

A Contrast of the East Asian Summer Monsoon–ENSO Relationship between 1962–77 and 1978–93*

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ABSTRACT

Using station rainfall data and the NCEP–NCAR reanalysis, the authors investigate changes in the interannual relationship between the east Asian summer monsoon (EASM) and El Niño–Southern Oscillation (ENSO) in the late 1970s, concurrent with the Pacific climate shift. The present study focuses on decaying phases of ENSO because changes in developing phases of ENSO are less significant. Remarkable changes are found in the summer rainfall anomaly in northern China and Japan. From pre- to postshift period, the summer rainfall anomaly in eastern north China during decaying phases of El Niño changed from above to below normal, whereas that in central Japan changed from negative to normal. Consistent with this, the barotropic anticyclonic anomaly over the Japan Sea changed to cyclonic; the associated anomalous winds changed from southerly to northerly over the Yellow Sea–northeastern China and from northeasterly to northwesterly over central Japan.

The change in the ENSO-related east Asian summer circulation anomaly is attributed to changes in the location and intensity of anomalous convection over the western North Pacific (WNP) and India. After the late 1970s, the WNP convection anomaly is enhanced and shifted to higher latitudes due to increased summer mean SST in the Philippine Sea. This induces an eastward shift of an anomalous low pressure from east Asia to the North Pacific along 30°–45°N during decaying phases of El Niño. Thus, anomalous winds over northeastern China and Korea switch from southeasterly to northeasterly. Before the late 1970s, an anomalous barotropic anticyclone develops over east Asia and anomalous southerlies prevail over northeastern China during decaying phases of El Niño. This may relate to anomalous Indian convection through a zonal wave pattern along 30°–50°N. After the late 1970s, anomalous Indian convection weakens, which reduces the impact of the Indian convection on the EASM.

1. Introduction

The east Asian summer monsoon (EASM) is a subtropical monsoon encompassing eastern China, Japan, and Korea (20°–45°N, 100°–140°E). Many previous works have studied the relationship between the EASM and El Niño–Southern Oscillation (ENSO; e.g., Huang and Wu 1989; Chen et al. 1992; Shen and Lau 1995;

Zhang et al. 1996; Weng et al. 1999; Lau and Weng 2001). The summer rainfall anomaly in eastern China shows a very different distribution when comparing the ENSO developing year to the following year for the period of 1951–80 (Huang and Wu 1989). Shen and Lau (1995) identified a pronounced quasi-biennial signal in the April–September rainfall variability over a broad region in the vicinity of the Yangtze River valley (28°–32°N, 110°–120°E). Their lag correlation analysis shows that above-normal rainfall in the Yangtze River valley is preceded by a warm equatorial eastern Pacific in the previous winter. Chang et al. (2000a) obtained a similar result. Thus, rainfall is significantly greater in the summer after an El Niño. The devastating Yangtze River flood in the summer of 1998 that followed the 1997 El Niño is such an example (Lau and Weng 2001). El Niño episodes, however, mature usually in boreal winter, and by the next summer the warming in the equatorial central-eastern Pacific has disappeared. How then does El

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Niño have its “delayed” influence on the EASM? Wang et al. (2000) revealed that the circulation system that conveys the impact of El Niño to east Asia is an anomalous low-level anticyclone over the Philippine Sea. The anomalous Philippine Sea anticyclone develops rapidly in the boreal fall of an El Niño year and persists until the ensuing summer, strengthening the western Pacific subtropical ridge in early summer, which then causes the abundant precipitation in the lower reaches of the Yangtze River valley. Chang et al. (2000a) also elaborated how the anomalous Philippine Sea anticyclone affects May–June rainfall in the Yangtze River valley. The persistence of the anomalous Philippine Sea anticyclone could be due to a positive thermodynamic feedback between atmospheric Rossby waves and the underlying warm pool ocean (Wang et al. 2000).

An interdecadal change was observed in the late 1970s in the tropical and North Pacific sea surface temperature (SST; e.g., Nitta and Yamada 1989; Trenberth and Hurrell 1994; Graham 1994; Wang 1995; Zhang et al. 1997) and the tropical Pacific convection (e.g., Nitta and Yamada 1989; Graham 1994). The ridge of the western North Pacific (WNP) subtropical high shifted southward after the late 1970s (Nitta and Hu 1996). The inverse relationship between the Indian summer rainfall and equatorial eastern Pacific SST has broken down in recent decades (Webster et al. 1998; Kumar et al. 1999). Several hypotheses have been proposed to explain the weakened relationship. Kumar et al. (1999) suggested that a southeastward shift in the Walker circulation anomalies associated with ENSO may lead to a reduced subsidence over the Indian region, thus favoring normal monsoon conditions. Another reason is increased surface temperature over Eurasia in winter and spring that may favor the enhanced land–ocean thermal contrast conducive to a stronger Indian monsoon. Chang et al. (2001) showed that a most likely cause for the broken relationship is the strengthening and poleward shift of the jet stream over the North Atlantic. The latter enhances the European surface temperature anomalies in winter and spring. The effect of the resulting meridional temperature contrast disrupts the influence of ENSO on the Indian monsoon.

As an important component of the Asian monsoon system, the east Asian monsoon has its distinct features (Lau and Li 1984; Tao and Chen 1987). A question to be addressed is whether the interannual relationship between the EASM and ENSO is altered concurrently with the change in the Pacific mean state. If so, what causes the change in the EASM–ENSO relationship? According to Chang et al. (2000a), above-normal May–June rainfall in the Yangtze River valley follows a warm equatorial eastern Pacific in the preceding winter for both 1951–77 and 1978–96. Wang et al. (2001) found that over the WNP and the Yangtze River region, the ENSO–monsoon relationship has been steady in the last 40 years. However, the present study will show that, different from the Yangtze River region and the WNP,

the ENSO–monsoon relationship in northern China and Japan has experienced a remarkable change since the late 1970s.

The main purpose of the present study is to investigate interdecadal changes in the relationship between the east Asian summer rainfall and the ENSO cycle. We analyzed changes in summer rainfall anomaly in both the developing and decaying phases of ENSO. Because the changes identified in the developing phases of ENSO are less significant than those in the decaying phases of ENSO, we will focus on changes in the decaying phases of ENSO in this study. The arrangement of the text is as follows. Section 2 describes data and methods used in this study. In section 3, we contrast correlation patterns of east Asian summer rainfall with previous winter Niño-3 (5°S – 5°N , 150° – 90°W) SST anomalies between the two periods of 1962–77 and 1978–93 and address relevant changes in ENSO-related circulation anomalies over Asia and the western Pacific. Then, we discuss roles of anomalous WNP and Indian convection and their interdecadal changes (section 4). The plausible causes for those changes are discussed in section 5. Summary and discussion are provided in section 6.

2. Data and methods

The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996) provides the atmospheric data for the period 1948–97. The variables used for this study include winds at 850 and 200 hPa; geopotential height at 200, 500, and 850 hPa; and vertical p velocity at 500 hPa. These data are on regular grids with a horizontal resolution of $2.5^{\circ} \times 2.5^{\circ}$. The skin temperature and winds at 10 m are derived from the NCEP–NCAR reanalysis on Gaussian grids of T62 with a horizontal resolution of about $1.9^{\circ} \times 1.9^{\circ}$. Over the open ocean, the skin temperature is fixed at the Reynolds SST (Reynolds and Smith 1994) in the NCEP–NCAR reanalysis.

Station rainfall data include 160 stations in mainland China, 28 stations in Japan, 10 stations in South Korea, and 4 stations in Taiwan. The Chinese station data span the period of 1951–97 and are provided by the Chinese Meteorological Data Center. Station rainfall data in Japan are obtained from the Global Historical Climatology Network (GHCN) of the National Climate Data Center (NCDC) of the National Ocean and Atmosphere Administration (NOAA, online at <http://www.ncdc.noaa.gov>). This dataset has been subjected to quality control checks (Vose et al. 1992). Only those stations with data up to 1993 are selected. The South Korea station rainfall data are by courtesy of J.-S. Kug, Seoul National University of Korea and span the period of 1961–2000. Rainfall data for four Taiwan stations for the period of 1897–2000 are provided by Dr. M.-M. Lu, Central Weather Bureau of Taiwan. All Indian monthly rainfall

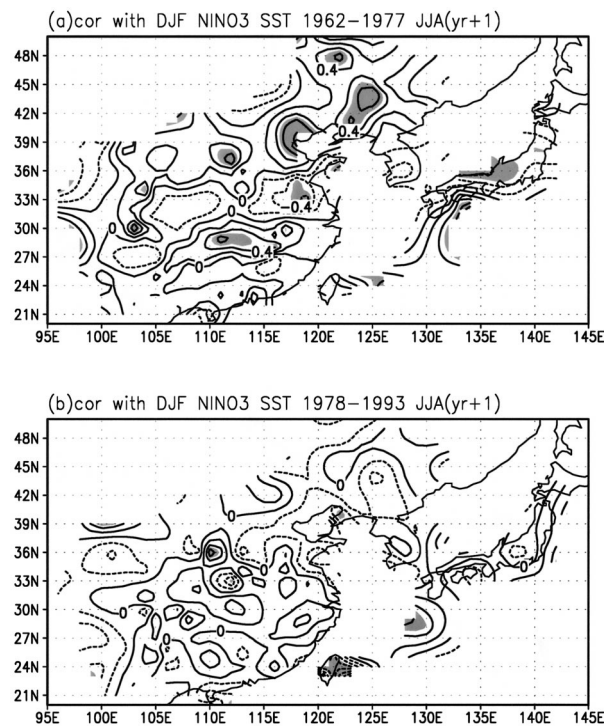


FIG. 1. Correlation of summer rainfall with previous winter Niño-3 SST in (a) 1962–77 and (b) 1978–93. Shading denotes correlation significant at 95% confidence level. Contour interval is 0.2.

data for the period of 1871–1994 are from Parthasarathy et al. (1994, 1995).

To document the plausible interdecadal change in the late 1970s, we take two epochs: 1962–77 and 1978–93, each epoch consisting of 16 yr. The selection of the two epochs is based on the sliding lag correlation analysis (section 3). The correlation and regression patterns with respect to previous winter [December–January–February (DJF)] Niño-3 SST anomalies are calculated separately for the two epochs. Using the Niño-3 SST index as a reference instead of regional mean monsoon rainfall allows identifying those regions where the relationship with ENSO may experience large changes.

3. Contrasting east Asian summer monsoon anomaly in the decaying phase of ENSO

a. Rainfall anomaly

To uncover where the east Asian summer rainfall–ENSO relationship experiences the most noticeable change in the late 1970s, we contrast the correlation of summer (JJA) rainfall with the preceding winter (DJF) Niño-3 SST between 1962–77 and 1978–93. The correlation pattern displays remarkable differences between the two epochs. In 1962–77 significant positive, negative, and positive correlation is located in the Yangtze River region (28° – 32° N, 110° – 120° E), the Huai River region (32.5° – 35.5° N, 115° – 120° E), South Korea–cen-

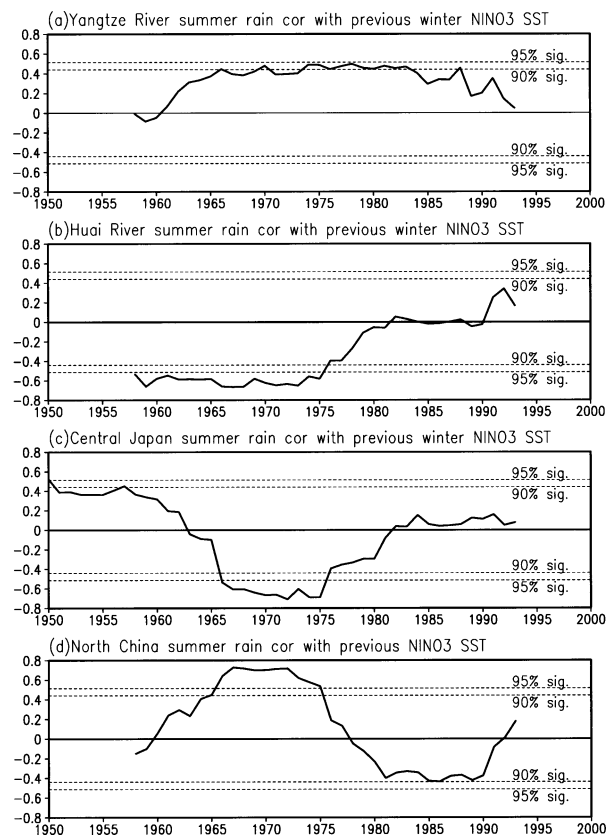


FIG. 2. Sliding correlation of area mean summer rainfall with previous winter Niño-3 SST with a window of 15 yr. (a) Yangtze River region (28° – 32° N, 110° – 120° E), (b) Huai River region (32.5° – 35.5° N, 115° – 120° E), (c) central Japan region (35° – 40° N, 132° – 144° E), and (d) north China region (38° – 44° N, 115° – 125° E). The dashed horizontal lines denote the 90% and 95% confidence levels for the correlation coefficient.

tral Japan (35° – 40° N, 132° – 144° E), and eastern north China (38° – 44° N, 115° – 125° E), respectively (Fig. 1a). In 1978–93 positive correlation appears in the Yangtze River region and negative correlation is present in eastern north China (Fig. 1b). Apparently, the largest contrast between the two epochs is found in eastern north China and central Japan. The correlation in the Yangtze River region does not show noticeable change, consistent with Chang et al. (2000a).

To validate the selection of the two periods, we show in Fig. 2 the evolution of correlation of summer rainfall in chosen regions with previous winter Niño-3 SST using a sliding window of 15 years. The Yangtze River region does not experience a large change during 1962–88 (Fig. 2a). The Huai River region and central Japan region show a significant negative correlation before the mid-1970s, followed by an obvious weakening of correlation in the late 1970s (Figs. 2b,c). The eastern north China region displays a remarkable change from significant positive correlation to negative correlation in the late 1970s (Fig. 2d). Another change can be noticed in the early 1960s in the Yangtze River, central Japan,

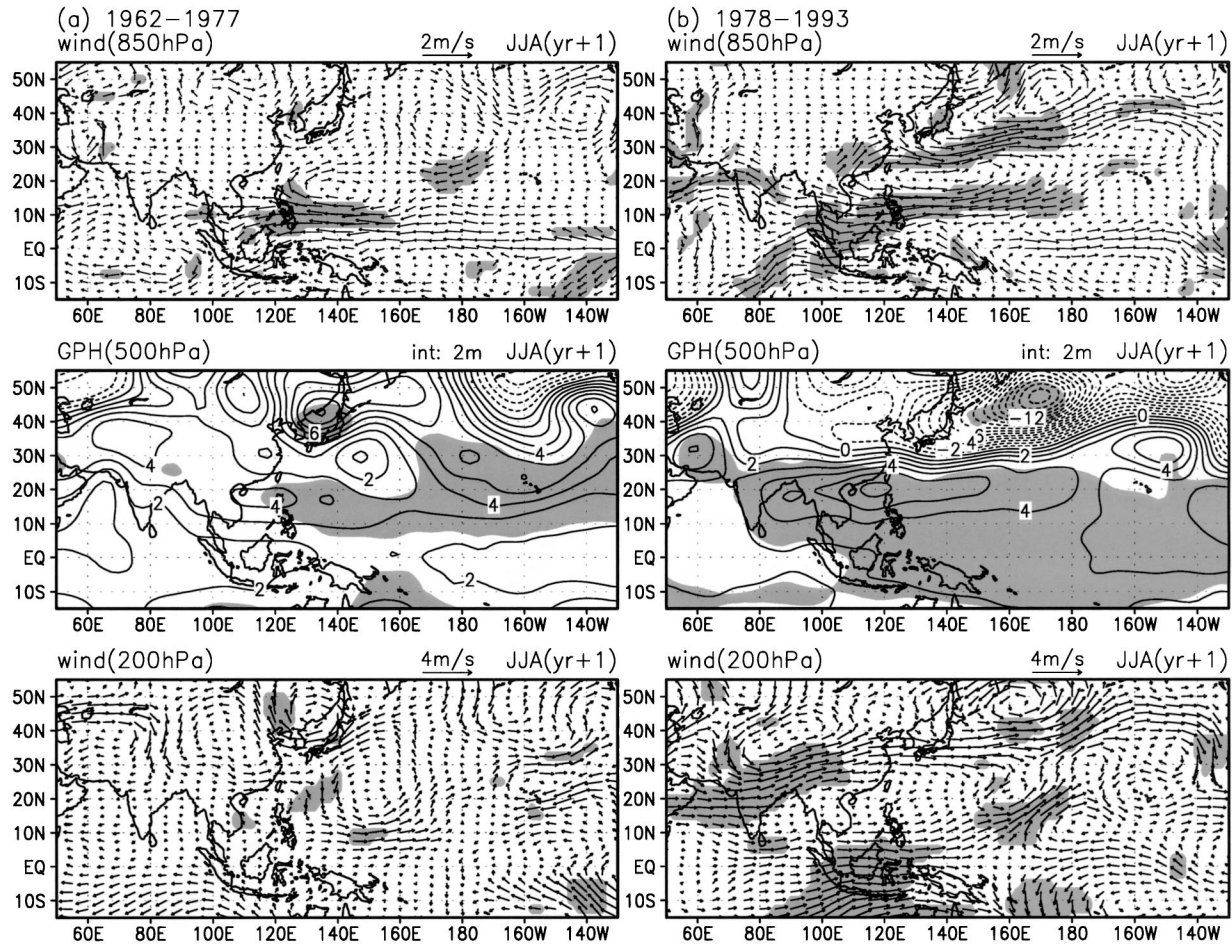


FIG. 3. Regression pattern of summertime circulation with respect to previous winter Niño-3 SST for (a) 1962–77 and (b) 1978–93. From top to bottom are 850-hPa wind, 500-hPa height, and 200-hPa wind, respectively. Shading denotes regions of significance at 95% confidence level. Contour interval for 500-hPa height is 1 m. The wind scale is indicated at top of respective panel.

and eastern north China (Figs. 2a,c,d). The temporal evolution of correlation clearly indicates the change in the late 1970s and the steadiness of correlation in the two epochs chosen before and after the late 1970s. Thus, it would be meaningful to take the two epochs and contrast rainfall and circulation anomalies.

b. Circulation anomaly

The ENSO-related circulation anomalies over east Asia in summer following the mature phase of ENSO show remarkable differences between the two epochs. Figure 3 shows regression patterns of wind and height anomalies with respect to previous winter Niño-3 SST for the periods 1962–77 and 1978–93. In 1962–77 an anomalous barotropic anticyclone spans the Japan–Japan Sea–northeast China region (Fig. 3a). Anomalous southerlies exist over the Yellow Sea and northeastern China. Anomalous northeasterlies overlie Japan. In contrast, in 1978–93 an anomalous cyclone extends from the North Pacific to the east Asian coast (Fig. 3b). Over

northeastern China–northern Japan there are anomalous northeasterly winds. Southern Japan is under the impact of anomalous northwesterly winds.

The circulation anomalies over midlatitudes of Asia display different features. In 1978–93 upper-level anomalous westerly winds are seen along 25°–40°N from the Tibetan Plateau all the way to the central North Pacific (Fig. 3b). In 1962–77 the upper-level anomalous flow displays an obvious wave pattern: an anticyclone northwest of India, a cyclone north of the Tibetan Plateau, and an anticyclone over northeastern China and the Japan Sea (Fig. 3a). Though wind anomalies in 1962–77 over Asia are not significant, the striking differences between the two epochs suggest important changes in the influence of ENSO on the Asian circulation in the late 1970s.

The circulation anomalies over the tropical WNP show both common and distinct features. Large anticyclonic wind anomalies are observed to the east of the Philippines at a low level in both epochs (Fig. 3). At 500 hPa there are positive height anomalies. The anti-

cyclone is capped by upper-level cyclonic convergent winds. These circulation anomalies feature a weakened convection over the WNP. In contrast, the low-level anomalous anticyclone is stronger and located at higher latitudes in 1978–93 than in 1962–77. The WNP subtropical high is intensified mainly over the South China Sea and Philippines in 1962–77 (Fig. 3a). In 1978–93 the positive and significant height anomalies extend all the way from India to the date line (Fig. 3b).

The changes in the circulation anomalies over east Asia are coherent with those in the rainfall anomalies. In 1962–77 anomalous southerly winds over the Yellow Sea and northeastern China are expected to bring more moisture to eastern north China and favor more precipitation. Anomalous northeasterly winds over Japan are consistent with less precipitation. In 1978–93 anomalous northeasterly winds over northeastern China indicate an inflow of drier air, which leads to less precipitation.

4. Impacts of anomalous convection over the WNP and India on east Asia

Why does the circulation anomaly over east Asia differ in the two epochs? Different factors, such as changes in location and intensity of tropical heating and inland surface conditions, may each play a role in the difference. The change in mean circulation over east Asia may also modulate the atmospheric response to tropical heating. In this section, we discuss impacts of anomalous tropical convection over the WNP and India and their changes. The plausible causes for the interdecadal changes in those impacts will be discussed in the next section.

a. Influence of anomalous WNP convection

The influence of the WNP convection anomaly on the east Asian circulation was indicated in previous studies (Nitta 1987; Tsuyuki and Kurihara 1989; Huang and Sun 1992). It was suggested that the heating due to anomalous convection near the Philippines can excite barotropic Rossby waves propagating to east Asia (Nitta 1987; Kurihara and Tsuyuki 1987). The induced height and wind anomalies tend to display a zonally elongated band structure in summer (Wang et al. 2001). Such a zonal band structure is observed in summer rainfall anomaly distribution as well, suggesting a link between the WNP convection and east Asian rainfall anomalies. However, it is not clear whether the influence of anomalous WNP convection on the east Asian summer rainfall experienced a change in the late 1970s. To answer this question, in the following we compare the relationship of the east Asian rainfall and circulation with the WNP summer monsoon variability between the two epochs. For convenience, the WNP summer monsoon is measured by an index [the WNP monsoon index (WNPMI)] defined as the difference of the 850-hPa zon-

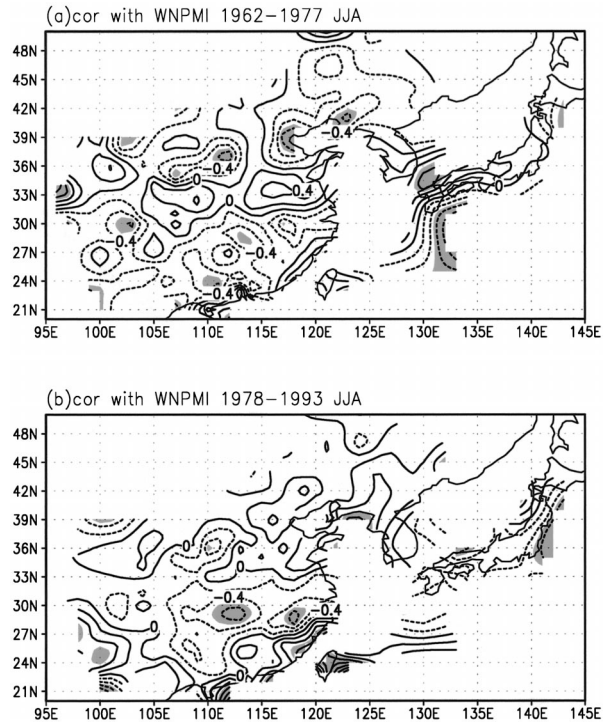


FIG. 4. Simultaneous correlation of summer rainfall with the WNPMI for the period of (a) 1962–77 and (b) 1978–93. Shading indicates correlation significant at 95% significance level. Contour interval is 0.2.

al wind anomaly between 5° – 15° N, 100° – 130° E and 20° – 30° N, 110° – 140° E (Wang et al. 2001).

The simultaneous correlation of the east Asian summer rainfall with the WNPMI shows noticeable changes in the late 1970s. This is demonstrated in Fig. 4, which compares the correlation distribution of summer rainfall with the WNPMI before and after the late 1970s. In 1962–77 the correlation displays alternative negative, positive, and negative distribution over southern China, the Huai River and Korea Strait, and north-northeastern China (Fig. 4a), respectively. In 1978–93 positive, negative, and positive correlation lies over the south China coast, Yangtze River–central Japan, and northeastern China (Fig. 4b), respectively. The correlation distribution in central and northern China is similar (with the sign reversed) to that shown in Fig. 1 for both epochs.

The simultaneous correlation of 500-hPa geopotential height with the WNPMI in summer shows both common features and noticeable differences. Figure 5 shows the correlation of summer 500-hPa height with the WNPMI in the two epochs before and after the late 1970s. Both epochs reveal a meridional wave-type response over east Asia and WNP and the height anomaly tends to display a zonally banded structure (Fig. 5), consistent with previous studies (Nitta 1987; Wang et al. 2001). A noticeable difference in the structure is found over east Asia. In 1978–93 a zonally elongated band of positive correlation extends from Huai River through Korea

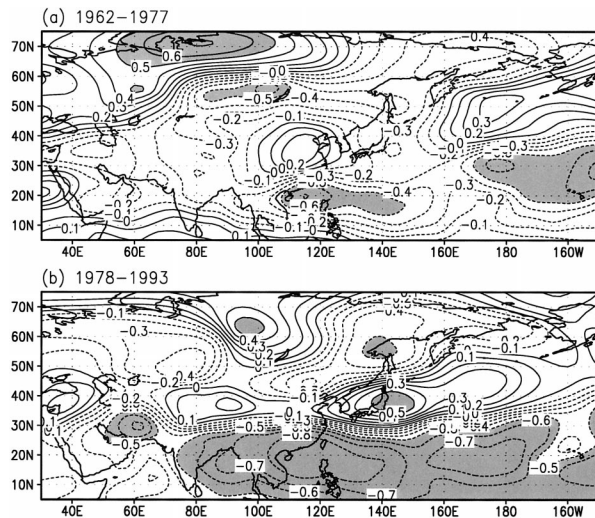


FIG. 5. Simultaneous correlation of summer 500-hPa geopotential height with the WNPMI for the period of (a) 1962–77 and (b) 1978–93. Shading indicates regions where correlation coefficient is over 0.5 (significant at 95% confidence level). The contour interval is 0.1. The correlation is calculated using interannual component (periods less than 8 yr) only.

and Japan to the North Pacific (Fig. 5b). As such, southwesterly (northeasterly) wind anomalies blow over north-northeastern China and Korea when the WNP summer monsoon is stronger (weaker) than normal. In contrast, in 1962–77 a negative correlation region is seen near Lake Baykal and a weak patch of positive correlation appears over eastern China (Fig. 5a). The anomalous pressure gradient over the Yellow Sea–Korea indicates anomalous northerly (southerly) winds when the WNP summer monsoon is stronger (weaker) than normal. Further examination shows that in 1962–77 the 850-hPa meridional wind has a significant negative correlation with the WNPMI over North Korea–southern northeast China. In 1978–93 the 850-hPa meridional wind has significant positive correlation in the Yellow Sea–Korea–Japan Sea region.

ENSO can induce large wind and convection anomalies over the WNP (Wang et al. 2000; Wu and Wang 2000). Thus, one way by which ENSO influences the EASM is through anomalous WNP heating. The identified change in the relationship between the WNPMI and the east Asian summer rainfall and circulation indicates that the impact of ENSO on the EASM through anomalous WNP convection is very different before and after the late 1970s.

b. Influence of anomalous Indian convection

Previous studies indicated a positive correlation between the Indian and northern China summer rainfall (Tao and Chen 1987; Kripalani and Singh 1993; Kripalani and Kulkarni 1997; Zhang et al. 1999; Kripalani and Kulkarni 2001) and a negative correlation between

the Indian and southern Japan summer rainfall (Krishnan and Sugi 2001; Kripalani and Kulkarni 2001). Composite analysis identified the presence of a barotropic anticyclone over northeastern China when the Indian summer monsoon is stronger than normal (Lau et al. 2000; Wang et al. 2001). A positive correlation is observed over northeastern China in the correlation field of 500-hPa geopotential height with Indian rainfall during summer on the intraseasonal timescale (Kripalani et al. 1997). The anomalous anticyclone over northeastern China seems to be responsible for the rainfall anomalies in east Asia in association with anomalous Indian summer monsoon. However, the process for the occurrence of the anomalous anticyclone over northeastern China is not clear. A reasonable scenario is speculated as follows. Anomalous Indian monsoon heating modulates the occurrence of an upper-level anomalous anticyclone northwest of India by a Rossby-type response to tropical heating located off the equator (Gill 1980). Such an anomalous anticyclone is observed in previous studies dealing with the Indian monsoon variability (Kawamura 1998; Lau et al. 2000; Wang et al. 2001). The modeling study with specified heating over the Bay of Bengal indicates that the heating can induce subsidence over west-central Asia in summer (Rodwell and Hoskins 1996). The anomalous anticyclone perturbs the upper-level flow in the midlatitude westerly and induces a response downstream through zonal Rossby wave propagation. One effect is the development of an anomalous anticyclone over northeastern China.

To verify the above point of view, we examined the correlation of geopotential height and meridional wind at different levels with reference to 200-hPa height at a grid (35°N, 60°E) near the center of an upper-level anomalous anticyclone induced by anomalous Indian monsoon heating. Figure 6 shows the correlation for geopotential height at 200, 500, and 850 hPa. A clear wave pattern is present along 30°–50°N, especially at 200 hPa (Fig. 6a). The negative correlation north of the Tibetan Plateau is weaker. The correlation of 200-hPa meridional wind shows a remarkable alternation of high positive and negative correlation from the Caspian Sea to Japan. The positive correlation over China and the negative correlation over Japan are as high as 0.8 and −0.6, respectively (figure not shown). This suggests the possibility for the development of an upper-level zonal wave pattern when a perturbation is induced west of the Tibetan Plateau. At 500 hPa, the positive correlation northwest of India shifts eastward (Fig. 6b) compared to that at 200 hPa. At 850 hPa, negative correlation lies over the Arabian Peninsula–Arabian Sea–India region (Fig. 6c). The baroclinic structure over west and south Asia indicates the role of tropical heating, and therefore anomalous Indian convection. In fact, the correlation fields of geopotential height with all-Indian summer rainfall (AISR) display a very similar spatial structure as Fig. 6. Krishnan and Sugi (2001) identified a similar

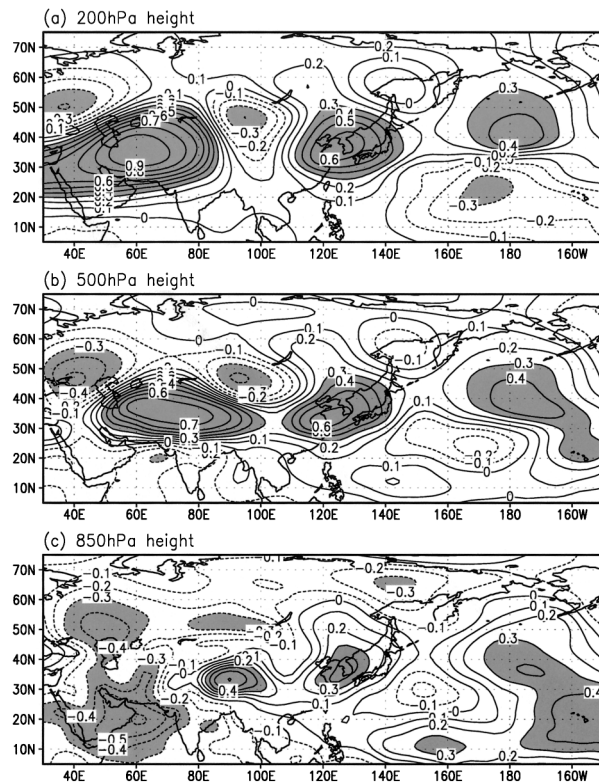


FIG. 6. Simultaneous correlation of summer 200-hPa height at 35°N and 60°E with geopotential height at (a) 200, (b) 500, and (c) 850 hPa for the period of 1948–97. Shading denotes regions where the correlation coefficient is over 0.3. The contour interval is 0.1. The correlation is calculated using interannual component only.

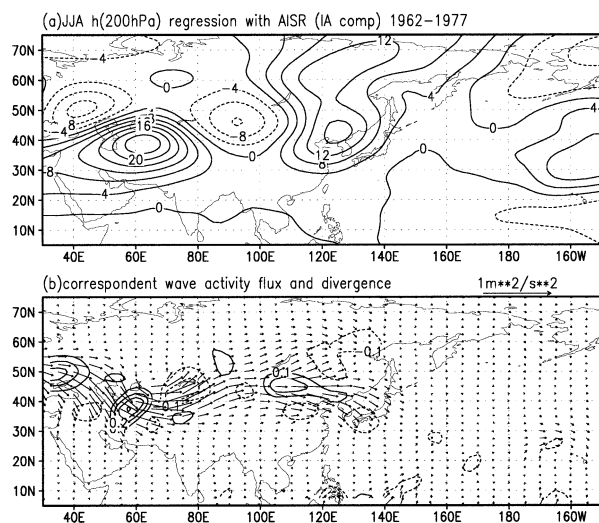


FIG. 7. Regression pattern of (a) summer 200-hPa height with respect to AISR for the period of 1962–77 and (b) correspondent wave activity flux and its divergence. The contour interval is 4 m in (a) and $0.1 \times 10^{-6} \text{ m s}^{-2}$ in (b). The zero contour in (b) is suppressed. The scale for the flux vector is shown on top-right of the panel. The regression is calculated using interannual component only.

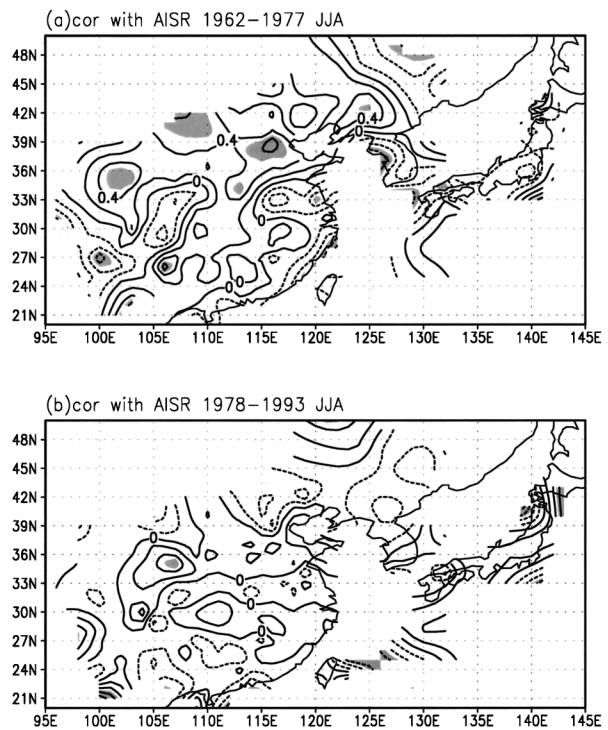


FIG. 8. Same as Fig. 4 but with AISR.

pattern for 200-hPa anomalous winds in the wet-minus-dry composite based on baiu rainfall.

To explore the possibility of propagation of stationary wave activity, we utilize a diagnostic tool for the horizontal components of stationary wave activity flux (Karoly et al. 1989; Barlow et al. 2001) that is a simplification of the three-dimensional case (Plumb 1985). The region where the wave flux has divergence indicates a source of wave activity. A weakness of this diagnostic

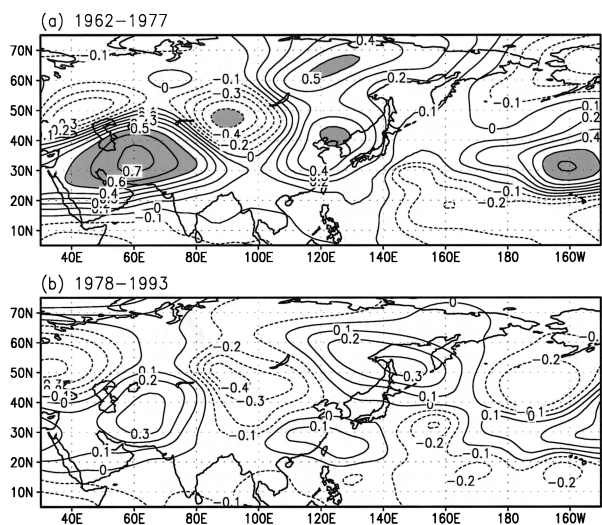


FIG. 9. Same as Fig. 5 except for the correlation is for 200-hPa height with AISR.

tool is its inability to identify wave propagation away from the tropical forcing (Karoly et al. 1989). Figure 7 displays the regression of summer 200-hPa height with respect to the AISR and the correspondent wave flux vector for the period of 1962–77 during which time the wave pattern is prominent. Clearly, there is propagation of wave activity along 30° – 50° N (Fig. 7b) corresponding with a zonal wave pattern for the 200-hPa height anomaly (Fig. 7a). In the region northwest of India, the wave flux has a large divergent component (Fig. 7b), suggesting that this region is a (but not the only) source region for wave activity. Note that the wave flux can be traced to southern Europe and there is a large source region near 50° N and 30° E, suggesting the influence from Europe.

The above discussions provide a plausible dynamical link between anomalous Indian heating and east Asian circulation. As ENSO plays an important role in the Indian summer monsoon variability (Webster et al. 1998), another way by which ENSO can influence the east Asian summer circulation is through anomalous Indian heating. A question is whether the influence of anomalous Indian convection on the EASM differs between the two epochs. To answer this question, we display in Fig. 8 the correlation distribution of summer rainfall with the AISR for the two epochs 1962–77 and 1978–93. The correlation of the east Asian summer rainfall with the AISR shows obvious differences between the two epochs (Fig. 8). The positive correlation in northern China is significant in 1962–77 (Fig. 8a) and becomes weak or negative in 1978–93 (Fig. 8b). Significant negative correlation extends from the Huai River valley of China through Korea to central Japan in 1962–77 (Fig. 8a). Notice that the correlation distribution in Fig. 8a is quite similar to that in Fig. 1a. The correlation distribution in Fig. 8b only shows weak similarity to that in Fig. 1b.

The change in the rainfall correlation is consistent with that in the correlation for height and meridional wind. In 1962–77 the positive correlation for 200-hPa height is significant over northeastern China and Korea, as is the negative correlation north of the Tibetan Plateau and the positive correlation northwest of India (Fig. 9a). The correlation field of 200-hPa meridional wind shows alternatively significant positive and negative correlation from the Caspian Sea to the Japan Sea along 30° – 50° N (figure not shown). In the correlation field for 850-hPa meridional wind, the positive correlation over northern China and negative correlation over the Japan Sea are as high as 0.5. In 1978–93 the correlation for 200-hPa height is weak and the positive correlation is displaced away from northeastern China (Fig. 9b). The correlation for 200-hPa meridional wind shows a similar weakening over east Asia, and the correlation centers for 850-hPa meridional wind have very different locations (figures not shown).

The above-mentioned weakened relationship between anomalous Indian convection and east Asian rainfall and

circulation indicates a weakened influence of ENSO on the EASM through anomalous Indian convection. This applies to both the developing and decaying phases of ENSO, though the ENSO-induced Indian rainfall anomaly is larger in the former than in the latter. This argument is supported by the biennial nature of ENSO for the period of 1962–77 (e.g., An and Wang 2000) during which the developing phase of El Niño is also the decaying phase of La Niña. As such, the simultaneous correlation between the AISR and east Asian summer rainfall and circulation in 1962–77 does not distinguish the two phases. In 1978–93 ENSO is of a low-frequency nature. We calculated the correlation as Figs. 8b and 9b but for only the decaying phase of ENSO. Comparison shows that the correlation pattern is very similar to Figs. 8b and 9b, although the correlation coefficient is higher.

5. Discussion: Causes for the change in the EASM–ENSO relationship

a. Causes for the change in the influence of the WNP convection

The identified changes in the correlation fields with the WNPMI (section 4a) raise a question: What caused the change in the response of the east Asian circulation to anomalous WNP convection? Plausible reasons are the change in location and intensity of anomalous heating over the WNP and the change in mean circulation over east Asia and WNP. The distinction of ENSO-related wind and height anomalies over the WNP between the two epochs (Fig. 3) indicates anomalous WNP convection is stronger and located at higher latitudes in 1978–93 than in 1962–77.

The response of the east Asian circulation is sensitive to the location of anomalous WNP heating. This is demonstrated by the change in the correlation of 500-hPa height anomaly with reference to 500-hPa vertical motion at different grids. Figure 10 shows three representative cases with the reference grid of 500-hPa vertical motion located at 15° N and 130° E, 20° N and 130° E, and 20° N and 150° E. Those grids were selected to reflect the change in location of anomalous heating. The correlation is calculated using the interannual component only to avoid the interference from decadal and interdecadal variation (periods over 8 yr). When the reference grid is located at 20° N, the response in the subtropics is an elongated band extending to the date line (Figs. 10b,c). When the reference grid is at 15° N, the response in the subtropics is limited in zonal extension and the correlation center is at a westward location (Fig. 10a). As such, anomalous winds over northeastern China have a southerly component in 1962–77 and a northerly component in 1978–93 when the WNP convection is stronger than normal. This is consistent with the change in the correlation of 500-hPa height with the WNPMI (Fig. 5).

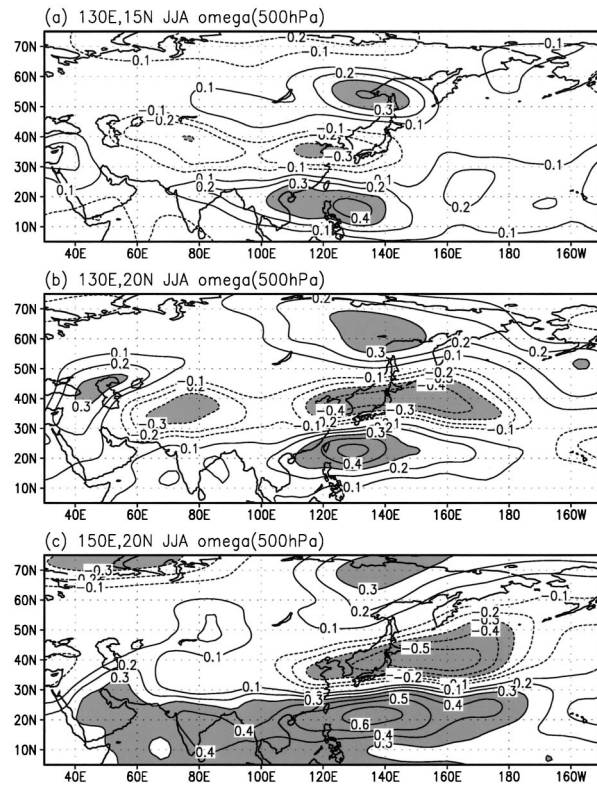


FIG. 10. Simultaneous correlation of summer 500-hPa height with 500-hPa vertical p velocity at reference grid of (a) 15°N and 130°E , (b) 20°N and 130°E , and (c) 20°N and 150°E for the period of 1948–97. Shading indicates regions where correlation coefficient is over 0.3. The contour interval is 0.1. The correlation is calculated using interannual component only.

Considering that the reliability of 500-hPa vertical motion may be lower before the late 1970s, we examine the correlation for the recent period (1978–97). It turns out that the pattern as seen in Fig. 10 is very obvious. In addition, the correlation is stronger and shifts eastward when the reference grid of 500-hPa vertical motion is at 20°N and 150°E than at 15°N and 130°E . To verify the 500-hPa vertical motion in recent decades, we calculate the local simultaneous correlation between 500-hPa vertical motion and the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) rain rate (Xie and Arkin 1997). In summer (JJA), a high correlation region with correlation coefficient over 0.6 is seen over 10° – 40°N and 110°E – 180° . The regression patterns of summer 500-hPa vertical p velocity with respect to previous winter Niño-3 SST (figures not shown) are quite similar to the correlation pattern of station rainfall (Fig. 1) over east Asia. The above evidences suggest that the large-scale pattern obtained based on 500-hPa vertical motion is reasonable.

The change in location of anomalous WNP convection is related to the difference of mean summer SST between the two epochs. The summer SST shows an increase in the equatorial central and eastern Pacific and

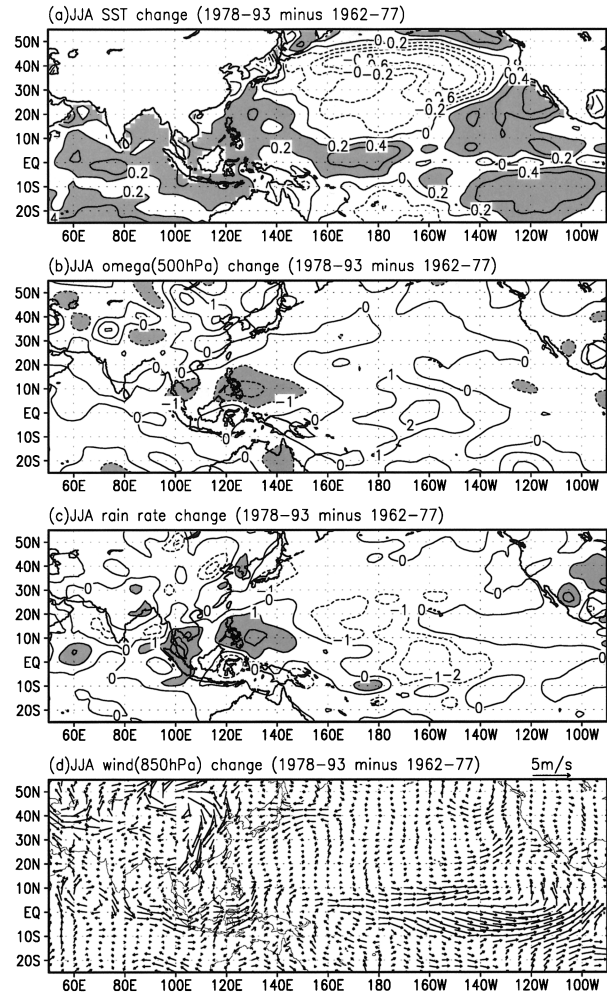


FIG. 11. (a) Difference of mean summer SST, (b) 500-hPa vertical p velocity, (c) rain rate, and (d) 850-hPa winds between 1978–93 and 1962–77. The contour interval is 0.2°C in (a), 0.01 Pa s^{-1} in (b), and 1 mm day^{-1} in (c). The wind scale is shown on the top-right of the panel.

a decrease in the North Pacific (Fig. 11a). An increase in SST is also observed in the tropical WNP, consistent with Nakamura and Yamagata (1999) who showed the presence of a higher SST in the tropical WNP in their EOF2 mode for decadal SST variability in the summertime. Note that the summer SST change in the WNP is of similar magnitude to that in the equatorial central Pacific. Due to the dependence of convection on SST, a higher mean SST in the tropical WNP indicates a stronger mean convection. This is supported by the enhancement in mean upward motion at 500 hPa (Fig. 11b) and the increase in the mean rain rate (Fig. 11c) in summer. The low-level mean wind change shows a large cyclone (Fig. 11d), consistent with the intensification of mean convection.

The numerical results of an atmospheric general circulation model (AGCM) described below demonstrate that anomalous WNP convection induced by anomalous

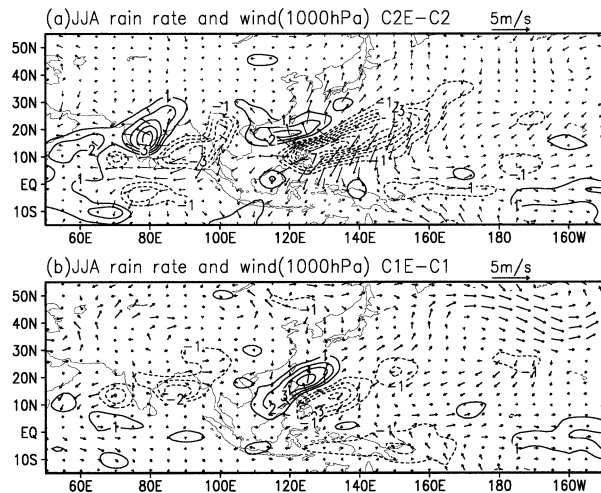


FIG. 12. Summer rain rate and 1000-hPa wind response to anomalous SST forcing under mean SST for the period of (a) 1962–77 and (b) 1978–93. The contour interval is 1 mm day^{−1} with zero contour suppressed. The wind scale is shown on the top-right of the panels. The results are based on 6-yr ensemble means using COLA AGCM.

SST forcing is enhanced under stronger mean low-level convergence and mean convection. The convection anomalies also extend to higher latitudes. The AGCM used in the numerical experiments is the Center for Ocean–Land–Atmosphere Studies (COLA) AGCM (Kirtman and DeWitt 1997) with a horizontal resolution of R30 and 19 levels in the vertical. Four experiments were designed. In the first two experiments, the specified SST forcing is climatological monthly mean SST derived from the period of 1962–77 and 1978–93, respectively. For simplicity, these two experiments are denoted as C1 and C2, respectively. In the last two experiments, we superimpose anomalous monthly mean SST derived from a composite of SST anomalies for the years of 1983, 1987, and 1992. These two experiments are annotated as C1E and C2E. The derived difference between the experiment with mean plus anomalous SST forcing and that with mean only SST forcing is taken as the response of the atmosphere to the anomalous SST forcing. For each of the four experiments, we made six integrations, each for 1 yr, starting from 1 January with different initial conditions. The results shown in Fig. 12 are ensembles of six integrations. Negative rain-rate anomalies over the WNP are larger in the experiment with mean SST forcing derived from the period of 1978–93 than from the period of 1962–77. So are low-level anticyclonic wind anomalies over the WNP. The anomalous precipitation and anticyclone also extend to higher latitudes under a higher mean SST.

Similar experiments were done with composite SST anomalies for the years of 1964, 1966, 1970, 1973, and 1977. The convection and anticyclone anomalies over WNP display similar sensitivity to mean SST change though the anomalies are weaker. We also did experi-

ments based on composite SST anomalies for La Niña cases. The WNP convection and cyclone anomalies, especially those to the east of 145°E, show a tendency to shift to higher latitudes under a higher mean SST. Compared to Fig. 12, the anomalies are much weaker and occur at 1–2 months earlier.

b. Causes for the change in the influence of the Indian convection

One plausible reason for the weakened influence of the Indian convection on the east Asian circulation is the reduction in the strength of anomalous Indian convection. Wang et al. (2001) indicated a decrease in the Indian summer monsoon variability after the late 1970s. The standard deviation of the AISR in 1978–93 (60 mm) is about 89% of that in 1962–77 (67 mm). The decreased Indian rainfall variability may be related to a weakened influence of ENSO on the Indian summer monsoon in recent decades, as indicated by the broken correlation between the AISR and simultaneous Niño-3 SST index (Kumar et al. 1999). The plausible reasons for the weakening Indian summer monsoon–ENSO relationship were discussed in Kumar et al. (1999) and Chang et al. (2001).

Another plausible reason is the change in the distribution of summer rainfall variability in homogeneous regions in India (for the distribution of homogeneous regions in India, refer to Munot and Kothawale 2000). We found that the decrease in summer rainfall variability mainly happens in west-central India (standard deviation decreases by about 24%). In peninsular India, central-northeast India, and northeast India the standard deviation of summer rainfall increases by about 24%, 11%, and 21%, respectively. The change in the distribution of summer rainfall variability can modify the circulation response over east Asia since the circulation anomaly induced by anomalous heating in different subregions of India may vary. Rodwell and Hoskins (1996) indicated that the atmospheric response depends critically on the latitudinal location of the heating imposed. The anticyclone generated northwest of India is much stronger and extends to higher latitudes when the heating is at 25°N, 90°E than at 10°N, 90°E.

The dependence of the circulation response on the location of anomalous Indian heating is demonstrated using Fig. 13, which shows the simultaneous correlation of summer 200-hPa height with summer rainfall in northwest, west-central, and peninsular India. Note that those regions have different latitudinal extent. Northwest India is between 20° and 32°N and west of 78°E. West-central India lies mainly between 15° and 25°N. Peninsular India is located mainly south of 16°N. The zonal mean part of the height anomaly has been removed before calculating the correlation in order to emphasize the wave-type feature. Apparently, the regions at high latitudes are more likely to induce wave-type response over Asia and have a larger influence on east Asia. The

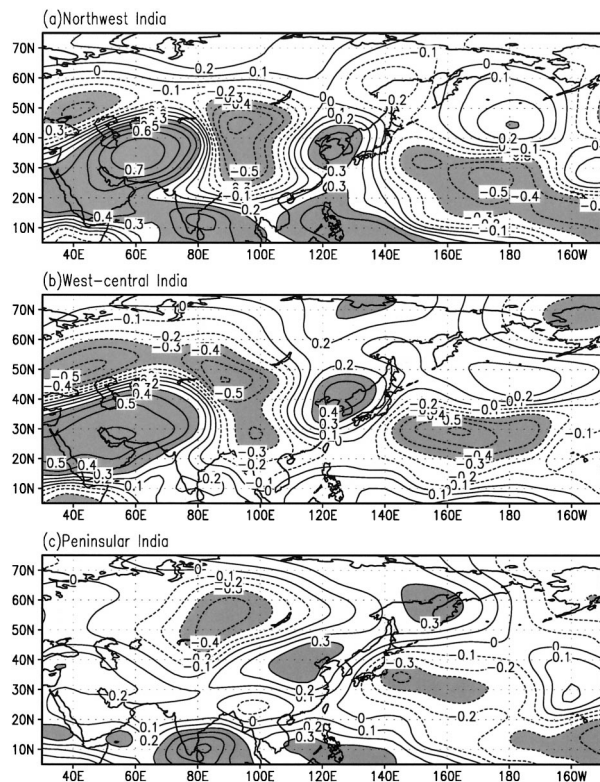


FIG. 13. Simultaneous correlation of summer 200-hPa height with summer rainfall in (a) northwest India, (b) west-central India, and (c) peninsular India for the period of 1948–94. Shading indicates regions where correlation coefficient is over 0.3. The contour interval is 0.1. The correlation is calculated using the zonal departure part of the interannual component only.

positive correlation over east Asia shifts westward in the correlation map with the peninsular Indian rainfall (Fig. 13c) compared to that with the northwest and west-central Indian rainfall (Figs. 13a,b). We also calculated the correlation of 200-hPa height with reference to 500-hPa vertical motion at different grids in the Indian monsoon region. The results suggest that the response of midlatitude circulation over Asia varies largely. When the reference grid is at 20°–25°N and over the Indian land, the correlation distribution shows a clear wave pattern over the midlatitudes of Asia. The correlation pattern is weaker when the reference grid is at a lower latitude (10°–15°N). This is reasonable since the circulation response depends on the distance from the heating to the midlatitude westerly and consistent with Rodwell and Hoskins (1996). This agrees with statistical relations presented in previous studies. Krishnan and Sugi (2001) showed that only in northern part of India the July–August rainfall has significant correlation with that in southern Japan. In the correlation field of rainfall in India with 500-hPa height at 45°N, 130°E during summer, significant correlation is found only in northern India (Kripalani et al. 1997).

c. Relative contribution of anomalous WNP and Indian convection

The relative contribution of anomalous WNP and Indian convection to the EASM variability may depend upon the sign, location, and intensity of the two heating regions. We examined ENSO-related summertime 500-hPa vertical p -velocity anomaly over India (10°–20°N, 75°–85°E) and the WNP (10°–20°N, 120°–160°E). It turns out that the anomalous convection over the WNP and India tends to be of opposite sign and has the same intensity in the decaying phase of ENSO during 1962–77. Considering the insignificance of 500-hPa height correlation over eastern China (Fig. 5a), the Indian heating may be more important. On the other hand, significant negative correlation of 850-hPa meridional wind and station rainfall (Fig. 4a) with the WNPPI over northeastern China indicates an additive contribution from the WNP heating, especially for the development of anomalous meridional winds over the Yellow Sea–northeastern China. It suggests that the ENSO influences the EASM through both anomalous WNP and Indian convection in the decaying phase of ENSO during 1962–77.

In the decaying phase of ENSO during 1978–93, anomalous convection is much stronger over the WNP than over India. As such, the response of the east Asian circulation to anomalous WNP heating is larger than that to anomalous Indian heating. The resemblance (opposite sign) of Fig. 5b to Fig. 3b and Fig. 4b to Fig. 1b supports the role of anomalous WNP convection. The weakness of the upper-level wave pattern over the mid-latitudes of Asia suggests that the influence of anomalous Indian convection on the EASM is weak. Thus, it is speculated that the influence of ENSO on the EASM is mainly through anomalous WNP heating in the decaying phase of ENSO during 1978–93.

6. Summary and discussion

Using station rainfall data and the NCEP–NCAR reanalysis, this study contrasts the correlation of the east Asian summer rainfall with the preceding winter Niño-3 SST anomaly between 1962–77 and 1978–93. A pronounced different correlation is found in northern China and Japan. In eastern north China, the correlation was strongly positive in 1962–77 and switched to negative in 1978–93. In central Japan, the correlation changed from significant and negative to weak. In the Huai River region and South Korea, the negative correlation experienced a similar weakening as that in central Japan. The correlation in the Yangtze River region did not show obvious change.

Consistent changes are identified in the summertime Asian circulation anomaly in the decaying phase of El Niño. In 1962–77 an anomalous barotropic anticyclone is centered over the Japan Sea. Anomalous southerly winds over the Yellow Sea and northeastern China bring

more moisture to eastern north China. Anomalous northeasterly winds over Japan reduce the precipitation. In 1978–93 an elongated anomalous cyclone extends from the North Pacific to the east Asian coast. Anomalous northeasterly winds over northeastern China bring more dry air and reduce the precipitation. At the upper levels, an obvious zonal wave pattern developed over the mid-latitudes of Asia along 30°–50°N in 1962–77. In 1978–93 the above wave pattern was weak. Instead, strong westerly anomalies exist over east Asia through the North Pacific along 25°–40°N. Over the WNP, an anomalous anticyclone is observed, but it is stronger and extends to higher latitudes in 1978–93 than in 1962–77. In the decaying phase of La Niña, circulation anomalies are opposite.

There are two ways by which ENSO can influence the summer rainfall anomaly over northern China and Japan. One is through anomalous WNP heating that induces a meridional wave pattern along the east Asian coast. The other is through anomalous Indian heating. It is proposed that the impact of anomalous Indian convection on the east Asian circulation may be through modulating a zonal wave pattern at upper levels along 30°–50°N via a baroclinic Rossby wave response to anomalous convective heating over India.

One cause for the contrasting circulation anomaly over east Asia in the decaying phase of ENSO before and after the late 1970s is the change in location and intensity of the WNP convection anomaly. This change leads to a very different influence of anomalous WNP convection on the east Asian circulation. Anomalous WNP convection becomes stronger after the late 1970s, thereby imposing a larger impact on the east Asian circulation. The anomalous convection over the WNP extends to higher latitudes in 1978–93 than in 1962–77. This results in an eastward shift of the anomalous low (high) pressure from east Asia to the North Pacific along 30°–45°N in summer in the decaying phase of El Niño (La Niña). The associated low-level anomalous winds over northeastern China change from southerly (northerly) in 1962–77 to northerly (southerly) in 1978–93.

Another cause for the change in the east Asian circulation anomaly is the weakening of the Indian monsoon variability after the late 1970s. This weakening is mainly due to the decrease in summer rainfall variability in west-central India. Due to the decrease in the Indian monsoon variability along with the change in spatial distribution of summer rainfall variability, the influence of anomalous Indian heating on the circulation anomaly over east Asia experienced a significant weakening after the late 1970s.

The change in location and intensity of anomalous WNP convection may relate to an increase in summer mean SST in the Philippine Sea. The higher mean SST induces a stronger mean convection and low-level mean convergence. Thereby, the convection anomaly over the WNP is enhanced and extends to higher latitudes. This effect is especially large in the decaying phase of ENSO.

The decrease in the Indian monsoon variability may be due to a weakened influence of ENSO on the Indian monsoon after the late 1970s. The latter may relate to the change in the frequency of ENSO.

The relative roles of the two heating sources are different before and after the late 1970s. In 1962–77 ENSO affects the EASM through both anomalous WNP and Indian heating. In 1978–93 the influence of ENSO on the EASM is mainly through changing the WNP heating. In the decaying phase of El Niño during 1962–77, the Indian summer monsoon is stronger than normal and the WNP summer monsoon is weaker than normal. Anomalous strong Indian convection induces a zonal wave pattern over the midlatitudes of Asia with an anomalous barotropic anticyclone centered over North Korea. At the same time, anomalous weak WNP convection also contributes to the development of southerly wind anomalies over the Yellow Sea and northeastern China. In the decaying phase of El Niño during 1978–93, the Indian summer monsoon is near normal. Thus, the impact of anomalous Indian convection is weak. On the other hand, the WNP summer monsoon is substantially weaker than normal. As such, anomalous WNP convection generates large circulation anomalies along the east Asian coast. However, the induced circulation anomalies over east Asia are very different from those in 1962–77 because of a shift of anomalous convection to higher latitudes. The most important difference is the eastward shift of the anomalous low pressure from east Asia to the North Pacific. This induces strikingly different wind anomalies over northern China and Japan: northerly over the Yellow Sea and northeastern China and northwesterly over central Japan.

The change in tropical mean SST in the interdecadal timescale influences both the evolution of the SST anomaly and the location and intensity of anomalous WNP convection. Its impact on the evolution of the SST anomaly relates to the difference of summer SST anomalies between the two epochs as discussed by Chang et al. (2000a). Its modification on the location and intensity of the WNP convection anomaly affects the circulation anomaly over east Asia. The change in tropical mean SST can also alter tropical mean convection. The latter could modulate the mean circulation over east Asia whose influence on the response of the east Asian circulation to tropical heating has not been discussed in this study.

The present study is mainly based on correlation analysis. Some of the correlation features in the east Asian region are not as significant as in the Tropics, partly because the influence of ENSO on the EASM is indirect and because other factors, for example, the Eurasian land surface condition, also affect the EASM variability. This imposes limitations on some of the results of the present study.

The relationship between the east Asian rainfall and eastern equatorial Pacific SST anomalies is complex. It displays seasonal dependency (Zhang et al. 1999), sub-

seasonal change (Chang et al. 2000a,b) meridional structure (Huang and Wu 1989; Shen and Lau 1995; Chang et al. 2000b), and interdecadal modulation (Chang et al. 2000a,b). More work is needed to document and understand the complexity of the east Asian monsoon–ENSO relationship. It is also interesting to explore the possible role of the change in midlatitude mean circulation in modulating this relationship since the east Asian monsoon is a subtropical monsoon that involves both tropical and midlatitude systems (Tao and Chen 1987). As the European land surface temperature change may play a role in the recent weakening of the Indian summer monsoon–ENSO relationship (Chang et al. 2001), another topic for future study is the interdecadal change in the Eurasian land surface condition and its possible impact on the east Asian monsoon–ENSO relationship.

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