northeastern India during both epochs. It can be concluded that the weakened early summer relationship between the Indian and north China rainfall after the late 1970s is not attributed to the change in the Indian rainfall–CGT relation. This conclusion is different from that proposed by Wu and Wang (2002) who highlighted the role of the reduced magnitude of the Indian rainfall anomaly, which weakened its connection to the midlatitude wave train and, subsequently, summer rainfall in north China after the late 1970s during the summer season. Therefore, this result suggests a different mechanism for the decadal change in the early summer relationship.
between the Indian and north China rainfall from that proposed by Wu and Wang (2002).

The possible mechanism for the decadal change in the early summer relationship between the Indian and north China rainfall is that the CGT’s effect on early summer rainfall in north China changes after the late 1970s. As shown in Fig. 5b, the correlation between the CGTI and NCRI is significant before the late 1970s, with a correlation coefficient of 0.75 during the pre-1980 epoch, while it reduces after the late 1970s, with a correlation coefficient of 0.37 during the post-1980 epoch. Moreover, the weakened relationship between the CGT and north China rainfall happens since 1980, concurrent with that between the Indian and north China rainfall. We also examine spatial pattern change between rainfall in East Asia and the CGT (Figs. 6c–e). In the positive phase of the CGT (Fig. 4c), rainfall increases significantly in north China during the first epoch (Fig. 6c) but not during the second epoch (Fig. 6d). The CGT–north China rainfall is weakened after the late 1970s, in good agreement with the weakened correlation between the AIRI and NCRI (Fig. 5b). Also noted is that excessive rainfall related to the CGTI displaces eastward to the Yellow Sea during the second epoch (Fig. 6e), similar to that related to the AIRI (Fig. 2e). In addition, we further regress H200 anomalies upon the NCRI during the two epochs. A CGT pattern is clearly identified during the first epoch while a zonally extended belt of positive H200 over midlatitude Asia is seen during the second epoch (figure not shown), confirming the weakened impact of the CGT on north China rainfall after the late 1970s. In a word, all the results reveal a weakened impact of the CGT on north China rainfall after the late 1970s, which causes the weakened Indian–north China rainfall relationship since the relationship between Indian rainfall and the CGT remains significant during both epochs (Fig. 5b). Additionally, we also notice a significant positive correlation between the CGTI and NCRI and between the CGTI and AIRI after mid-1990s, specifically in late 1990s and early 2000s (Fig. 5b), but a weak positive correlation between the AIRI and NCRI. The results suggest a different physical process, instead of the CGT, may be
responsible for the weak link between the Indian and north China rainfall after the mid-1990s, which is confirmed by a simultaneously negative correlation between the AIRI and NCRI after subtracting the CGT’s impact (Fig. 5b). In this study we focus on the role of CGT in the change of the Indian–north China relationship around the late 1970s; the change after the mid-1990s is beyond the scope of the present study. Therefore, the 18-yr epoch of 1980–97 is used in this study when the CGT’s impact on the north China rainfall is weak in contrast with the 18-yr epoch of 1962–79 when the CGT’s impact on the north China rainfall is strong (Fig. 5b). We have obtained similar results using a different second period with the endpoint year shifting forward and backward slightly in the mid-1990s.

Ding and Wang (2005) pointed out that the CGT can modulate rainfall in East Asia through an CGT-related upper-tropospheric anticyclonic vortex over East Asia. As shown in Fig. 7a, the anticyclonic vortex related to the CGT is clearly identified over East Asia as well as southwestern Asia during both epochs. The anticyclonic vortex remains stable over southwestern Asia. The anticyclonic vortex over East Asia, however, displaces eastward during the second epoch. The eastward shift of the East Asian vortex is more clearly seen in the CGT-related H200 anomalies averaged meridionally over 30°–45°N, with the center shifting eastward approximately 7° longitude from 127.5° to 135°E. The eastward-displaced anticyclonic anomaly is consistent with an eastward shift of the associated ascending motion over East Asia (Figs. 8a,b). A strong ascent related to the CGT is identified over north China during the pre-1980 epoch (Fig. 8a), which is caused by upper-level divergence due to the positive vorticity advection in the upstream half of the positive geopotential height anomaly over East Asia (Holton 1992). The ascent anomaly shifts eastward to the Yellow Sea during the post-1980 epoch (Fig. 8b). In the lower troposphere, related to the CGT, a southerly anomaly prevails over eastern China during the first epoch (Fig. 8c), bringing more moisture to north China and, together with the enhanced ascending motion, favoring rainfall there. During the second epoch, a significant southwesterly anomaly blows from the South China Sea to the Yellow Sea (Fig. 8d), which favors rainfall around the Yellow Sea with the local anomalous ascending motion. Note that the southwesterly anomaly is part of an anticyclone anomaly over the subtropical western North Pacific (Fig. 8d), which is southward tilted from the upper-tropospheric anticyclone anomaly related to the CGT (Fig. 8b). The southward-displaced anticyclone anomaly

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**FIG. 5.** (a) Interannual variations of the CGTI in June for the period 1951–2014. (b) Correlation coefficient between the AIRI and NCRI before (red line with open circles) and after (red dashed line) removing effect of the CGT, between the AIRI and CGTI (blue line with filled circles), and between the CGTI and NCRI (black line) with a sliding window of 15 yr. The effect of the CGT is removed by subtracting linear component regressed upon the CGTI. The black dashed lines depict significance at the 95% confidence level by using the Student’s t test.
in the lower troposphere may be caused by the feedback of suppressed rainfall in the subtropical western North Pacific (Fig. 6e).

Therefore, the eastward shift of the anomalous ascent and lower-tropospheric moisture transport (Fig. 8) leads to the associated eastward shift of rainfall after the late 1970s from north China to the Yellow Sea (Fig. 6). The same eastward shift, related to the AIRI, in the upper-tropospheric anticyclone anomaly over East Asia from west to east of the Korea Peninsula is also identified.

Fig. 6. As in Fig. 2, but for correlation coefficients of (a)–(d) CRU and (e) GPCP rainfall anomalies with the CGTI in June over (a),(b) the Indian subcontinent and (c)–(e) East Asia for the period 1962–79 in (a),(c) and 1980–97 in (b),(d),(e).

Fig. 7. (a) Anomalies of H200 regressed upon the CGTI in June for the period 1962–79 (shading) and 1980–97 (contours). Contours and shading intervals are 10 gpm. (b) Meridional-mean H200 averaged over 30°–45°N from (a) for the period 1962–79 (line with open circles) and 1980–97 (black line with filled circles). The vertical red and black lines depict the longitudes of the maximum H200 anomalies over East Asia for the period 1962–79 and 1980–97, respectively.
after the late 1970s (Fig. 9a). The center of meridional-averaged H200 anomalies over 30°–45°N shifts eastward about 5° longitude from 125°E during the first epoch to 130°E during the second epoch (Fig. 9b). Meanwhile, the eastward shift of the anomalies of lower-tropospheric anticyclones and ascent over East Asia, related to the AIRI, are also identified after the late 1970s (figure not shown). To conclude, the Indian rainfall-induced CGT pattern changes after the late 1970s, with the anticyclonic anomaly over East Asia shifting eastward, leading to the subsequent eastward shift of the increased rainfall from north China to the Yellow Sea (Figs. 2 and 6). Accordingly, the effect of the Indian early summer monsoon on north China rainfall weakens after the late 1970s.

c. Role of early summer basic-state change

The eastward shift of the anticyclonic anomaly over East Asia reported in section 3b, associated with the Indian rainfall-induced CGT pattern, is captured by the linear barotropic vorticity equation model depicted in section 2 based on the basic state at 200 hPa in June during the two epochs with an identical prescribed divergence forcing centered over northeastern India (Fig. 10). Two similar midlatitude wave trains are seen in response to the forcing over India based on the basic state during the two epochs, associated with an eastward propagation of wave-activity flux from northern India to East Asia. The location of the anticyclonic anomaly over southwestern Asia is nearly identical. However, the anticyclonic anomaly over East Asia shifts eastward approximately by 5° longitude during the second epoch (contour) in comparison with that during the first epoch (shading). The simulated change in the wave train in response to the forcing over India is consistent with that the CGT pattern change revealed in the observation (Figs. 7 and 9), though the centers of anomalies are located southward and eastward in the model simulations.
Since the only difference between the two simulations is in the basic state, the simulations highlight that the basic-state change may be a main cause of the observed CGT pattern change over East Asia after the late 1970s.

The change in the basic state over East Asia in early summer is characterized by a westerly anomaly in the northern flank of the upper-tropospheric westerly jet and an easterly anomaly in the southern flank (Fig. 11a). Subsequently, the East Asian upper-tropospheric westerly jet tends to shift northward after the late 1970s, consistent with that revealed by Yu and Zhou (2007). But how does this basic-state change lead to the CGT pattern change over East Asia?

As shown in Fig. 7, the eastward shift of the anticyclonic anomaly over East Asia and the fixed one over southwestern Asia related to the CGT indicates an increased zonal wavelength of the CGT pattern about 7° longitude over East Asia. In theory, wavelength or wavenumber of the CGT is determined by the basic state (Hoskins and Ambrizzi 1993; Ding and Wang 2005). According to Hoskins and Ambrizzi (1993), the stationary Rossby wavenumber KS is calculated as

$$\text{KS} = \sqrt{\frac{\beta}{\frac{\partial^2 U}{\partial y^2}}}$$

where $\beta$ is meridional gradient of the Coriolis parameter, and $U$ is climatology of zonal westerly wind. Figure 11b shows the stationary Rossby wavenumber in

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**Fig. 10.** The 30-day mean H200 response, calculated from streamfunction following the geostrophic balance, to the prescribed divergence forcing over northeastern India subcontinent (red contour) in a barotropic vorticity equation model using the basic flow at 200 hPa during 1962–79 (shading) and 1980–97 (contours). Also shown is wave-activity flux (vectors) associated with the H200 responses during 1962–79, with the scale plotted at the bottom-right corner. The maximum of the divergence forcing is $1 \times 10^{-8} \text{s}^{-1}$, and intervals of shading and contours are 4 gpm.
June at 200 hPa for the basic flow during the pre-1980 epoch. A maximum wavenumber appears around the westerly jet, acting as a waveguide. The wavenumber is reduced within the waveguide region where the KS contour is larger than 6 during the post-1980 epoch (Fig. 11b). The reduced wavenumber corresponds to an increase in wavelength within the East Asian westerly jet waveguide ranging from 3° to 9° longitude (Fig. 11c), close to the changes of 5° longitude in the model simulations (Fig. 10) and 7° longitude in the observations (Fig. 7). The increased wavelength corresponds to the eastward shift of the anticyclonic anomaly over East Asia related to the CGT (Figs. 7 and 10), with the fixed anticyclonic anomaly over southwestern Asia due to the fixed heating forcing over India (Fig. 10).

d. July and August situations

Wu and Wang (2002) have noticed the weakened impact of the Indian rainfall on north China rainfall during summer after the late 1970s. So far we have concentrated on the decadal change in early summer. What happened to the relationship in July and August between the Indian and north China rainfall? Figure 12 shows the correlation coefficients between the AIRI and NCRI in July and August, with a sliding window of 15 yr. The relationship in August is weakened after the 1970s, similar to that in June. We regress August CRU rainfall against the simultaneous AIRI during the two epochs (figures not shown). A significant positive rainfall anomaly is revealed in north China in the first epoch (1962–79), resembling that in June (Fig. 2a), but its center shifts northwestward in August. In the second epoch (1980–97), however, there is no significant rainfall signal found in north China, consistent with the weakened correlation after the late 1970s shown in Fig. 12.

To investigate circulation anomalies related to the decadal change in August, Fig. 13 shows the CGT pattern over Eurasia related to the August AIRI during the two epochs, respectively. The anticyclonic anomaly over East Asia is significant in both epochs, but it shifts northeastward from north China in the first epoch to northeastern China in the second epoch. Meanwhile, the anticyclonic anomaly over southwestern Asia displaces northeastward. The CGT pattern change is probably due to different mechanisms responsible for the midlatitude wave train formation. As shown in Fig. 13a, the westward extension of the anticyclonic anomaly from southwestern Asia to northern Africa in the first epoch is probably the result of Rossby wave response to the Indian rainfall (Rodwell and Hoskins 1996). In the second epoch (Fig. 13b), there is a significant anticyclonic anomaly over Europe and wave-activity flux propagated from Europe to southwestern Asia, although it cannot be clearly shown in Fig. 13b. This upstream anomaly may imply enhanced impact of high-latitude disturbances on the CGT pattern and then Indian rainfall (Syed et al. 2012).

In July, the decadal change in the relationship between the Indian and north China rainfall is different from that in June and August. It is enhanced after the late 1970s and then weakened after mid-1990s (Fig. 12). Correspondingly, the CGT pattern is clear in the upper troposphere related to the AIRI during 1980–97, but not during 1962–79 and weak during 1998–2014 (figures not shown). The CGT pattern is possibly induced by the AIRI-related rainfall anomaly in northwestern India during 1980–97 (Ding and Wang 2005). Before the late 1970s and after the mid-1990s, the CGT pattern in July is
probably attributed to the impact of tropical western Indian Ocean rainfall anomaly (Chen and Huang 2012).

The above analysis therefore shows that the weakened impact of the Indian rainfall on north China rainfall during summer is caused by the change in the India–north China rainfall relationship in June and August, but not July. Meanwhile, these mechanisms for the decadal changes in India–north China rainfall relationship in July and August are different from that for the June relationship. Since this study focused on the impact of basic-state change on the CGT, we only briefly discussed the India–north China rainfall relationship in July and August. The present results suggest that the decadal changes in the India–north China subseasonal rainfall relationship are complicated and request further investigation.

### 4. Conclusions and discussion

In this study, a significant change is found in the early summer relationship between the Indian and north China rainfall around the late 1970s. The Indian rainfall is significantly linked to north China rainfall in early summer before the late 1970s; specifically, heavier Indian rainfall relates to increased rainfall in north China. However, this link is weakened after the late 1970s.

The weakening of the early summer relationship is attributed to the weakened impact of the Indian early summer rainfall-induced midlatitude wave train on the north China rainfall. The wave train over the Asian continent is trapped within the strong Asian upper-tropospheric westerly jet, dominated by two anticyclonic anomalies over southwestern Asia and East Asia, respectively. After the late 1970s, with the one over southwestern Asia remaining at the nearly same location, the anticyclonic anomaly over East Asia displaces eastward, leading to associated eastward shift of the ascending motion and rainfall from north China to the Yellow Sea. Therefore, the impact of the Indian early summer rainfall on rainfall in north China is weakened after the late 1970s.
Moreover, it is proposed that the eastward displacement of the anticyclonic anomaly over East Asia is due to the change in the basic flow (i.e., a northward shift of the East Asian upper-tropospheric westerly jet in early summer after the late 1970s). The change in the basic state causes a decrease in the stationary Rossby wave number along the East Asian upper-tropospheric westerly jet waveguide. Accordingly, the wavelength of the midlatitude wave train increases so that the anticyclonic anomaly over East Asia displaces eastward corresponding to the fixed anticyclonic anomaly over southwestern Asia caused by the anomalous forcing over India. The impact of the basic-state change is confirmed by the results of simulation with a barotropic vorticity equation model.

Related to the CGT pattern in June, rainfall over the Indian subcontinent also shows a slightly eastward shift after the late 1970s with decreased rainfall anomaly in northwestern India (Fig. 6b). The eastward shift of rainfall anomalies is probably attributed to the change in the lower-tropospheric circulation. Related to the CGT, a westerly anomaly is identified across the east coast of the Indian subcontinent forming a cyclonic anomaly to the north over central-northeastern India during the second epoch, while a significant cyclonic anomaly is seen over northwestern India during the first epoch (figures not shown). The cyclonic anomalies may enhance lower-tropospheric convergence, ascent, and then rainfall in northwestern India during the first epoch and in central-northeastern India during the second epoch, consistent with the reduced rainfall in northwestern India and the eastward shift of the rainfall anomalies.

The present study highlights the role of the basic-state change over East Asia in the weakened connection in early summer between the Indian and north China rainfall after the late 1970s. This mechanism differs from that proposed by Wu and Wang (2002) who postulated that the reduction of interannual variability of the Indian summer rainfall weakened its connection with the midlatitude wave train and then summer rainfall in north China this study reveals that the connection of all Indian rainfall to the midlatitude wave train remains stable around the late 1970s. This difference may arise from the complicated decadal change in summer season, which is implied by the weakened relationship in June and August, but enhanced relationship in July after the late 1970s (Fig. 12).

The mechanism of basic-state change, however, may not be capable of explaining the weakened impact of the Indian rainfall on north China rainfall in summer season presented by Wu and Wang (2002), because of a subseasonal difference between June and July–August in decadal change of the basic flow over East Asia after the late 1970s. The westerly jet tends to shift northward in June (Fig. 11a of this study) and southward in July and August (Yu and Zhou 2007; Zhang and Huang 2011). The opposite change may lead to only a weak change in the summer-mean westerly jet and then the CGT wave pattern after the late 1970s.

This study suggests that decadal changes in circulation can significantly modulate the interannual relationship in rainfall between the remote regions through modifying the shape of the CGT pattern. Some studies have shown that the Asian jet may intensify in the late-twenty-first century under global warming (Zhang and Guo 2010; Dai and Lu 2013), which may imply possible change in the spatial pattern of the CGT and then rainfall relationship between India and East Asia. Further investigations on the change of the rainfall relationship under global warming are requested.

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