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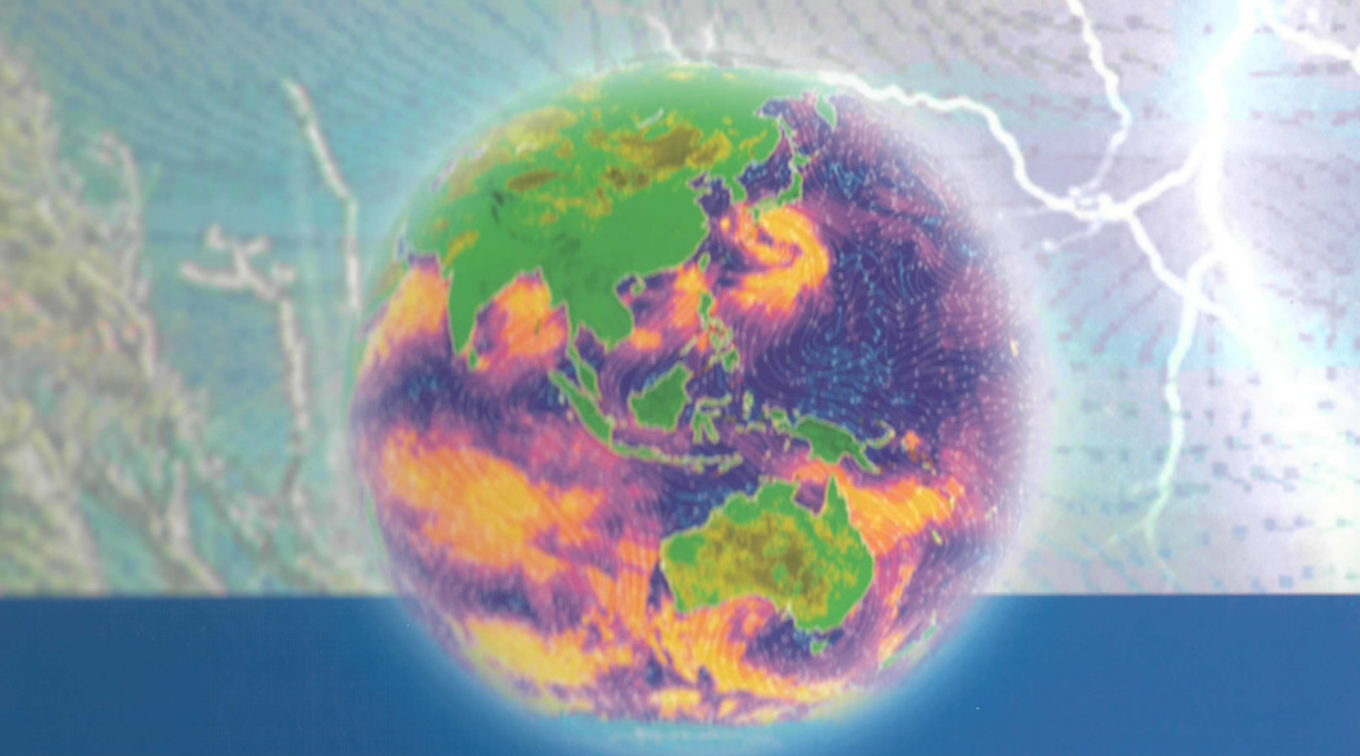
Vol. 2

East Asian Monsoon



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East Asian Monsoon

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East Asian Monsoon

edited by

C.-P. Chang

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PREFACE

The term "Asian summer monsoon" most frequently refers to the heavy summer rainfall season in South Asia, particularly around India. To some in the western world, "East Asian summer monsoon" is a less familiar term. This is partly because seasonal mean maps have a large area of high rainfall in South Asia during boreal summer, while rainfall over a comparably sized area in East Asia is significantly lower. In fact, the East Asian summer monsoon has complex space and time structures that are distinct from the South Asian summer monsoon. It covers both subtropics and midlatitudes and its rainfall tends to be concentrated in elongated rain belts that stretch for many thousands of kilometers and affect China, Japan, Korea, and the surrounding seas. The narrow meridional scale and the tendency of alternating stationary and northward progression stages make the seasonal averaged rainfall in East Asia less concentrated than in South Asia. However, the impacts of floods and droughts on human lives and economics during to East Asian summer monsoon are just as great because the finer intraseasonal space-time structure, coupled with the narrow rivers, is more sensitive to interannual variations. For example, a flood event can result from heavier-than-normal rainfall during the normal subseasonal rain period or from sustained rainfall outside of the normal period or both.

The East Asian summer monsoon is also closely linked with the West Pacific summer monsoon. Both are part of the global climate system and are affected by El Nino – Southern Oscillation (ENSO) and surface temperature variations in the western Pacific and surrounding oceans, the tropospheric biennial oscillation, and the South Asian summer monsoon. In addition, typhoons in the western North Pacific are most active during the East Asian summer monsoon. They may be considered a component of the East Asian summer monsoon as they contribute substantial amounts of rainfall and have major impacts on the region.

The East Asian winter monsoon, or "Asian winter monsoon", may be a more familiar weather system to the western world. Its circulation encompasses a very large meridional domain. Cold air outbreaks emanate from the Siberian high and penetrate deeply into the equatorial region, where the center of maximum rainfall is found in the Maritime Continent region. This center has long been recognized as a major planetary scale heat source that provides a significant amount of the energy that drives the global circulation during boreal winter. The East Asian winter monsoon is directly connected with the Australian summer monsoon by vigorous circulations extending across the equator, and both are affected by ENSO and other oscillations in the Pacific and Indian Ocean. In addition, a number of global teleconnections associated with the East Asian winter monsoon have recently been found. These include a correlation of temperature and precipitation anomalies over North America with fluctuations in the East Asian jet stream, and an association of winter storm development as far away as Europe with precursor convection over the Maritime Continent.

Because of its impacts on nearly one third of the world population and on the global climate system including effects on climate change, the study of East Asian monsoon has received increased attention in both East Asian countries and in the United States. This book presents reviews of the recent research on the subject. The book is organized into five parts with three chapters each: East Asian summer and winter monsoon, interannual variations, general circulation modeling, synoptic and mesoscale processes, and interactions with other circulations.

The idea of this book originated when the World Meteorological Organization/ Commission on Atmospheric Science Working Group on Tropical Meteorology Research, under the past chair, Dr. Greg Holland, and the present chair, Prof. Lianshou Chen, organized the International Panel for East Asian Monsoon under the leadership of Prof. Shiyan Tao. The book would not have been possible without the efforts of the many reviewers. Besides the members of the Editorial Board of the World Scientific Series on Meteorology of East Asia who served as reviewers, the book benefited tremendously from the following individuals who provided critical reviews of individual chapters: Drs. Hanna Annamalai, Johnny Chan, Russ Elsberry, Bill Frank, Bob Haney, Harry Hendon, Brian Hoskins, Dick Johnson, T. N. Krishnamurti, Bill Kuo, Bill Kyle, Mark Lander, Peggy LeMone, Jerry Meehl, Takio Murakami, Julia Slingo, Wei-Kuo Tao, Mingfang Ting, Chuck Wash, and Song Yang. The technical editing was performed by Ms. Hway-Jen Chen. This work was supported in part by the National Science Foundation under Grant ATM-0101135, the National Oceanic and Atmospheric Administration under Grant NA01AANRG0011, and the Office of Naval Research Marine Meteorology Program.

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V. Interactions with Other Circulations

6. CLIMATE VARIATIONS OF THE SUMMER MONSOON OVER CHINA

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The progress in recent studies on climate variations of the summer monsoon in China is reviewed in this paper. Especially, the characteristics of the intraseasonal, interannual and interdecadal variations of the summer monsoon in China and its surrounding regions, which can cause severe drought and flooding disasters there, are discussed from the analyses of observed data. Moreover, some causes and physical processes of these variations are also sought from the components of the East Asian climate system. Besides, the scientific problems which need to be studied in the near future are suggested.

1. Introduction

China is in East Asia, which is one of the strong monsoon regions in the world. There are many characteristic weather systems in different seasons, such as the Meiyu in China, the Baiu in Japan, the Changma in Korea, tropical cyclones in summer, and persisting northwesterly and northeasterly winds and cold surges in winter. Climate variations including interdecadal, interannual, and intraseasonal variabilities of the summer monsoon over China and its surrounding regions have been found to be significant. And due to these variations, serious climatic disasters such as extensive droughts and floods frequently occur there. Since the 1980s, these climatic disasters have become more severe and caused great damage to agricultural and industrial production and to the daily life of people in China. In each year of the 1980s and the 1990s, the climatic and synoptic disasters caused grain loss of about 20 billion kg and economic loss of about 200 billion RMB (Chinese yuan), approximately 3% - 6% of the GNP of China. The extremely severe flood that occurred in the Yangtze River valley in the summer of 1998 caused an economic loss of over 260 billion RMB (e.g., Huang *et al.* 1998; Huang and Zhou 2002). Thus, it is an important issue for Chinese meteorologists to study the regularity and causes of climate variations of the summer monsoon over China, so that seasonal prediction of climate variations in China can be improved and the predictability of interannual variations of climate in China can be understood further.

Since the East Asian monsoon has a great impact on climate in China and its surrounding regions, more than sixty years ago, Zhu (1934) first proposed the possible relationship between summer rainfall in China and the East Asian summer monsoon. Later on, Tu and Huang (1944) investigated the advance and retreat of the Asian summer monsoon in East Asia. These studies opened the way to search for the regularity of the East Asian monsoon variability and its impact on climate variations in China. Following Tu and Huang's (1944) investigation, many Chinese meteorologists (e. g., Tao and Chen 1957; Yeh *et al.* 1959; Gao and Xu 1962) studied the characteristics of monsoon climate in China and the influence of the onset and retreat of the East Asian summer monsoon on climate in China. Tao and Chen (1987) made a systematic review on these studies. After their review, studies on the regularity and causes of different time scale climate variations of the summer monsoon over China have been greatly made using observed data for a longer period. For example, many scholars (e. g., Huang and Li 1987; Chen *et al.* 1991; Ding 1992; Huang *et al.* 1993; Huang and Sun 1994b) investigated the intraseasonal variations of the summer monsoon over China. Recently, many scholars also studied the regularity and causes of interannual variations of the summer monsoon over China (e.g., Miao and Lau 1990; Lau and Shen 1992; Huang and Sun 1992; Huang and Sun 1994a; Chang *et al.* 2000; Huang *et al.* 2003). In addition, the decadal and interdecadal variations of the summer monsoon over China and its surrounding regions have attracted many researchers' interest (e.g., Wang 1993; Nitta 1993; Huang *et al.* 1999; Chang *et al.* 2000; Huang 2001).

Since the late 1970s, many investigations emphasized the thermal effect of the Tibetan Plateau on the interannual variations of the summer monsoon over China and its surrounding regions (e.g., Ye and Gao 1979; Nitta 1983; Luo and Yanai 1984; Huang 1984, 1985). The thermal effect of the West Pacific warm pool on the interannual variations of the summer monsoon over East Asia have also been investigated by many scholars (e.g., Nitta 1987; Huang and Li 1987; Kurihara 1989; Huang and Sun 1992).

Since observed data for relatively short periods were used in previous studies of climate variations of the summer monsoon over China and its surrounding regions, it is necessary to re-examine the summer monsoon climate variabilities using longer datasets. Thus, the observed datasets of monthly precipitation and surface air temperatures at 160 observational stations of China from 1951 to 2000 and daily precipitation at 598 stations of China from 1961 to 2000, daily snow cover at 72 stations in the Tibetan Plateau, provided by the National Climate Center (NCC) of the China Meteorological Administration (CMA), and the NCEP/NCAR reanalysis data of height, moisture, and wind fields, the observed high cloud amount published by the Monthly Report on Climate System, JMA, and the dataset of GISST, provided by the Hadley Center, UK are used to analyze the intraseasonal, interannual, and interdecadal variabilities of the summer monsoon over China in this chapter. Moreover, recent studies of climate variations of the summer monsoon over China and the relationship between these variabilities and the East Asian climate system including the East Asian monsoon, the western Pacific subtropical high and the disturbances in the middle latitudes in the atmosphere, the thermal effect of the West Pacific warm pool, the ENSO cycle in the tropical Pacific, the land surface processes of the Eurasian continent, and the thermal and dynamical effects of the Tibetan Plateau are also reviewed in this chapter.

2. Intraseasonal Variability of the Summer Monsoon over China

2.1. *The Summer Monsoon Climate over China*

2.1.1. Precipitation

In order to study climate variations of the summer monsoon over China, the climatological mean monthly precipitation and surface air temperature over China averaged for 30 years from 1961 to 1990 are analyzed by using observed monthly precipitation and surface air temperature at 160 stations of China, whose locations are shown in Figure 2.1. Figures 2.2a-f are the distributions of climatological monthly precipitation in April, May, June, July, August, and September, respectively. These figures feature the zonal characteristics in rainfall distributions, which are known as the rainband in China, Japan and Korea. The rainband exhibits a quasi-stationary state during a period and moves from South China to North China from April to August. In April and May, large precipitations of about 250 mm are mainly concentrated in South China and to the south of the Yangtze River. This period is called the

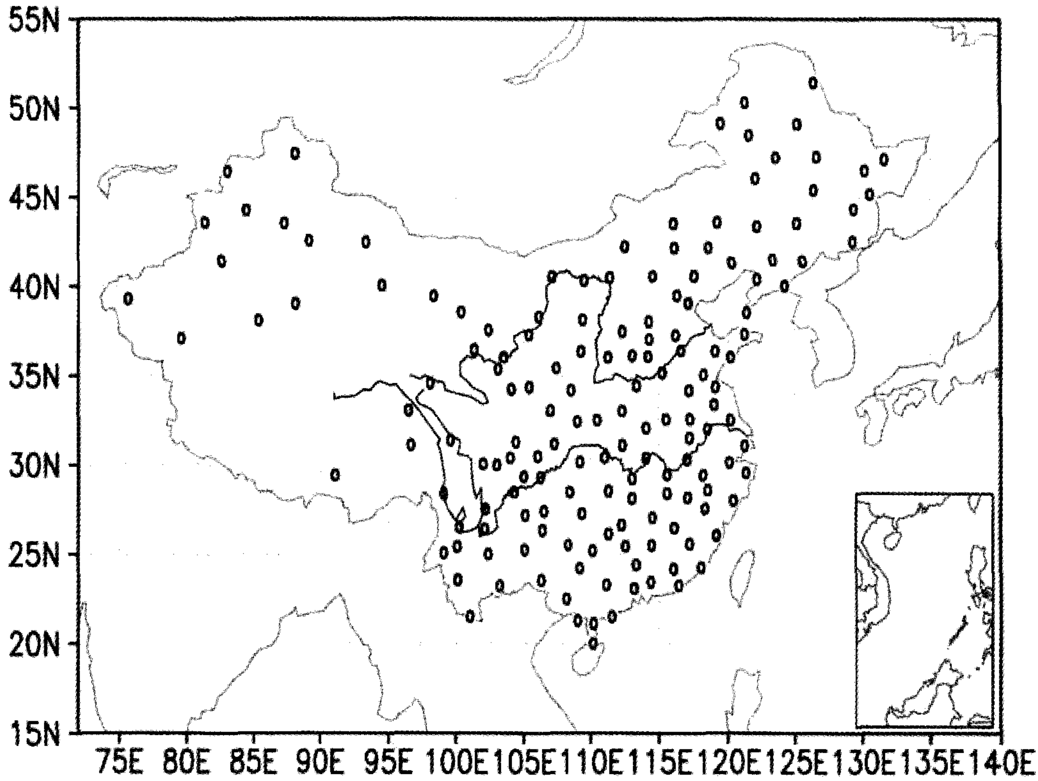


Figure 2.1. Distribution of 160 meteorological observation stations in China.

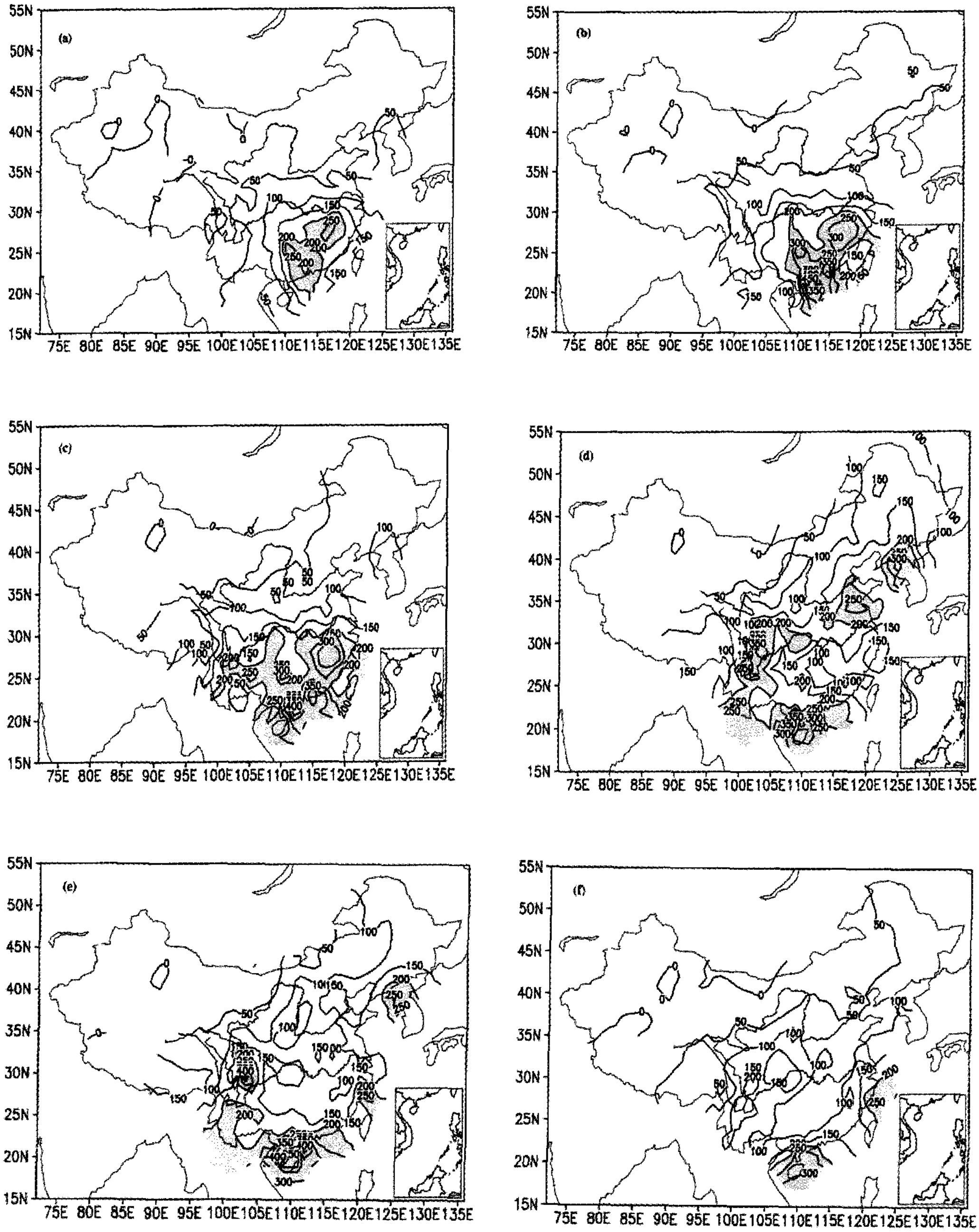


Figure 2.2. Distributions of climatological monthly precipitation averaged for 30 years from 1961 to 1990. Units: mm. (a) April; (b) May; (c) June; (d) July; (e) August; (f) September. The areas of monthly precipitation amount over 200 mm are shaded, and contour interval is 50 mm in these figures.

pre-rainy season of South China, but the precipitation is small in North China and Northeast China in the period. From June, precipitation begins to increase in the middle and lower reaches of the Yangtze River (i.e., the area denoted by "3" in Fig. 3.1) and the Yangtze River and Huaihe River valley (see Fig. 3.1, i.e., the Jianghuai valley in Chinese. This valley is the area between the Yangtze and Huaihe Rivers and the area denoted by "2" in Fig. 3.1), and the monthly precipitation can reach about 250 mm in these regions in June and July, respectively. This period is called the Meiyu season in China. From July to August, precipitation begins to increase in North China and Northeast China, and large amounts of precipitation totally about 150-200 mm are mainly concentrated in the upper reaches of the Yangtze River, North China and Northeast China. From September, the areas of large precipitation retreat southward and are located in the upper and lower reaches of the Yangtze River and South China.

From these distributions of climatological monthly precipitation, it is clearly seen that the belt of large precipitation moves northward from April to August, which is schematically depicted in Fig. 2.3, but the rainband moves southward from September. This may be closely associated with the advance and retreat of the East Asian summer monsoon (EASM) (e.g., Tu and Huang 1944).

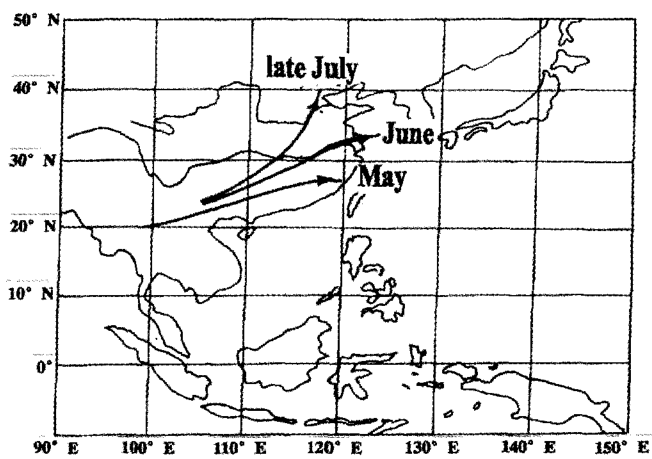


Figure 2.3. Schematic diagram of the locations of the summer monsoon rainband over East Asia.

2.1.2. *Surface Air Temperature*

Figures 2.4a-f are the distributions of climatological monthly mean surface air temperature in April, May, June, July, August and September, respectively. These figures have the same characteristics as the distributions of monthly precipitation shown in Figs. 2.2a-f. In April and May, the area of higher surface air temperature is mainly over South China and to the south of the Yangtze River. This area abruptly extends northward to North China and Northeast China from June and is maintained in the eastern part of China in July and August. However, the area of higher surface air temperature retreats southward to the Yangtze River valley and South China from September. This may be associated with not only the solar radiation, but also the retreat of the EASM.

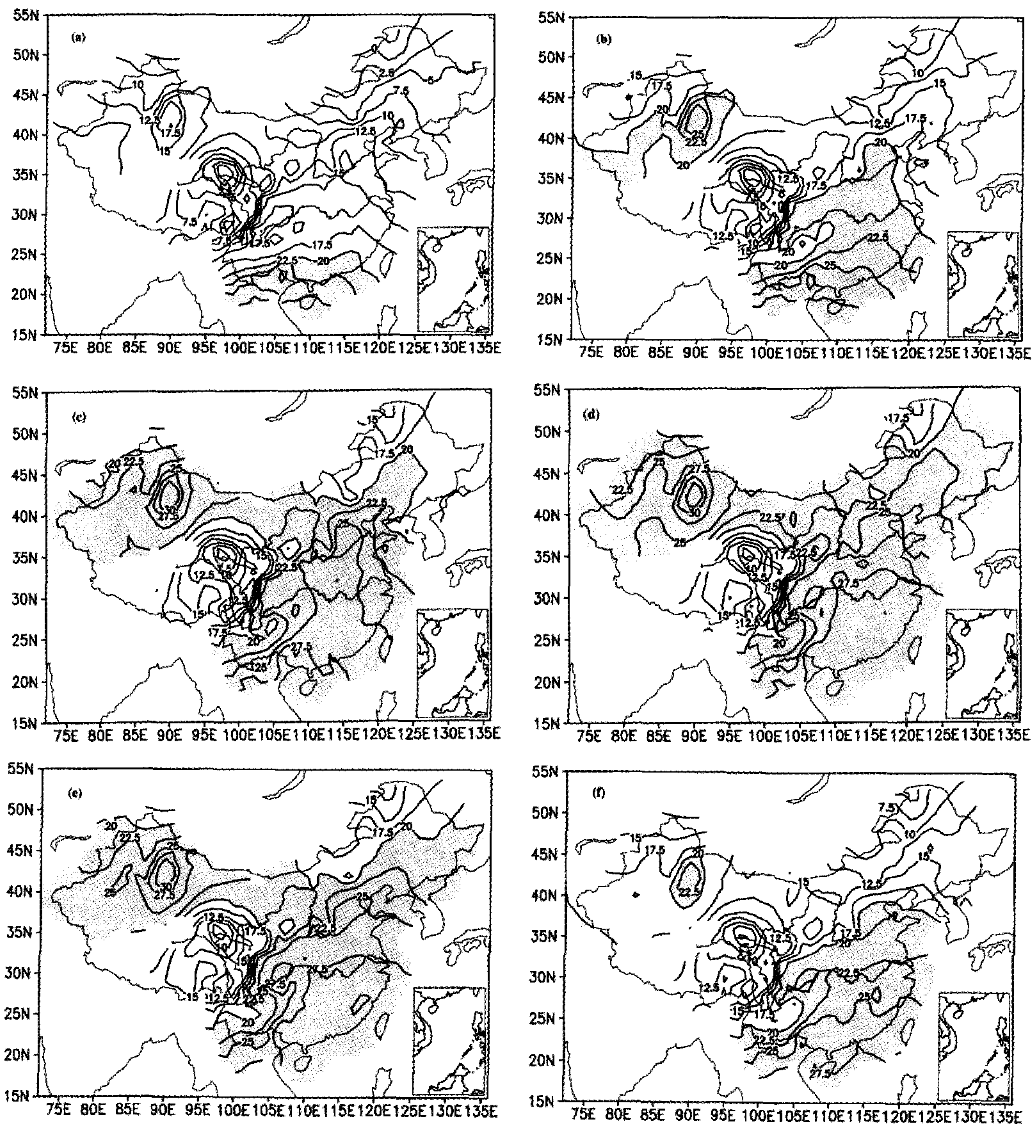


Figure 2.4. Distributions of climatological monthly mean surface air temperature for 30 years from 1961 to 1990. Units: °C. (a) April; (b) May; (c) June; (d) July; (e) August; (f) September. The areas of monthly mean surface air temperature over 20°C are shaded, and contour interval is 2.5°C in these figures.

2.2. *Intraseasonal Variability of the Summer Monsoon Rainband over East Asia*

Perhaps the variation of the monsoon rainband over East Asia from May to August is the most important intraseasonal variability of the EASM. In order to analyze well the intraseasonal variations of the monsoon rainband over China, a detailed dataset of precipitation is needed. Therefore, the observed dataset of daily precipitation at 598 stations in China, provided by NCC/CMA, is used to analyze the intraseasonal variability of the summer monsoon rainband over East Asia. However, considering the spatial distribution and the time length of the establishment of these observational stations, only the observed data at 255 stations are selected from the dataset of daily precipitation at 598 stations to be used in this study. Figure 2.5 is the latitude-time cross section of 5-day precipitation along 115°E (averaged between 110°E - 120°E) averaged for 30 years from 1961 to 1990 by using the observed daily precipitations at 255 observational stations of China. The intraseasonal variability of the summer monsoon rainband can be clearly seen from Fig. 2.5. As shown in the figure, the monsoon rainband is located over the area to the south of the Yangtze River during the period from May to early June, and then it moves abruptly northward and is located over the Yangtze River and Huaihe River valleys of China in mid-June. This is the

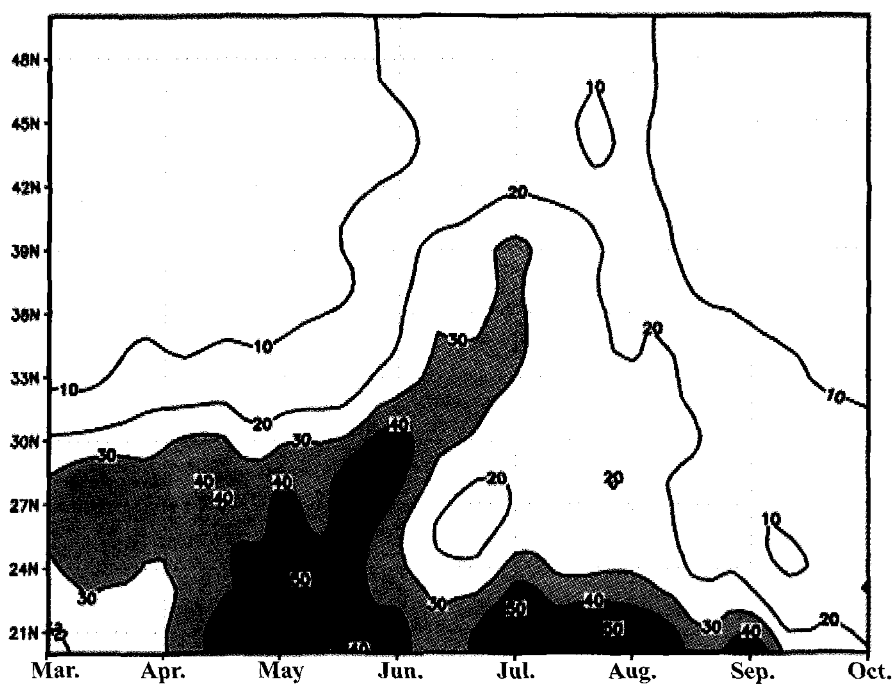


Figure 2.5. Latitude-time cross section of 5-day precipitation along 115°E (averaged between 110°E - 120°E) averaged for 30 years from 1961 to 1990. Units: mm. The precipitation amounts over 30 mm are shaded, and contour interval is 10 mm in the figure.

beginning of the Meiyu season in China (i.e., the Changma season in South Korea and the Baiu season in Japan). Moreover, the rainband again moves abruptly northward to North China in mid-July. Therefore, the northward movement of the maximum rainfall area shown in Fig. 2.5 is in agreement with the intraseasonal variations of the summer monsoon rainband shown schematically in Fig. 2.3.

The abrupt movement of the rainband from the south to the north is closely associated with the abrupt northward shift of the western Pacific subtropical high (e. g., Huang and Sun 1992; Ding 1992). Yeh *et al.* (1959) first discovered that the abrupt movement of the summer rainband is closely associated with the abrupt change of the large scale circulation over East Asia during early and mid-June. This abrupt change of planetary-scale circulation will bring the onset of the EASM in the Yangtze River and Huaihe River valley. Later on, Krishnamurti and Ramanatahan (1982), and McBride (1987) also pointed out the abrupt change of the Indian and Australian summer monsoon circulations.

The transition from the winter to the summer monsoon circulation and the northward movement of the monsoon rainband over East Asia are abrupt in the climatological mean sense (see Yeh *et al.* 1959). However, it is different in different years and is dependent on the thermal state of the tropical western Pacific, i.e., dependent on the SST anomaly in the surface and subsurface of the tropical western Pacific. Thus, this is discussed in the following two cases.

2.3. *Relationship between the Intraseasonal Variations of the Summer Monsoon Rainband and Convective Activities around the Philippines*

As well known, the tropical western Pacific is a region of the highest SST in the global sea surface and is known as "the warm pool". Due to the warm state of this region, the air-sea interaction is very strong, and the ascending branch of the Walker circulation is in the region. Thus, the strong convergence of the air and moisture leads to strong convective activities and heavy rainfall there.

Nitta's (1986) study showed that the convective activities over the tropical western Pacific are closely associated with the sea surface temperature anomaly of the West Pacific warm pool. When the West Pacific warm pool is in the warming state, i.e., the SST anomaly in the area of NINO-west (EQ - 14°N, 130°E - 150°E) is positive, the convective activities are strong around the Philippines, and vice versa. Therefore, the thermal state of the West Pacific warm pool can be described using the convective activities around the Philippines because the SST anomalies in the tropical western Pacific are relatively small. Moreover, Nitta (1987), Huang and Li (1987), Kurihara (1989), Tsuyuki and Kurihara (1989), and Huang and Sun (1992) all showed a close relationship between the summer monsoon over East Asia and the convective activities around the Philippines. Huang and Sun (1992) and Huang *et al.* (1993) showed that the intraseasonal variations of the summer monsoon over China during the summers with strong convective activities around the Philippines are different from those during the summers with weak convective activities in this region. In a summer with strong convective activities around the Philippines, the abrupt northward shift of the summer monsoon rainband from the Yangtze River valley to the Yellow River valley is obvious in early or mid-July, and the summer monsoon rainfall may be weak in the Yangtze River and Huaihe River valley of China, South Korea and Japan. On the other hand, in a summer with weak convective activities around the Philippines, the abrupt northward shift of the summer monsoon rainband from the Yangtze River valley to the Yellow River valley may be not obvious, and the summer monsoon rainfall may be heavy in the Yangtze River and Huaihe River valley of China, South Korea and Japan. Therefore, in the following, the intraseasonal variability of the summer monsoon over China is discussed by using the observed data for a longer period.

2.3.1. In the Summers with Strong Convective Activities around the Philippines

Figure 2.6 is the interannual variations of normalized high cloud amount anomaly averaged for spring (March-May), summer (June-August), autumn (September-November) and winter (December-February) around the Philippines (i.e., 10°N - 20°N, 110°E - 140°E). If the normalized high cloud amount anomaly averaged for a summer season (June-August) is positive around the Philippines, the convective activities may be considered as a case of strong convective activities in the summer (e.g., Monthly Report on Climate System, JMA). By this way, it can be seen from Fig. 2.6 that in the summers of 1978, 1981, 1985, 1988, 1994 and 1999, the convective activities were strong around the Philippines. On the other hand, if the normalized high cloud amount anomaly averaged for a summer is negative around the Philippines, the convective activities may be considered as a case of weak

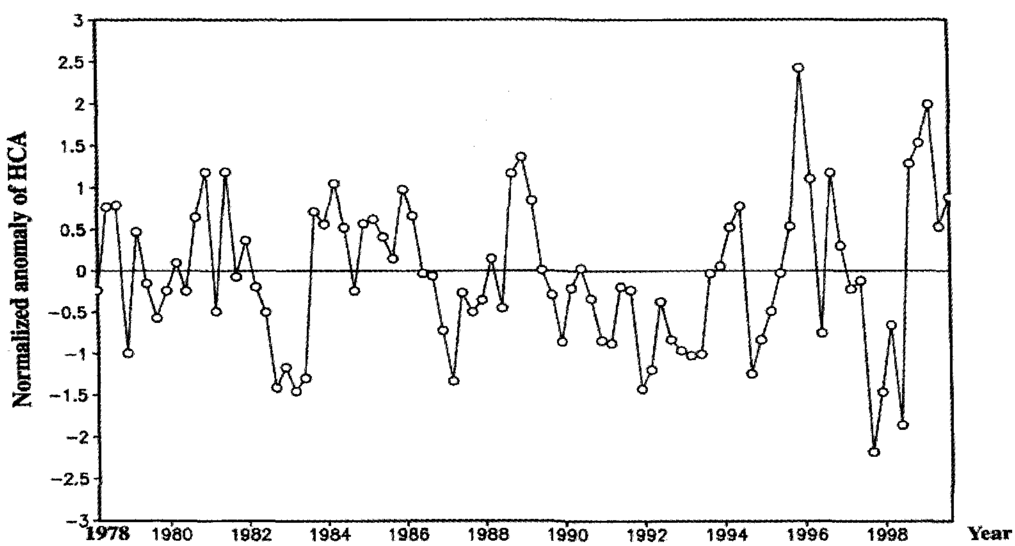


Figure 2.6. Interannual variations of the normalized seasonal mean high cloud amount (HCA) around the Philippines (i.e., 10°N - 20°N, 110°E - 140°E). The seasons mentioned in this analysis include spring (March-May), summer (June-August), autumn (September-November) and winter (December-February). Data are obtained from Monthly Report on Climate System, JMA.

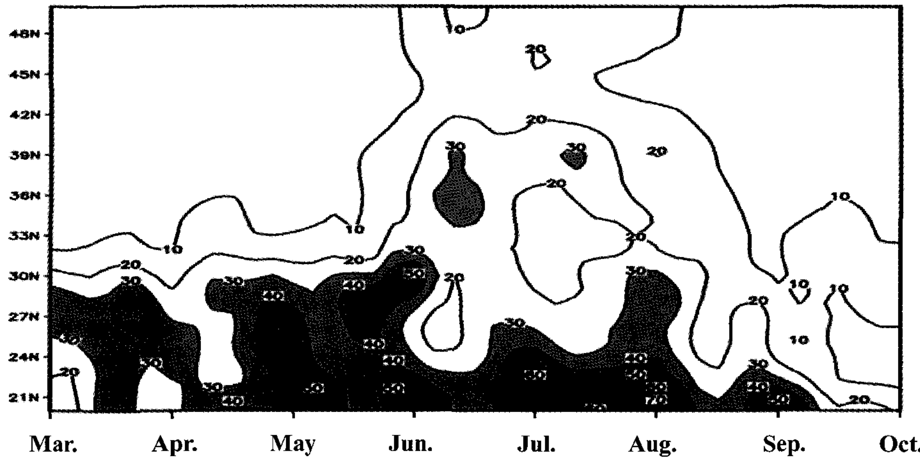


Figure 2.7. The composite latitude-time cross section of 5-day precipitation along 115°E (averaged between 110°E - 120°E) for the summers with strong convective activities around the Philippines (i.e., 10°N - 20°N, 110°E - 140°E). Units: mm. The precipitation amounts over 30 mm are shaded, and contour interval is 10 mm in the figure.

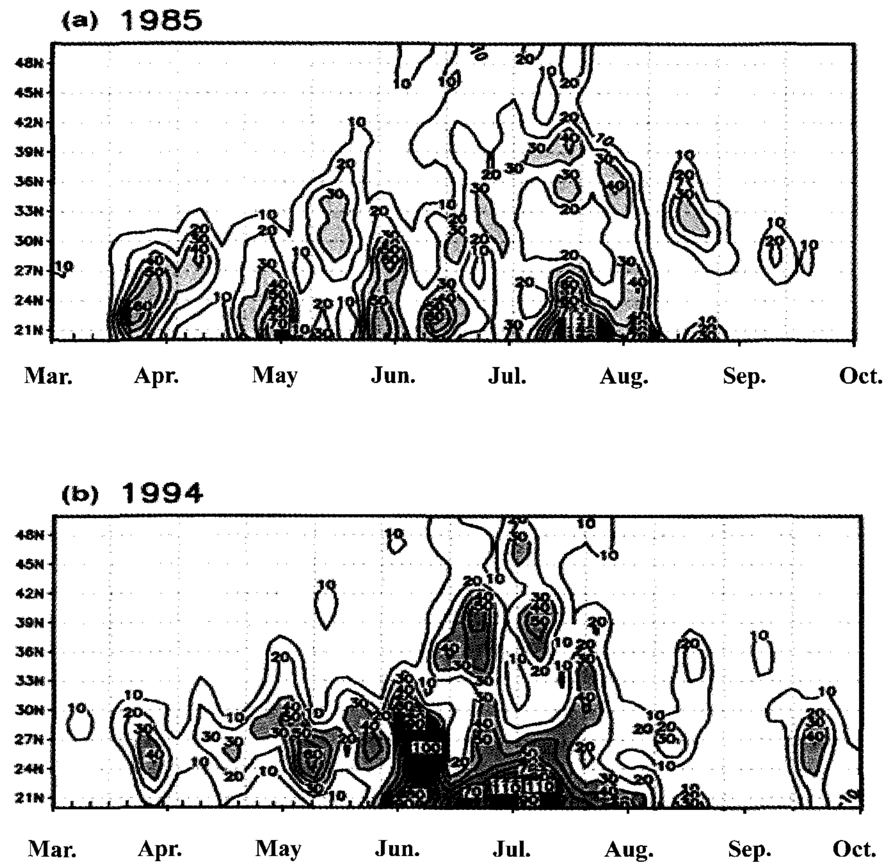


Figure 2.8. The latitude-time cross section of 5-day precipitation along 115°E (averaged between 110°E - 120°E) in the summers of (a) 1985 and (b) 1994. Units: mm. The precipitation amounts over 30 mm are shaded, and contour interval is 10 mm in the figure.

convective activities in the summer. It may be also seen from Fig. 2.6 that in the summers of 1980, 1982, 1983, 1991, 1992, 1993, 1996 and 1998, the convective activities were weak around the Philippines, and in this case, the convective activities were intensified over the equatorial central Pacific (e.g., Huang *et al.* 1998).

In order to discuss the different characteristics of the intraseasonal variability of the summer monsoon over China, the composite latitude-time cross sections of monsoon rainfall for the strong and weak convective activities around the Philippines are analyzed respectively.

Figure 2.7 is the composite latitude-time cross section of 5-day precipitation along 115°E (averaged between 110°E - 120°E) for the summers with the strong convective activities around the Philippines. Fig. 2.7 features the abrupt northward-shift of the summer monsoon rainband to the Yangtze River and Huaihe River valley in mid-June and then to the Yellow River valley in mid-July. To illustrate further this feature, the intraseasonal variations of the summer monsoon rainfall in the summers of 1985 and 1994 are taken as examples and shown in Figures 2.8a and 2.8b, respectively. It may be seen from Figs. 2.8a and 2.8b that during May and early June, the monsoon rainband was maintained over the area to the south of the Yangtze River, and the rainband abruptly moved northward to the Yellow River valley in mid-July. In these two summers, the monsoon rainfalls were very weak and were 30-50% below normal in the Yangtze River and Huaihe River valley. Due to the anomalous northward movement of the rainband, hot and dry summers occurred in the Yangtze River and Huaihe River valley in 1985 and 1994.

2.3.2. In the Summers with Weak Convective Activities around the Philippines

As shown in Fig. 2.6, in the summers of 1980, 1982, 1983, 1987, 1991, 1992, 1993, 1996 and 1998, the convective activities were weak around the Philippines, and the convections were

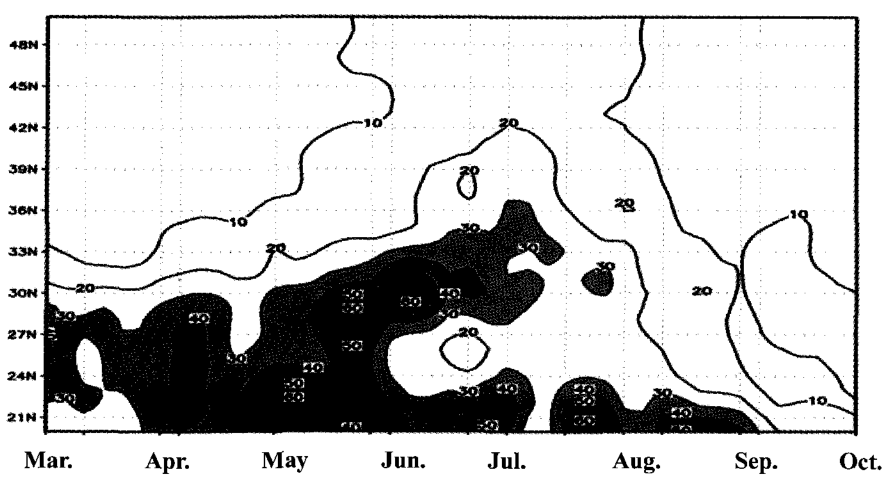


Figure 2.9. As in Fig. 2.7 except for the summers with weak convective activities around the Philippines (i.e., 10°N - 20°N, 110°E - 140°E). Units: mm.

intensified over the equatorial central Pacific in this case (e.g., Huang *et al.* 1998). Figure 2.9 is the composite latitude-time cross section of 5-day precipitation along 115°E (averaged between 110°E - 120°E) for the summers with weak convective activities around the Philippines. The figure shows that the northward shift of the summer monsoon rainband was not abrupt, and the summer monsoon rainband was maintained for a longer period over the Yangtze River and Huaihe River valley. In this case, the intraseasonal variability of summer rainfall is not remarkable in the eastern part of China.

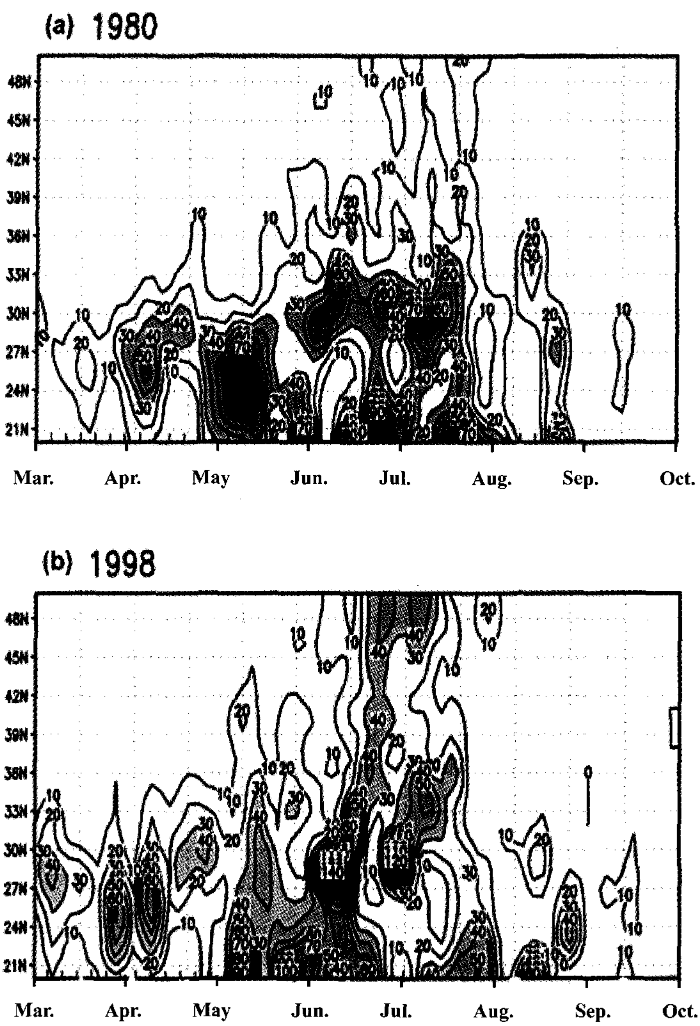


Figure 2.10. As in Fig. 2.8 except for the summers of (a) 1980 and (b) 1998.

Similarly, the summers of 1980 and 1998 are taken as examples, and the intraseasonal variations of rainfall in these two summers are shown in Figs. 2.10a and 2.10b. It can be seen that in these two summers, the abrupt northward shift of the monsoon rainband from the

Yangtze River valley to North China was not obvious and the monsoon rainband was maintained over the Yangtze River and Huaihe River valley from late May to late July. In the summer of 1980, the precipitation anomaly percentage was 30% above normal in the Yangtze River and Huaihe River valley and severe flooding occurred in this region. And in the summer of 1998, the precipitation was double normal amount and particularly severe flooding also occurred in the Yangtze River valley, as shown in Figure 2.11.

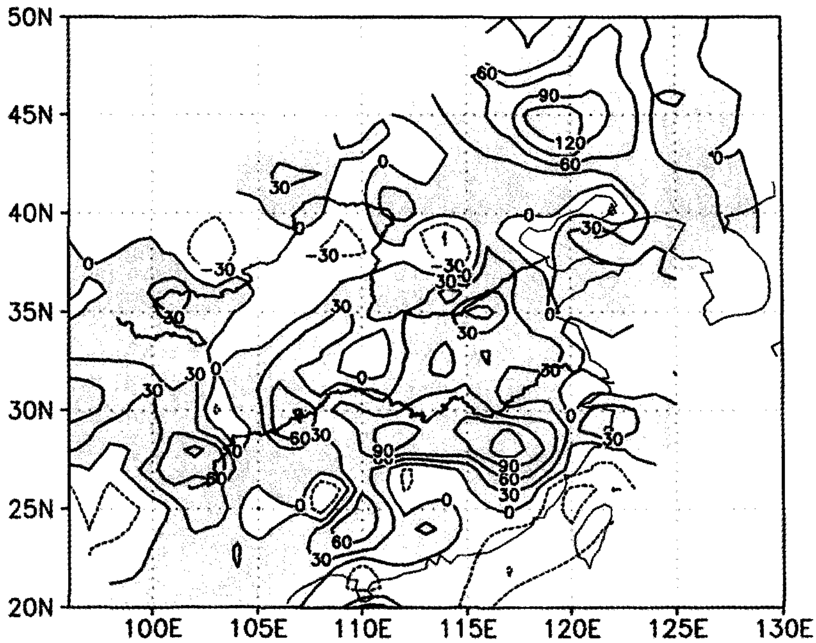


Figure 2.11. Distribution of monsoon rainfall anomaly in percentage in China in the summer (June-August) of 1998. The solid and dashed contours denote positive and negative anomalies of rainfall, respectively. The areas of positive rainfall anomalies are shaded, and contour interval is 30% in the figure. The distributions of climatological monthly precipitation in June, July and August shown in Figs. 2.2 c-e are taken as the normals, respectively.

2.4. *Relationship between the Intraseasonal Variations of the Western Pacific Subtropical High and Convective Activities around the Philippines*

Since the intraseasonal variations of summer monsoon rainfall over China are closely associated with the intraseasonal evolution of the western Pacific subtropical high, it is necessary for the further understanding of the causes of intraseasonal variations of summer monsoon rainfall over China to discuss the characteristics of the intraseasonal evolution of the western Pacific subtropical high in different cases of convective activities around the Philippines.

2.4.1. In the Summers with Strong Convective Activities around the Philippines

As described above, in the summers with strong convective activities around the Philippines, i.e., the summers of 1978, 1981, 1985, 1988, 1994 and 1999, the abrupt northward movements of the summer monsoon rainband in either mid-June or mid-July were obvious over China, as shown in Fig. 2.7. This is closely associated with the abrupt variation of the summer monsoon circulation system, especially the abrupt northward shift of the western Pacific subtropical high in mid-June over East Asia. Figure 2.12 is the composite distribution of latitude-time cross section of the 500 hPa height along 135°E for the summers with strong convective activities around the Philippines. It can be seen from the figure that the ridge line of the western Pacific subtropical high shifts abruptly from 18°N to 25°N in mid-June, which will cause the abrupt movement of the summer monsoon rainband to the Yangtze River and Huaihe River valley. Moreover, the subtropical high abruptly shifts northward again in mid-July and its ridge moves to 33°N, which will cause the abrupt shift of the summer monsoon rainband to the Yellow River valley and North China and the close of the Meiyu season in the Yangtze River and Huaihe River valley.

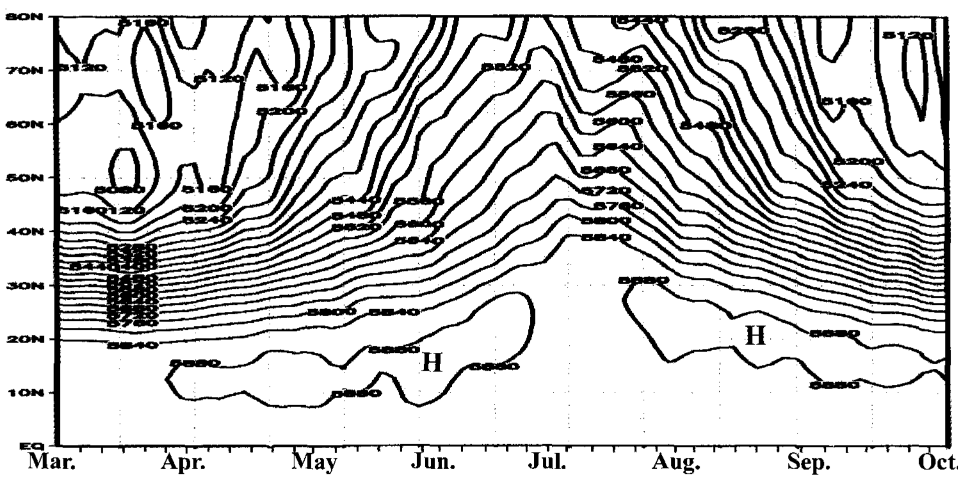


Figure 2.12. The composite distribution of latitude-time cross section of the 500 hPa height along 135°E for the summers with strong convective activities around the Philippines (i.e., 10°N - 20°N, 110°E - 140°E). Units: gpm. Contour interval is 20 gpm in the figure.

2.4.2. In the Summers with Weak Convective Activities around the Philippines

In the summers with weak convective activities around the Philippines, i.e., the summers of 1980, 1982, 1983, 1987, 1991, 1992, 1993, 1996 and 1998, the abrupt northward movements of the summer monsoon rainband were not remarkable over China either in mid-June or in mid-July, and the monsoon rainband stayed easily in the Yangtze River and Huaihe River valley, as shown in Fig. 2.9. This mainly is due to the fact that the western Pacific subtropical high remained over the region to the south of the Yangtze River. Figure 2.13 is the composite distribution of latitude-time cross section of the 500 hPa height along 135°E for the summers with weak convective activities around the Philippines. As shown in the figure, the abrupt northward shifts of the subtropical high are not obvious in either mid-June or mid-July, as it slowly moves northward and its ridge line remains at about 20°N from early June to mid-July.

From the above-mentioned analyses, it may be shown that the character of the intraseasonal variability of the western Pacific subtropical high in a summer with strong convective activities around the Philippines is different from that in a summer with weak convective activities around the Philippines. What causes this difference? Recently, Cao *et al.* (2002) applied the theory of multiple equilibrium proposed by Charney and Devore (1979) to the study of the nonlinear evolution of the western Pacific subtropical high. Their result showed that whether the northward shift of the subtropical high in mid-June is abrupt or not, it may depend on the thermal forcing due to convective activities around the Philippines. If the thermal forcing by convective activities around the Philippines is strong and exceeds a critical value, due to the strong wave-mean flow interaction and wave-wave interactions among the waves responding to the thermal forcing, the abrupt transition from a winter circulation pattern to a summer pattern can occur over East Asia in mid-June. In this case, the western Pacific subtropical high can abruptly shift northward. In contrast, if the thermal forcing by convective activities around the Philippines is weaker, since the wave-mean flow interaction and wave-wave interactions among the waves responding the thermal forcing are very weak, the transition from a winter circulation pattern to a summer pattern slowly occurs over East Asia, and the subtropical high oscillates with the thermal forcing oscillation.

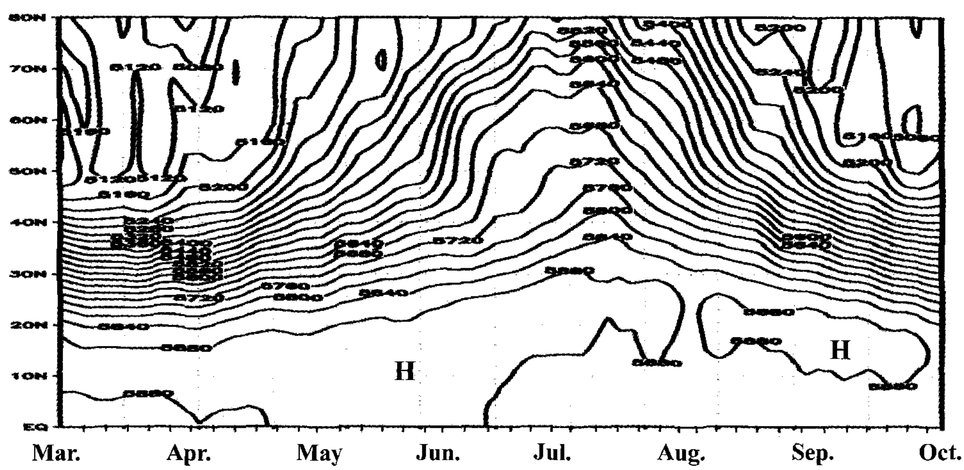


Figure 2.13. As in Fig. 2.12 except for the summers with weak convective activities around the Philippines (i.e., 10°N - 20°N, 110°E - 140°E).

Besides, the influence of convective activities around the Philippines on the intraseasonal variability of the summer monsoon over China and its surrounding regions may be associated with the Madden-Julian Oscillation. Huang (1994), Sun and Huang (1994) showed the intraseasonal variability of the summer monsoon rainfall in East Asia is closely associated with the 30-60 day oscillation propagated from the tropical eastern Indian Ocean and the Indo-China Peninsula to the East Asian summer monsoon region.

3. Interannual Variability of the Summer Monsoon over China

The summer climate in China, Korea and Japan is mainly influenced by the EASM, and summer rainfall anomaly in East Asia is also an important criterion measuring the interannual variability of the EASM (e. g., Tao and Chen 1987; Huang *et al.* 2003). Therefore, the interannual variability of the summer monsoon rainfall in China is emphasized in this section. However, the EASM is influenced by not only the Indian monsoon, but also the western Pacific subtropical high (e. g., Tao and Chen 1987; Huang and Sun 1992), thus, the interannual variations of the summer monsoon over China are complex (e. g., Huang *et al.* 2003).

3.1. *Interannual Variability of Summer Monsoon Rainfall in China*

Because China is located over the regions from the tropics to the temperature zone, and the eastern part of China is faced with the Pacific and the Tibetan Plateau is in the western part of China, the interannual variations of summer rainfall and surface air temperature in various regions of China are very different. Considering the characteristics of climate variability, China is generally divided into about 7 climate regions as shown in Figure 3.1 in the study of climate variability over China. The division of these climate regions of China and the number of observational stations used in the study of the intraseasonal and interannual variations of summer rainfall in these regions are shown in Table 3.1.

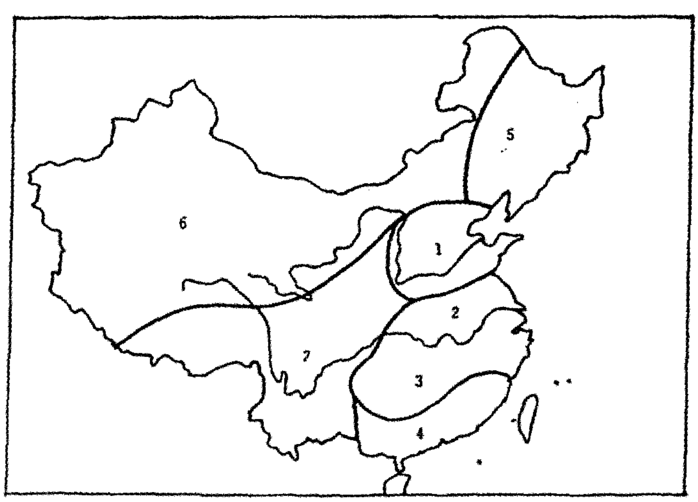
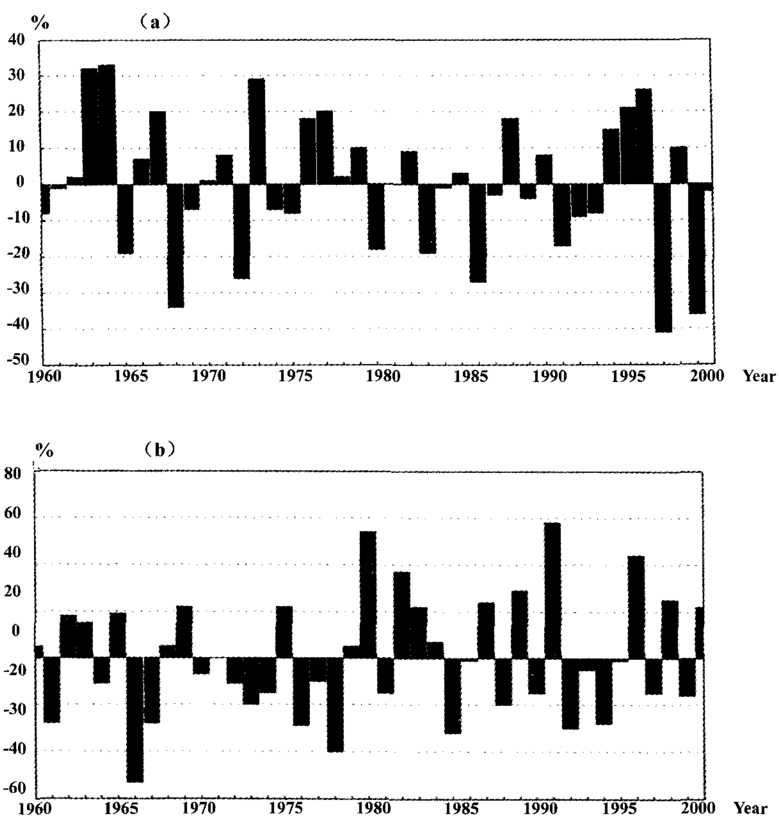


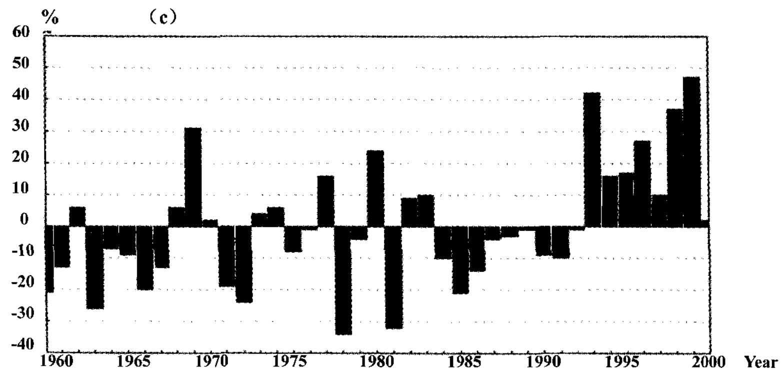
Figure 3.1. A division of climate regions of China.

Table 3.1 The division of climate regions of China and the number of observational stations in each region used in this study.

Region number	1	2	3	4	5	6	7
Name of Region	North China	The Yangtze River and Huaihe River valley	The middle and lower reaches of the Yangtze River	South China	Northeast China	Northwest China	Southwest China
Number of stations	24	20	33	28	33	68	49

Moreover, being the same as the analysis of intraseasonal variability of the summer monsoon over China described in the previous section, the observed data at 255 observational stations selected from the observed dataset of daily precipitation at 598 stations in China are used to study the interannual variations of summer rainfall in various climate regions of China, especially in the Yangtze River and Huaihe River valley, the middle and lower reaches of the Yangtze River, South China, and North China, whose locations are shown in Fig. 3.1. And the interannual variations of summer rainfall in these regions are discussed in this section.





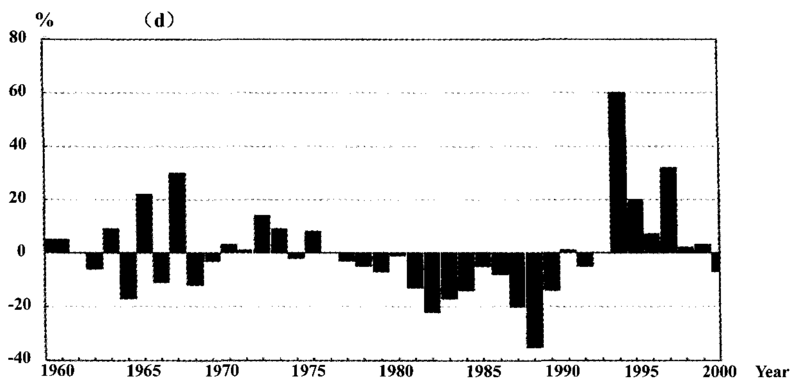


Figure 3.2. Interannual variations of the summer (June-August) rainfall anomalies (in percentage) averaged for the regions of (a) North China, (b) the Yangtze River and Huaihe River valley, (c) the middle and lower reaches of the Yangtze River, and (d) South China. The climatological mean summer (June-August) rainfalls in various regions averaged for 30 years from 1961 to 1990 are taken as the normals in the various regions.

Since the EASM mainly influences the interannual variability of summer rainfall in the southern and eastern parts of China, the observed data of daily precipitation at the selected observational stations shown in Table 3.1 in North China, the middle and lower reaches of the Yangtze River, the Yangtze River and Huaihe River valley and South China are used to analyze the interannual variations of summer (June-August) monsoon rainfall in these regions. Moreover, the climatological mean of precipitation in the various regions averaged for 30 summers from 1961 to 1990 is taken as the normal of precipitation in each region. In this section, the interannual variations of summer monsoon rainfall in these regions from 1978 are emphasized because the observed data of high cloud amount over East Asia and the tropical western Pacific are available from 1978. Figures 3.2 a-d are the interannual variations of summer rainfall anomalies (in percentage) in North China, the Yangtze River and Huaihe River valley, the middle and lower reaches of the Yangtze River and South China, respectively. Figures 3.2a-d feature large interannual variations of the summer monsoon rainfall in these regions of the eastern and southern parts of China. It may be clearly seen from Fig. 3.2b that in the summers of 1978, 1981, 1985, 1988, 1990, 1992, 1994, 1997 and 1999, the monsoon rainfall was 10% - 20% below normal in the Yangtze River and Huaihe River valley and droughts occurred in this region in the summers of 1978, 1985, 1988, 1992 and 1994, respectively. On the other hand, in the summers of 1980, 1982, 1983, 1987, 1989, 1991, 1996, 1998 and 2000, the summer rainfall was over 20% above normal, and severe floods occurred there in the summers of 1980, 1982, 1991 and 1996 respectively, and the rainfall was at least 30% greater than the mean in this region in these summers.

It may be also seen from Fig. 3.2a that the interannual variations of summer rainfall are also significant in North China. After 1978, the rainfall was more than 10% below normal in the summers of 1980, 1983, 1986, 1991, 1997, 1999 and 2000, respectively, and in particular, severe droughts occurred in the summers of 1997 and 1999, the rainfall was at least 30% less than the mean in this region in these two summers.

Comparing Fig. 3.2a with Fig. 3.2b, it is clearly seen that after 1978, flood summers frequently appeared in the Yangtze River and Huaihe River valley, but drought summers frequently appeared in North China. Thus, the interannual variations of summer rainfall in these two regions seem to be opposite.

Similarly, from 1978 to 1993, the summer monsoon rainfall was below normal, and drought summers frequently occurred in the middle and lower reaches of the Yangtze River and South China. However, flood summers continuously appeared in these two regions from 1994 to 1999, as shown in Fig. 3.2c and Fig. 3.2d, respectively.

From the above-mentioned results, it may be seen that there are obvious interannual oscillations of 2-6 years in the variations of the summer monsoon rainfall in North China, East China and South China. However, the interannual variability of summer monsoon rainfall is different in these regions although these regions all are located in the East Asian monsoon region.

3.2. *The Quasi-Biennial Oscillation of Summer Monsoon Rainfall in China and its Link with that in Its Surrounding Regions*

In order to reveal the regularity of the interannual variations of summer monsoon rainfall in China, the Empirical Orthogonal Function (EOF) analysis method is applied. Figures 3.3a and 3.3b are the spatial distribution and corresponding time-coefficient series of the first component of EOF analysis (EOF1) (it explains 16.28% of the variance) of summer rainfall in China. As shown in Fig. 3.3a, the spatial distribution of EOF1 of summer monsoon rainfall shows a meridional tripole pattern in China, and the large-scale strong negative signal is in the middle and lower reaches of the Yangtze River and in the Yangtze River and Huaihe River valley, and the large-scale positive signals are in South China and North China. This also shows that the monsoon rainfall anomaly in North China is opposite to that in the Yangtze River and Huaihe River valley in summer. The corresponding time-coefficient series shown in Fig. 3.3b obviously exhibits a prevailing quasi-biennial oscillation from the mid-1970s to the late 1990s, but the quasi-biennial oscillation was not obvious from the 1950s to the mid-1970s. By using entropy spectrum analysis, the dominant oscillation period of the EOF1 of summer rainfall in China is about 2.2 years, as shown in Fig. 3.4.

From the above-mentioned result, it is clearly seen that there is an obvious quasi-biennial oscillation of the summer monsoon rainfall in East China. This result is in agreement with that studied by Miao and Lau (1990), Yasunari (1991), Lau and Shen (1992).

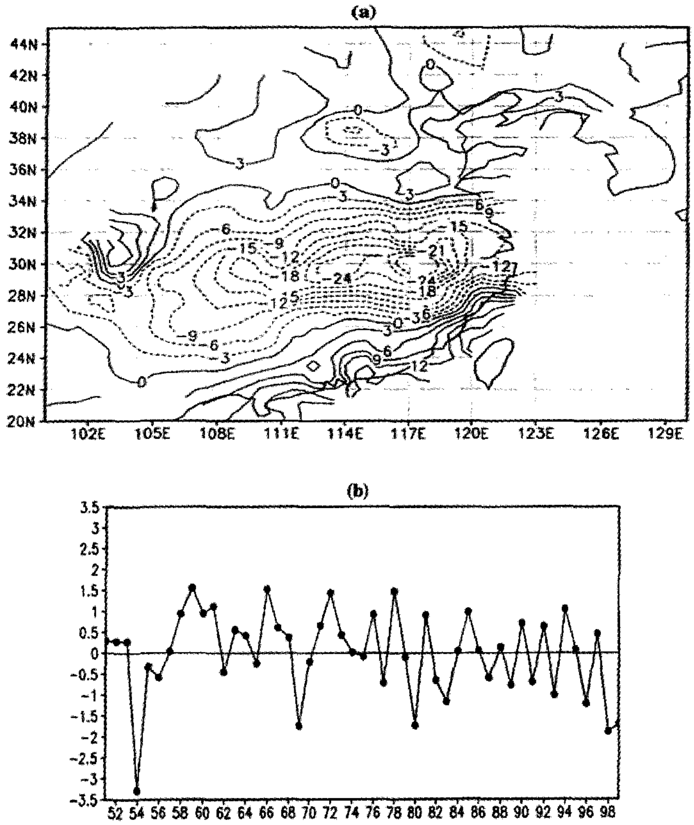


Figure 3.3. (a) The spatial distribution and (b) corresponding time-coefficient series of the first component of EOF analysis (EOF1) of summer rainfall in China from 1951 to 1999, analyzed using the dataset of monthly precipitation in 160 observational stations of China. The solid and dashed contours in Figure 3.3a indicate positive and negative signals, respectively, and the first EOF explains 16.28% of the variance.

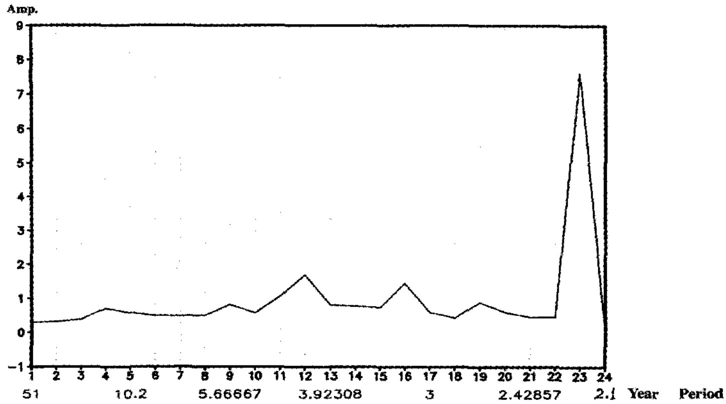


Figure 3.4. The entropy spectrum of the corresponding time-coefficients of EOF1 of summer rainfall in China

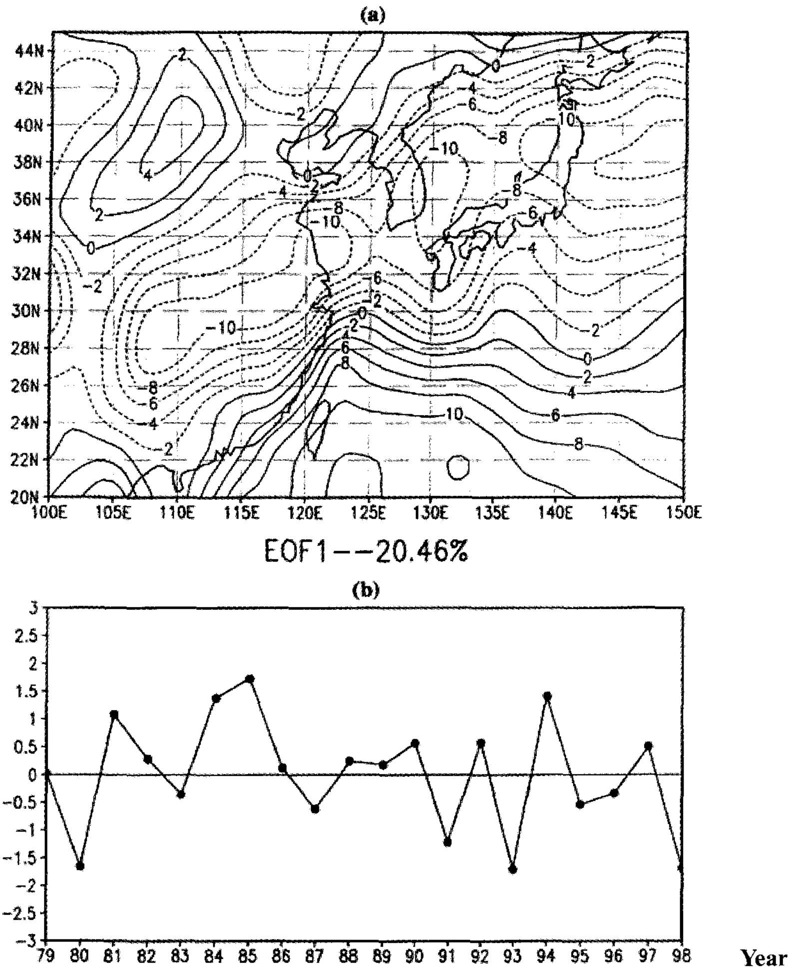


Figure 3.5. As in Fig. 3.3, except for using the dataset of precipitations analyzed by Xie and Arkin (1997) for 20 summers from 1979 to 1998. Note that the first EOF explains 20.46% of the variance.

The interannual variability of summer monsoon rainfall in Korea and Japan is also similar to that in the Yangtze River and Huaihe River valley of China. In order to investigate the linkage between the interannual variation of summer monsoon rainfall in China and that in its surrounding regions, the dataset of precipitation analyzed by Xie and Arkin (1997) for 20 summers from 1979 to 1998 is used to analyze the spatial distribution and the corresponding time-coefficient series of EOF analysis of summer monsoon rainfall in China and its surrounding regions. Figures 3.5a and 3.5b are the spatial distribution and the corresponding time-coefficient series of the EOF1 (it explains 20.46% of the variance) of summer rainfall in China and its surrounding regions, respectively. It may be found from Fig. 3.5a that the spatial distribution of the summer rainfall signal also exhibits a meridional tripole pattern. The large-scale negative signal is in the area from Japan, Korea to the middle and lower reaches of the Yangtze River, and the Yangtze River and Huaihe River valley of China. This can explain that there is a close linkage between the summer monsoon rainfall in the Yangtze River and Huaihe River valley of China and that in Japan and Korea. Moreover, the large-scale positive signal is in the area from the tropical western Pacific, the South China Sea and South China to the Indo-China Peninsula. Beside, another weaker positive rainfall signal is in North China. This pattern features that the large-scale rainfall signal in the Yangtze River and Huaihe River valley of China is the same as that in Korea and Japan and the opposite to that in the tropical western Pacific and Southeast Asia, and North China and Northeast China. Lu *et al.* (1995) also showed that the interannual variation of summer rainfall in the Yangtze River and Huaihe River valley is generally similar to those in Korea. Moreover, Fig. 3.5b also exhibits the characteristic of the quasi-biennial oscillation in the interannual variations of summer rainfall in East Asia and the tropical western Pacific.

3.3. *The Quasi-Biennial Oscillation of Water Vapor Transport by the EASM and its Impact on Summer Monsoon Rainfall over China and its Surrounding Regions*

As mentioned above, the interannual variability of the East Asian climate system has a great impact on the summer monsoon over China and its surrounding regions. Furthermore, the interannual variability of the summer monsoon over China greatly influences summer rainfall in China through water vapor transport. According to the result investigated by Huang *et al.* (1998), the characteristics of water vapor transport over the East Asian monsoon region including South China, East China, Korea and Japan are greatly different from those over the Indian monsoon region. Over the latter, the zonal component of water vapor transport is dominant. However, over the East Asian monsoon region, the meridional component of water vapor transport is large and has the same order as the zonal component. Moreover, the convergence of water vapor, which can form the monsoon rainfall, is mainly caused by the moisture advection over South China, East China, Korea and Japan. Therefore, the interannual variability of water vapor transport caused by monsoon flow is very important for the interannual variability of summer monsoon rainfall in China and its surrounding regions. Therefore, discussion of the impact of the water vapor transport by the EASM on summer rainfall in China and its surrounding regions is needed.

Assuming no water vapor above 100 hPa, the transport flux vector of water vapor $\vec{Q} = (\vec{Q}_\lambda, \vec{Q}_\varphi)$ can be described as follows:

$$Q_\lambda = \frac{1}{g}\int_{100}^{P_0} q \cdot udp \tag{3.1}$$

and

$$Q_\varphi = \frac{1}{g}\int_{100}^{P_0} q \cdot vdp \tag{3.2}$$

where Q_λ and Q_φ are the zonal and meridional components of water vapor transport flux, respectively, q is the specific humidity, u and v are the zonal and meridional components of wind field, respectively, and P_0 =1000 hPa for the sake of simplicity.

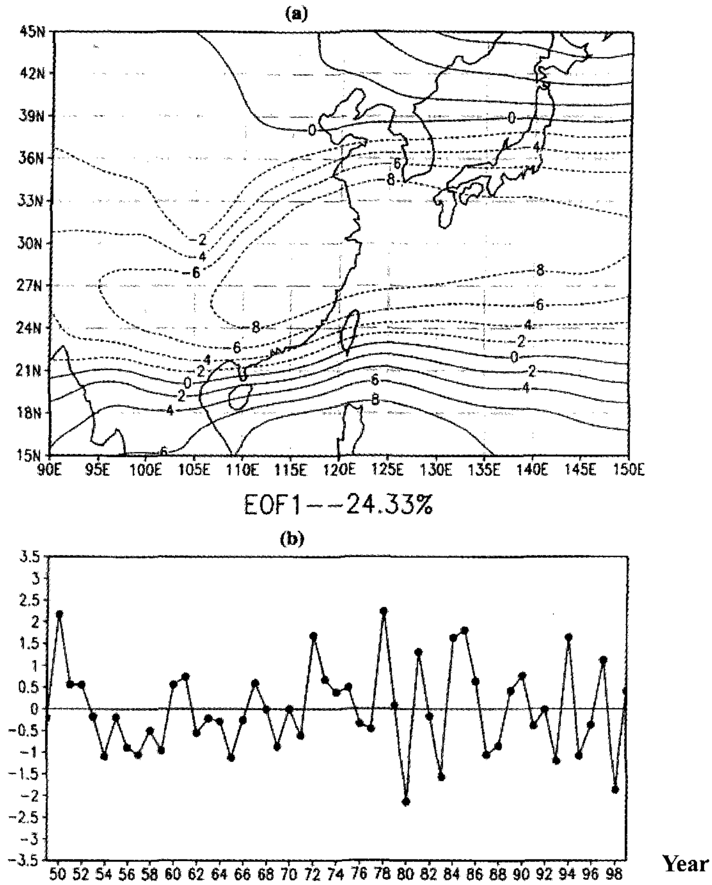


Figure 3.6. (a) The spatial distribution and (b) corresponding time-coefficient series of the first EOF component of zonal water vapor transports during summer, analyzed using the NCEP/NCAR reanalysis dataset of moisture and wind fields from 1951 to 1999. The solid and dashed contours in Figure 3.6a indicate positive and negative anomaly signals of zonal water transports, respectively, and the first EOF explains 24.33% of the variance.

The water vapor transport fluxes over East Asia and the tropical western Pacific in the summers (June-August) of 1951-1999 are calculated using Formulas (3.1) and (3.2) and the NCEP/NCAR reanalysis data of moisture. Moreover, the EOF analysis is applied to study the characteristics of spatial-temporal variability of the zonal and meridional components of water vapor transport over the regions. Figures 3.6a and 3.6b are the spatial distribution and the corresponding time-coefficient series of the first EOF component (EOF1) of zonal water vapor transport in summer, respectively. The spatial distribution of the EOF1 of zonal water vapor transport shown in Fig. 3.6a exhibits a meridional tripole pattern, which is the same as the spatial pattern of the EOF1 of summer rainfall in China and its surrounding regions shown in Fig. 3.3a and Fig. 3.5a. Moreover, it may be seen from the corresponding time-coefficient series of the first component of EOF shown in Fig. 3.6b that there is an obvious oscillation of two-three years from the mid-1970s. That is to say, there is a characteristic of quasi-biennial oscillation in the interannual variation of zonal water vapor transport over China and its surrounding regions from the mid-1970s. This is in agreement with the characteristic of quasi-biennial oscillation in the interannual variations of summer rainfall in China, Korea and Japan shown in Fig. 3.3b and Fig. 3.5b. Moreover, the corresponding time-coefficient series of the second EOF component of meridional water vapor transport also exhibits the characteristic of quasi-biennial oscillation (figure omitted).

The above-mentioned analysis shows that there is an obvious quasi-biennial oscillation in the interannual variations of water vapor transport by the summer monsoon flow over East Asia and the tropical western Pacific, which may cause the quasi-biennial oscillation of summer monsoon rainfall over China and its surrounding regions.

4. Interdecadal Variation of the Summer Monsoon over China

As in other monsoon regions, the decadal and interdecadal variations of the summer monsoon over China are also significant (e.g., Chang *et al.* 2000; Huang 2001). Thus, the discussion on decadal and interdecadal variations of the summer monsoon over China will be emphasized in this section. Moreover, being different from the interannual variability of summer surface air temperature, the interdecadal variability of summer surface air temperature is as significant as the summer monsoon rainfall in East Asia. Therefore, in order to explain the interdecadal variability of the summer monsoon over China, the observed dataset of monthly precipitation and surface air temperature at 160 observational stations of China, whose distribution is shown in Fig. 2.1, provided by NCC/CMA, are analyzed in this section.

4.1. *Decadal Variations of Surface Air Temperature and Precipitation in China*

In order to analyze the decadal variability of summer (June-August) surface air temperature and precipitation in China, the climatological mean monthly precipitation and surface air temperature averaged from 1961 to 1990 as shown in Figs. 2.2b-d and Figs. 2.4b-d are taken as the normals of summer (June-August) precipitation and surface air temperature, respectively. By this way, the decadal-mean anomaly distributions of summer surface air temperature and precipitation can be obtained by averaging for various decades from the 1950s to the 1990s.

Figures 4.1a-e indicate the decadal-mean anomaly distributions of summer (June-August) surface air temperature over China in the 1950s, 1960s, 1970s, 1980s and 1990s, respectively. From these figures, it can be found that a warming trend was obvious in North China, Northwest China, Northeast China and South China. From the 1980s to the 1990s, especially in the 1990s, the surface air temperature anomalies reached 0.4-0.8°C in North China and

Northwest China. However, there was a cooling trend in the middle and upper reaches of the Yangtze River from the 1980s to the 1990s, and the maximum anomaly of surface air temperature reached –0.6°C.

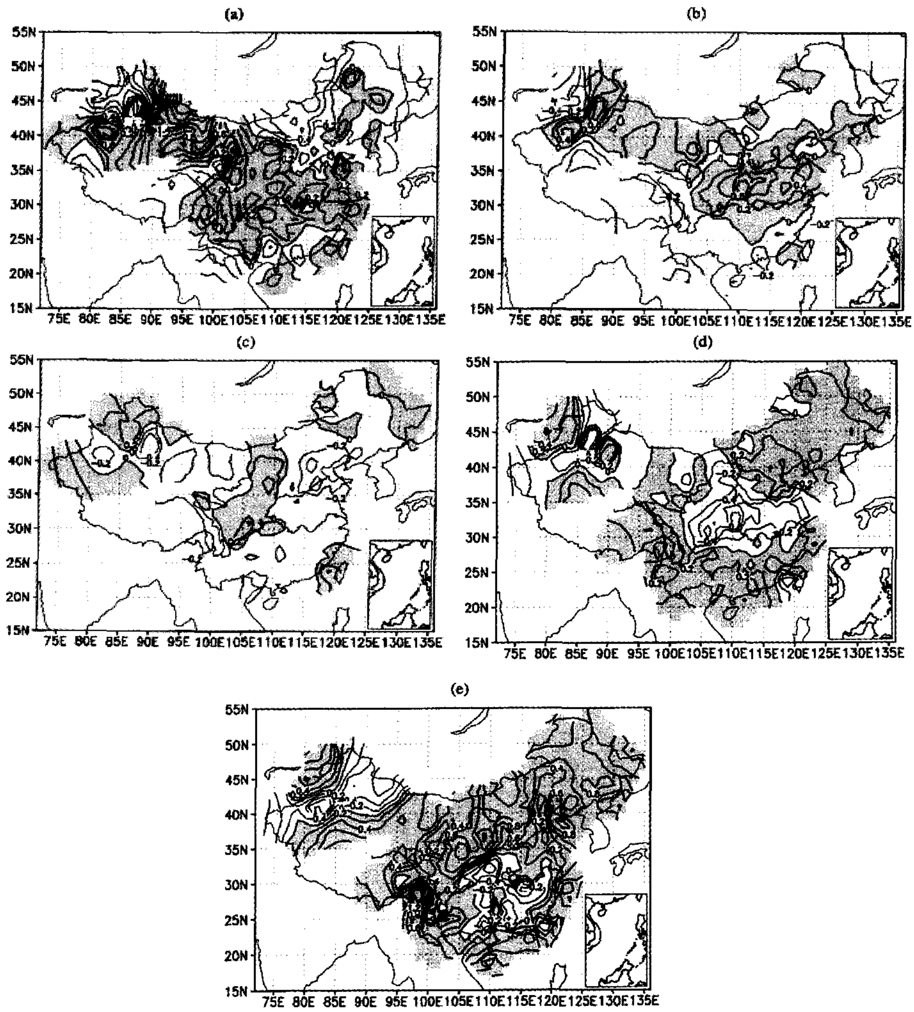


Figure 4.1. Decadal-mean anomaly distributions of summer (June-August) surface air temperature anomalies at 160 observational stations of China in (a) the 1950s, (b) the 1960s, (c) the 1970s, (d) the 1980s, and (e) the 1990s. Units: °C. The solid and dashed contours (contour interval: 0.2°C) indicate positive and negative temperature anomalies, respectively, and positive anomalies of surface air temperature are shaded. The distributions of climatological mean monthly surface air temperature in June, July and August shown in Figs. 2.4c-e are taken as the respective normals.

Compared with the decadal variations of summer surface air temperature shown in Figs. 4.1a-e, the interdecadal fluctuation of summer precipitation is more obvious in China. Figures 4.2a-e are the decadal mean anomaly distributions of summer (June-August) precipitation in the 1950s, 1960s, 1970s, 1980s and 1990s, respectively. It may be seen from these figures that the anomaly distributions of summer precipitation in the 1980s and 1990s were obviously different from those in the 1950s, 1960s, 1970s. The summer monsoon rainfall obviously increased in the 1980s and 1990s in the Yangtze River valley, especially in the middle and lower reaches of the Yangtze River where the summer precipitation anomalies were 30%-40% greater than the climatological mean. Thus, flood disasters frequently

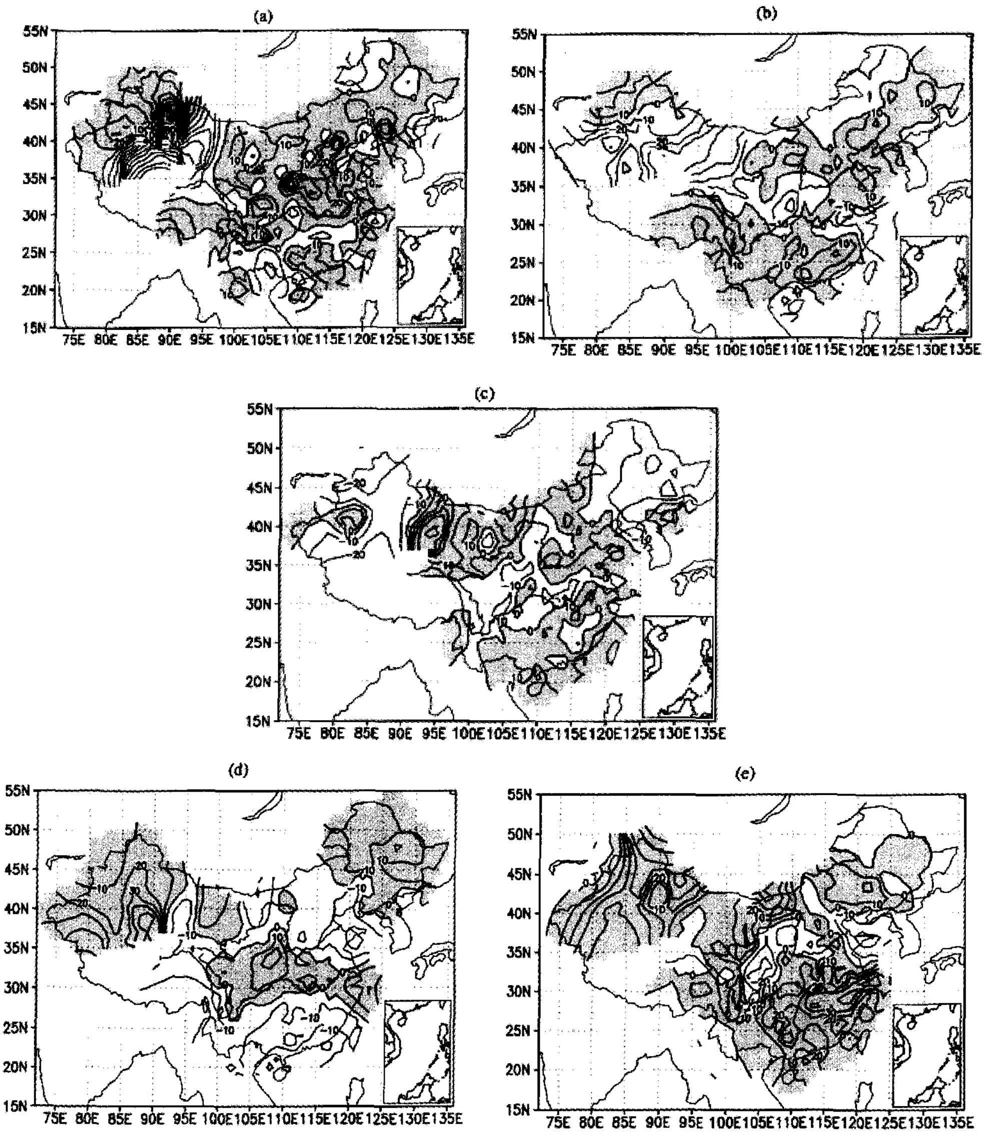


Figure 4.2. As in Fig. 4.1, except for the summer rainfall anomalies (in percentage). The solid and dashed contours (contour interval: 10%) indicate positive and negative rainfall anomalies, respectively, and positive rainfall anomalies are shaded. The distribution of climatological mean monthly rainfall in June, July and August shown in Figs. 2.2c-e are taken as the respective normal.

occurred in this region in the 1990s, as shown in Fig. 3.2c. Moreover, the monsoon rainfall obviously increased in South China in the 1990s although it decreased in the 1980s. However, the opposite phenomenon appeared in North China and the Yellow River valley. In the 1980s and 1990s, the summer precipitation was 10%-20% less than the climatological mean and prolonged droughts occurred in North China. Besides, Figures 4.2a-e also show that the summer rainfall obviously increased in Northwest China in the period of the 1980s and 1990s.

4.2. *The Climate Jump in the Late 1970s*

The interdecadal variations of summer monsoon rainfall in China may be also seen from the interannual variations of summer precipitation in various regions of China. Figures 3.2a-d show that the interannual variations of summer precipitation in North China and the middle and lower reaches of the Yangtze River before 1976 were different from those after 1976. Thus, the difference between the summer precipitation anomalies (in percentage) averaged for 1977-2000 and those averaged for 1967-1976 over China is analyzed and is shown in Fig. 4.3. The figure shows that there was a large difference between the summer precipitation anomalies averaged for 1977-2000 and those averaged for 1967-1976 in North China, Northwest China and the Yangtze River valley. After the late 1970s, the summer monsoon rainfall obviously decreased in North China and South China, but the summer precipitation obviously increased in the Yangtze River valley and Northwest China. Thus, a climate jump of summer rainfall in China occurred in the late 1970s. Due to this climate jump, North China has become drier, while the Yangtze River valley has become wetter, and summer precipitation has obviously increased in Northwest China.

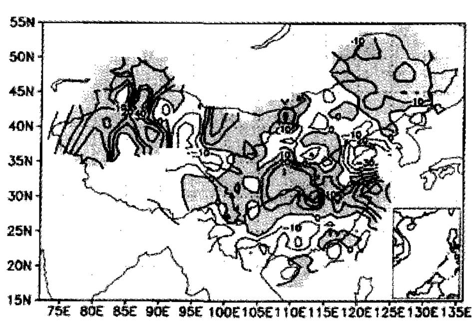


Figure 4.3. Differences between the summer precipitation anomalies (in percentage) averaged for 1977-2000 and those averaged for 1967-1976 at 160 observational stations in China. The solid and dashed contours (contour interval: 10%) indicate positive and negative values of rainfall differences, and positive rainfall differences are shaded. The distributions of climatological mean monthly precipitation in June, July and August shown in Figs. 2.2c-e are taken as the respective normal.

The climate jump can be also seen from the interdecadal variations of summer (June-August) surface air temperature. The difference between the summer surface air temperature anomalies averaged for 1977-2000 and those averaged for 1967-1976 in China is also analyzed. As shown in Fig. 4.4, there was also a large difference between the summer air temperature anomalies in China averaged for 1977-2000 and those averaged for 1967-1976. The figure shows the positive surface air temperature differences in North China, Northeast China, Northwest China and South China, respectively. Moreover, it may be seen that the negative surface air temperature difference appeared in the Yangtze River valley, with the maximum of about –0.6 in the upper reaches of the Yangtze River. This shows the decrease of surface air temperature in the upper reaches of the Yangtze River from the late 1970s, and it may be associated with the increase of summer rainfall there.

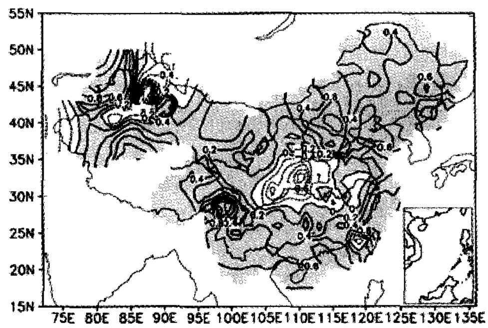


Figure 4.4. As in Fig. 4.3, except for summer surface air temperature anomalies in China. Units: °C. The solid and dashed contours (contour interval: 0.2°C) indicate positive and negative values of surface air temperature differences, respectively, and positive surface air temperature differences are shaded.

From Figs. 4.3 and 4.4, it may be found that the regional characteristics of the interdecadal variations of summer climate in China are significant. In the eastern and southern parts of China which are in the East Asian monsoon region, the increase (decrease) of summer monsoon rainfall is accompanied by a decrease (increase) of surface air temperature. For example, from the late 1970s to the 1990s, the summer monsoon rainfall obviously increased in the Yangtze River valley, which was accompanied by a decrease of surface air temperature in summer. Oppositely, the summer monsoon rainfall obviously decreased in North China, which was accompanied by an increase of surface air temperature in summer. However, in Northwest China which is not in the East Asian monsoon region, the increase of summer rainfall was accompanied by an increase of surface air temperature from the late 1970s to the 1990s.

Furthermore, the analysis of observed winter (December-February) surface air temperature also shows that the climate jump also occurred in the late 1970s. Since the late 1970s, the winter surface air temperature has obviously increased in Northeast China, North China and Northwest China. Because climate variations of the winter monsoon over China are described in another chapter, the detailed results will not be repeated in this section.

4.3. *The Interdecadal Variability of Water Vapor Transport by the EASM and its Impact on Summer Monsoon Rainfall over China*

Huang and Yan (1999) and Huang *et al.* (1999) investigated the cause of the interdecadal variations of summer climate in North China from the interdecadal variability of the southerly wind at 700 hPa over East Asia and pointed out that the southerly winds in the lower troposphere over North China began to become weak from 1965 and became more weak from the late 1970s. This led to the weakening of the meridional water vapor transport in North China and caused the beginning of drought in North China from 1965 and the prolonged droughts in this region from the late 1970s.

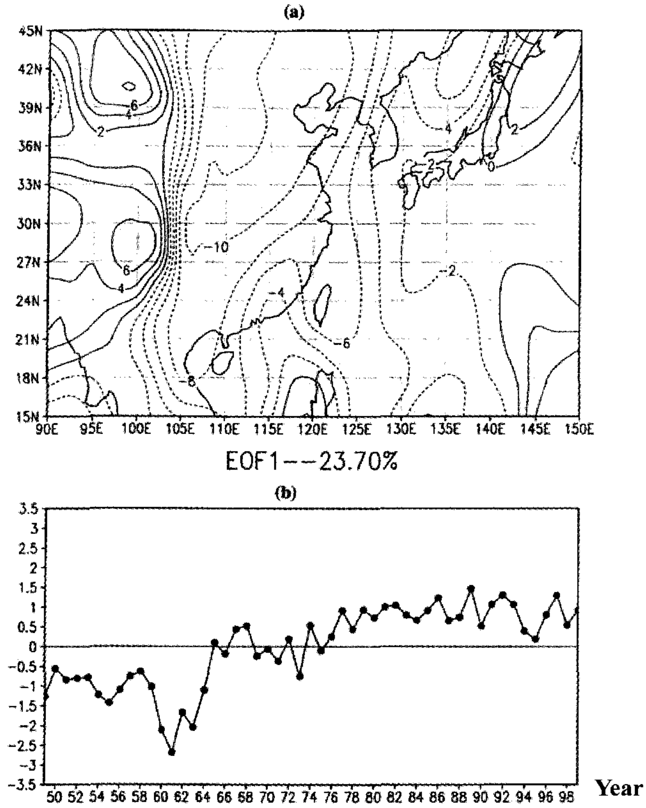


Figure 4.5. As in Fig. 3.6, except for meridional water vapor transport. The solid and dashed contours in Figure 4.5a indicate positive and negative anomaly signals of meridional water vapor transport, respectively, and the first EOF explains 23.70% of the variance.

The interdecadal variability of water vapor transport over East Asia is also analyzed using the water vapor transport fluxes over East Asia and the tropical western Pacific during summer (June-August) calculated in Section 3. As mentioned above, the meridional transport of water vapor is large over the eastern part of China, thus, the interdecadal variability of water vapor transport may be clearly shown in the meridional component of water vapor transport in China and its surrounding regions. Figures 4.5a and 4.5b are the spatial distribution and the corresponding time-coefficient series of the first EOF component of meridional water vapor transport in summer, respectively. It may be seen from Fig. 4.5a that the spatial distribution of meridional water vapor transport signals exhibits a negative single pattern over the eastern part of China, the Korean Peninsula and the western part of Japan, which is very different from the spatial distribution of the first EOF component of zonal water vapor transport shown in Fig. 3.6a. Moreover, it may be seen from the corresponding time-coefficient series of the first component of EOF shown in Fig. 4.5b that there is an obvious interdecadal variability in the interannual variations of meridional water vapor transport. As shown in this figure, in the summer of 1965, the time-coefficient of the EOF 1, which explains 23.7% of the variance, abruptly became a positive value from a negative value, and then oscillated about the normal up to 1976. Moreover, from 1977 to 1999, the time-coefficients of the EOF1 became larger positive values. Combined with the spatial pattern shown in Fig. 4.5a, it may be clearly seen that the meridional water vapor transport remarkably decreased over the eastern part of China, the Korean Peninsula and the western part of Japan from the late 1970s. This led to a serious decrease of water vapor transported into North China and North Korea. The serious decrease of water vapor transported into North China is closely associated with the weakening of southerly winds in the lower troposphere over North China (e.g., G. Huang 1999). This may be an important cause of the prolonged drought disaster which has occurred in North China and North Korea from the late 1970s.

5. The East Asian Climate System and its Impact on the Interannual Variations of the Summer Monsoon over China

As mentioned above, the interannual and interdecadal variabilities of the summer monsoon over China are significant, which have caused severe droughts and floods in China. Recently, the study of the East Asian climate system and its impact on the climate variation of the summer monsoon over China has been greatly promoted. Thus, the East Asian climate system influencing the interannual variations of the summer monsoon over China and its surrounding regions will be simply discussed in this section.

5.1. *The East Asian Climate System*

The summer monsoon climate variabilities in East Asia are mainly influenced by the interactions among the components of the East Asian climate system (e.g., Huang *et al.* 2003). As shown schematically in Fig. 5.1, the East Asian climate system includes the East Asian monsoon, the western Pacific subtropical high and the disturbances in middle latitudes in the atmosphere, the thermal effect of the West Pacific warm pool, the ENSO cycle in the tropical Pacific, the land surface processes in the Eurasian continent and the thermal and dynamical effects of the Tibetan Plateau. Due to the interactions among these components of the East Asian climate system, the climate variabilities of the summer monsoon with different timescales are caused in China and its surrounding regions. Therefore, the relationships between the interannual variations of the summer monsoon over China and these components of the East Asian climate system are analyzed using the observed data in the following subsections.

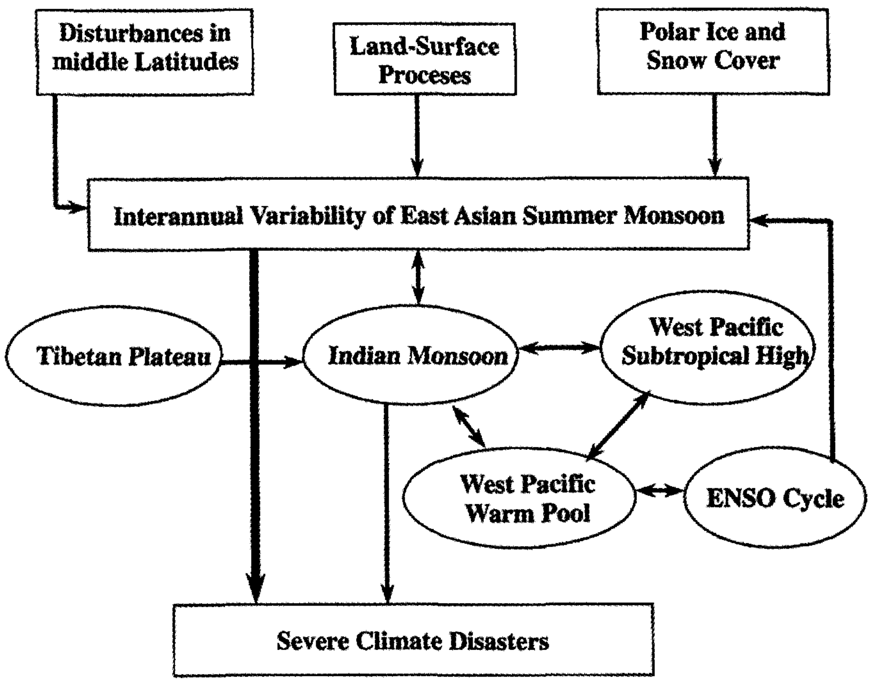


Figure 5.1. Conceptual diagram of the East Asian climate system.

5.2. *Dynamic and Thermal Effects of the Tibetan Plateau on the Interannual Variability of the Summer Monsoon over China*

The Tibetan Plateau has an important dynamical effect on the interannual variability of the EASM. By using the GFDL 9-layer GCM simulations, Hahn and Manabe (1975) showed that due to the dynamic effect of the Tibetan Plateau, the strong southwesterly (SW) flow can extend to the eastern part of China including South China, the Yangtze River valley, the Yangtze River and Huaihe River valley and North China from the Bay of Bengal. In the late 1970s, Ye and Gao (1979) first pointed out the thermal effect of the Tibetan Plateau on the Asian monsoon. Later on, many scholars also emphasized the thermal effect of the Tibetan Plateau on the Asian summer monsoon (e.g., Nitta 1983; Luo and Yanai 1984; Huang, 1984, 1985), and pointed out that the heating anomaly over the Tibetan Plateau has a large impact on the Asian summer monsoon circulation anomalies. Wu and Zhang (1997) explained that the Tibetan Plateau may play an air-pumping effect on the Asian summer monsoon onset through the sensible heating. Recently, Zhang *et al.* (2002) pointed out that the heating oscillation over the Tibetan Plateau can trigger the east–west oscillation of the South Asian high, which has a significant influence on the summer monsoon anomalies over the Yangtze River valley. As for the dynamic and thermal effects of the Tibetan Plateau on the interannual variability of the summer monsoon over China, these have been described in another chapter and will be omitted here.

5.3. *Thermal Effect of the Tropical Western Pacific on the Interannual Variability of the Summer Monsoon over China and the EAP Pattern Teleconnection*

The thermal state of the tropical western Pacific and the convective activities around the Philippines has a large impact on the interannual variability of the EASM. This impact may be through the western Pacific subtropical high and the onset of the South China Sea summer monsoon (SCSM). Thus, in this subsection and the next subsection, the impacts of the thermal state of the tropical western Pacific on the western Pacific subtropical high and the onset of the SCSM will be discussed.

5.3.1. Impact of the Thermal State of the Tropical Western Pacific on the Western Pacific Subtropical High

Since the summer monsoon rainband over the eastern part of China, Korea and Japan is located on the north side of the western Pacific subtropical high, the location of the western Pacific subtropical high greatly influences the distribution of the summer monsoon rainband over East Asia. Thus, the western Pacific subtropical high is an important component of the East Asian summer monsoon system (e.g., Tao and Chen 1987). If its location is shifted northward, a hot and drought summer may appear in the Yangtze River and Huaihe River valley. On the other hand, if its location is shifted southward, a flood summer may occur there.

The studies by many scholars (e.g., Nitta 1987; Huang and Li 1987; Kurihara 1989; Tsuyuki and Kurihara 1989; Huang and Sun 1992) showed that the thermal states of the tropical western Pacific and the convective activities around the Philippines play an important role in the interannual variations of the EASM. Moreover, Nitta (1987), Huang and Li (1987) and Kurihara (1989) all pointed out from the analyses of observed data that the thermal states of the tropical western Pacific and the convective activities around the Philippines can greatly influence the interannual variability of the western Pacific subtropical high.

The interannual variations of the western Pacific subtropical high are closely related to the thermal state of the tropical western Pacific or the convective activities around the Philippines. Figures 5.2a and 5.2b are the composite distributions of monthly mean 500 hPa height fields in July over East Asia for the summers with strong convective activities and for the summers with weak convective activities around the Philippines, respectively. From Fig. 5.2a, it can be seen that when the tropical western Pacific is in a warming state, i.e., the summer (June-August) SST anomaly in the area of NINO-west (EQ - 14°N, 130°E - 150°E) is above normal, the convective activities are strong around the Philippines, then the 500 hPa height field in July over East Asia will display a similar distribution to that shown in Fig. 5.2a.

In this case, the western Pacific subtropical high will shift unusually northward and will be located over the Yangtze River and Huaihe River valley. This will cause a hot and drought summer in the Yangtze River and Huaihe River valley. On the other hand, when the tropical western Pacific is in a cooling state, i.e., the summer SST anomaly in NINO-west is below normal, the convective activities are weak around the Philippines. In this case, the 500 hPa height field in July over East Asia will display a distribution as shown in Fig. 5.2b, which is very different from Fig. 5.2a. Under this situation, the western Pacific subtropical high will shift unusually southward and will be located to the south of the Yangtze River. This is favorable for the long-term maintenance of the Meiyu front in the Yangtze River and Huaihe River valley, thus, a flood summer will occur there.

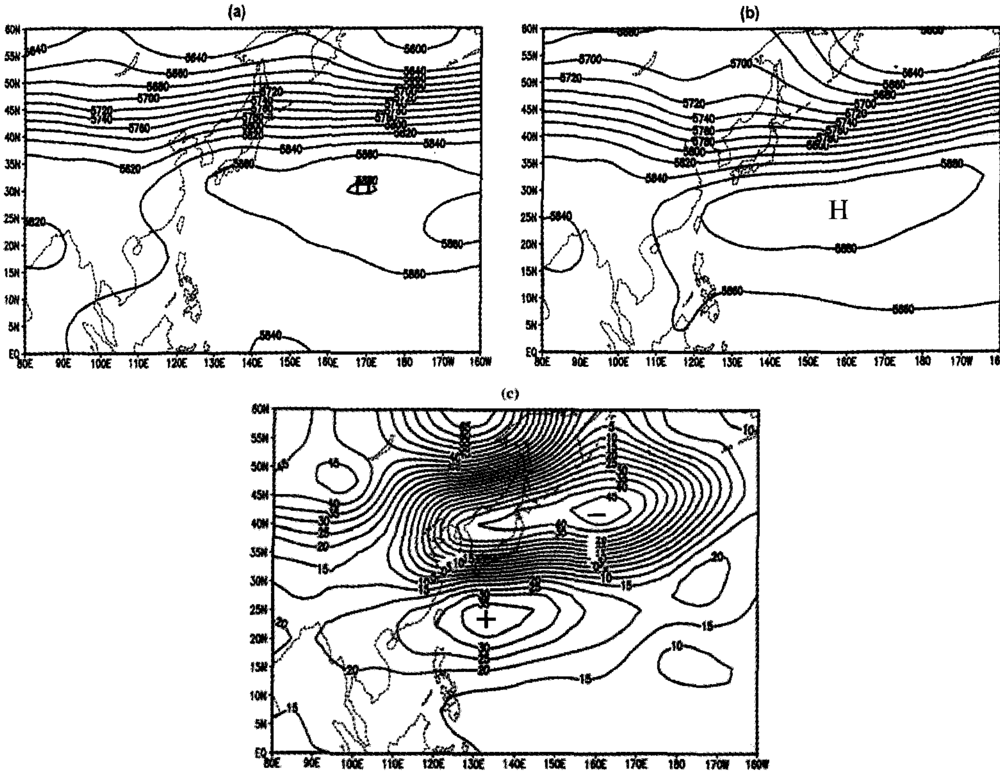


Figure 5.2. The composite distributions of monthly mean 500 hPa height field in July over East Asia for (a) summers with strong convective activities and (b) summers with weak convective activities around the Philippines (i.e., 10°N - 20°N, 110°E - 140°E) and (c) the difference between the distribution shown in Fig. 5.2b and that shown in Fig. 5.2a. Units: gpm. The NCEP/NCAR reanalysis data are used in this analysis. The contour interval is 20 gpm in Figs. 5.2a and 5.2b, and the solid and dashed contours (contour interval; 5 gpm) indicate positive and negative values of the 500 hPa height differences in Fig. 5.2c, respectively.

In order to show clearly the impact of the thermal state of the tropical western Pacific or the convective activities around the Philippines on the western Pacific subtropical high, the difference between the composite 500 hPa height field in July for the summers with weak convective activities around the Philippines and that for the summers with strong convective activities around the Philippines is shown in Fig. 5.2c. As shown in the figure, the positive differences are located in South China and eastern Siberia, respectively, and the negative differences appear in the Yangtze River and Huaihe River valley and North China. This features similar characteristics to the EAP pattern circulation anomalies over East Asia. It may explain that the location of the western Pacific subtropical high in a summer with weak convective activities around the Philippines is obviously different from that in a summer with strong convective activities around the Philippines.

In the above-mentioned analysis, the thermal state of the tropical western Pacific greatly influences the anomalous location of the western Pacific subtropical high. Recently, Lu (2001) and Lu and Dong (2001) also showed that the convective activities over the tropical western Pacific have a significant impact on the zonal shifts of the western Pacific subtropical high. They pointed out that if the convective activities are strong over the tropical western Pacific, the western Pacific subtropical high shifts eastward, but it will extend westward if convective activities are weaker over the tropical western Pacific.

5.3.2. Impact of the Thermal State of the Tropical Western Pacific on Convective Activities over the Tropical Western Pacific and East Asia

Nitta (1986) systematically studied the long-term variability of convective activities over the tropical western Pacific and showed that the interannual variations of high cloud amount in this region are closely associated with the thermal state of the tropical western Pacific. Nitta (1987) and Huang and Li (1987, 1988) pointed out that the thermal state of the tropical western Pacific can also influence the convective activities over East Asia through teleconnection.

In order to show the impact of the thermal state of the tropical western Pacific on the convective activities over the tropical western Pacific and East Asia, GMS-observed $1^\circ \times 1^\circ$ data of monthly mean black body temperature (TBB) from 1980 to 1998, provided by JMA, are used in this study. Figures 5.3a and 5.3b are the composite distribution of TBB anomalies in East Asia and the tropical western Pacific averaged for the summers (June-August) with the warming state and for the cooling state of the tropical western Pacific, respectively. Since TBB indicates the temperature of a black body at the surface in a cloud free area, the value of TBB is higher in the area. Thus, a high value of TBB or positive anomaly of TBB can denote weak convective activities. Oppositely, in a cloudy area, TBB indicates the temperature of a black body at the cloud top, and the value of TBB may be lower. Thus, a low value of TBB or negative anomaly of TBB can indicate strong convective activities in the area. By this way, it may be shown from Fig. 5.3a that in a summer with a warming state of the tropical western Pacific, i.e., the summer SST anomaly in the area of NINO-west (EQ - 14°N, 130°E - 150°E) is above normal, convective activities are weak from the Yangtze River and Huaihe River valley to South Korea and Japan, and strong from the Indo-China Peninsula and the South China Sea to the tropical western Pacific around the Philippines. In contrast, Fig. 5.3b shows that in a summer with a cooling state of the tropical western Pacific, i.e., the summer SST anomaly in the area of NINO-west is below normal, distributions of strong and weak areas of convective activities are just opposite to those shown in Fig. 5.3a. Moreover, Fig. 5.3b shows that convective activities are strong over the equatorial central Pacific in this case. Beside, from Figs. 5.3a and 5.3b, it may be seen that the distributions of convective activities over the tropical Pacific and East Asia also exhibit a meridional tripole pattern, which corresponds to that shown in Figs. 3.5a.

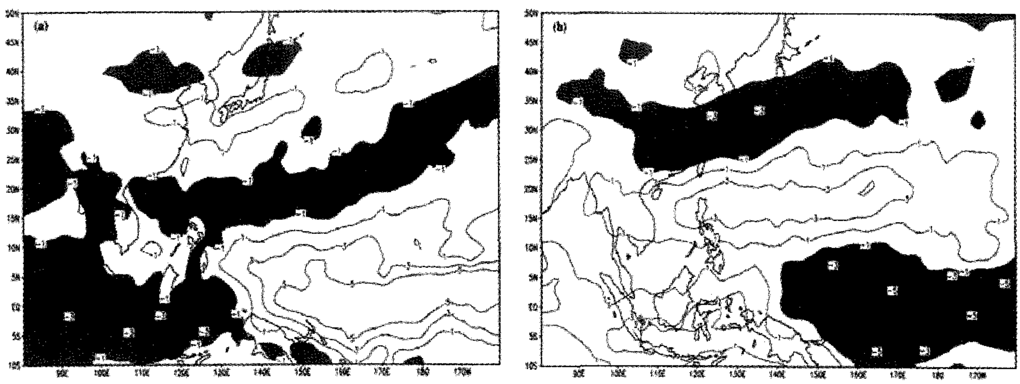


Figure 5.3. The composite distributions of TBB (black body temperature) anomalies in East Asia and the tropical western Pacific for (a) summers with the warming state and (b) summers with the cooling state of the tropical western Pacific. Units: K. The solid and dashed contours denote positive and negative TBB anomalies, respectively, and the TBB anomalies lower than -1K (e.g., strong convective activities) are shaded. The climatological mean summer (June-August) TBB is obtained by averaging the mean values of daily TBB from 1 June to 31 August. Data are obtained from JMA.

5.3.3. The East Asia/Pacific Pattern Teleconnection

The analyses of observed data showed that there is a teleconnection pattern of the summer circulation pattern over the Northern Hemisphere, i.e., the so-called Pacific-Japan (PJ) Oscillation (e.g., Nitta 1987) or the so-called East Asia/Pacific pattern teleconnection (e.g., Huang and Li 1987, 1988) and Lau (1992) also suggested a similar teleconnection pattern. The above-mentioned analyses of the detailed data can demonstrate further the existence of the EAP teleconnection pattern in the summer circulation anomalies over the Northern Hemisphere shown in Fig. 5.4, suggested by Nitta (1987) and Huang and Li (1987, 1988) from the observed OLR and circulation anomalies, respectively. It is seen from this teleconnection pattern that the quasi-stationary planetary wavetrain can propagate from Southeast Asia to the western coast of North America through East Asia during the Northern Hemisphere summer. What cause the formation of the EAP teleconnection pattern of the summer circulation anomalies over the Northern Hemisphere? This may be explained from the theory of planetary wave propagation (e.g., Huang and Li 1987, 1988; Huang and Sun 1992). As shown in Section 2, since the tropical western Pacific is a region of the highest SST in the global sea and the strong ascending branch of the Walker circulation is also over this region, the strong convergence of air and moisture leads to strong convective activities over this region. Thus, there is a strong heat source caused by strong convective activities. Huang and Li (1987, 1988) showed that due to the forcing by the heat source caused by strong convections, the quasi-stationary planetary waves responding to this forcing can propagate over the Northern Hemisphere. They calculated the propagating ray of planetary waves forced by an idealized heat source around the Philippines in a realistic summer basic flow with the formula of propagating ray path of planetary waves on the sphere, proposed by Hoskins and Karoly (1981). It may be seen from Fig. 5.5 that the planetary waves forced by a

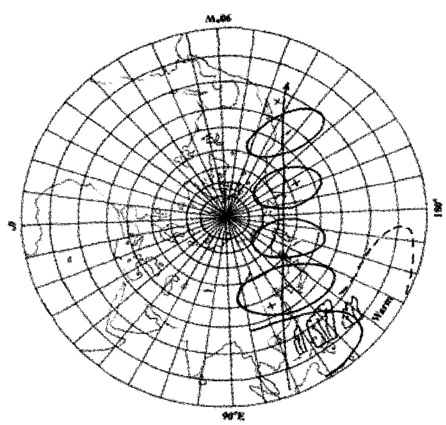


Figure 5.4. Schematic diagram of the East Asia/ Pacific (EAP) teleconnection pattern of summer circulation anomalies over the Northern Hemisphere (from Nitta 1987; Huang and Li 1987). + and − in the figure indicate positive and negative height anomalies at 500 hPa, respectively.

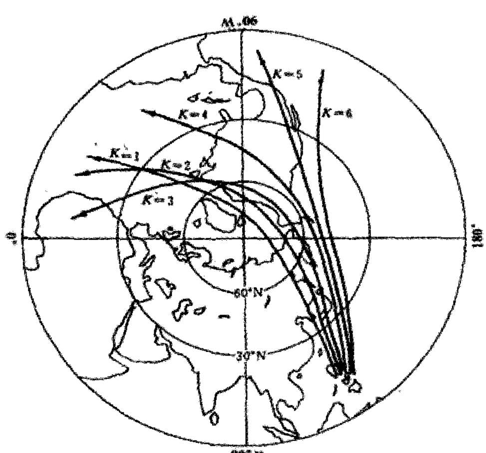


Figure 5.5. Propagating ray path of planetary waves forced by a heat source around the Philippines during boreal summer (from Huang and Li 1987, 1988).

heat source around the Philippines can propagate from the area around the Philippines toward the western coast of North America through East Asia and the North Pacific, which is in agreement with the teleconnection pattern shown in Fig. 5.4, obtained from the observed data (e.g., Nitta 1987; Huang and Li 1987; 1988; Lau 1992). Moreover, this teleconnection pattern has been demonstrated by Huang and Lu (1989), Nikaido (1989) and Huang and Sun (1992) using numerical simulations with general circulation models.

The above-mentioned analyses show that the impacts of the thermal state of the tropical western Pacific and convective activities around the Philippines on the interannual variations of the location of the western Pacific subtropical high, monsoon circulation and convective activities over East Asia during summer may be through the EAP pattern teleconnection, i.e., through the propagation of the quasi-stationary planetary wavetrain forced by the heat source around the Philippines.

5.4. *Impact of the Thermal State of the Tropical Western Pacific on the Onset of the South China Sea Monsoon (SCSM)*

The onset of the EASM is earlier in the area over the South China Sea, which generally is in May (e.g., Tao and Chen 1987), and this summer monsoon is called the South China Sea summer monsoon (hereafter SCSM). The onset of the SCSM is not only an important phenomenon of the intraseasonal variations of the EASM, but also has a large influence on the interannual variability of the Meiyu in the Yantze River and Huaihe River valley, the Baiu in Japan and the Changma in Korea. Figure 5.6 is the correlation between the summer (June-August) precipitation in China and the onset date of the SCSM defined by Lian and Wu

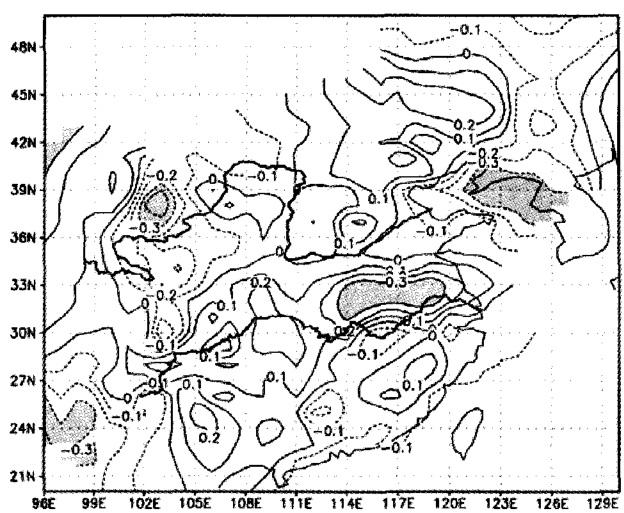


Figure 5.6. Correlation between the summer (June-August) rainfall in China and the onset date of the South China Sea monsoon (SCSM) during the period of 1951-1999. The shaded areas denote the correlations over the 95% confidence level, and the dashed and solid contours indicate negative and positive correlations, respectively. The onset date of the SCSM is provided by Lian and Wu (2002).

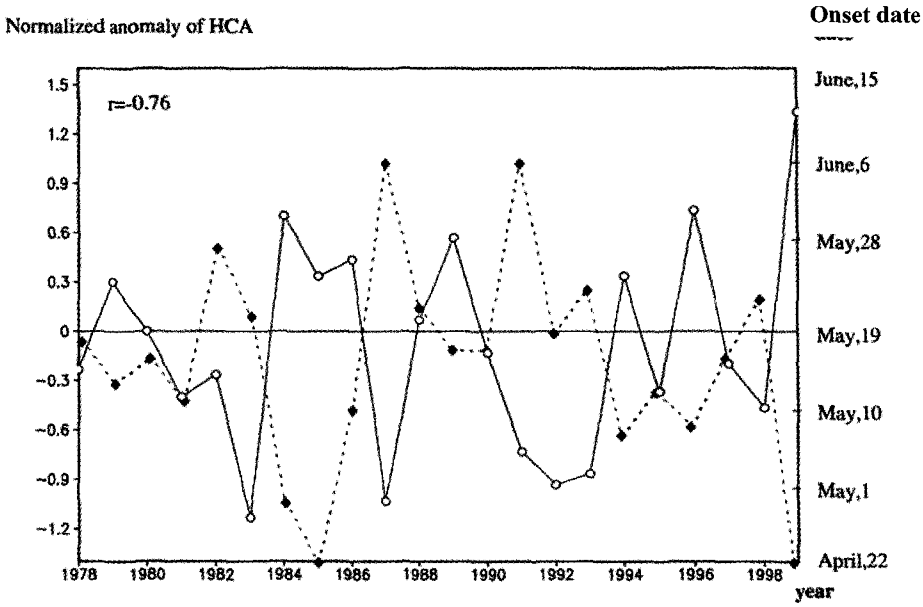


Figure 5.7. The interannual variations of onset date of the South China Sea summer monsoon (dashed line) (from Lian and Wu 2002) and the normalized high cloud amount anomaly (HCA) around the Philippines (i.e., 10°N - 20°N, 110°E - 140°E) in spring (March-May) (solid line) during the period of 1951-1999. Data of high cloud amount are obtained from Monthly Report on Climate System, JMA.

(2002) for 49 summers from 1951 to 1999. It may be clearly seen from the figure that there is an obvious positive correlation between the summer rainfall in the Yangtze River and Huaihe River valley and the onset date of the SCSM, the correlation coefficients between them are larger than 0.3 and exceed the 95% significance level. This is to say, if the onset of the SCSM is early, then the summer rainfall will be below normal in the Yangtze River and Huaihe River valley, but if the onset of the SCSM is late, such as in 1998, the summer rainfall will be above normal in this region. This has been used in prediction of summer monsoon rainfall anomalies in the Yangtze River and the Huaihe River valley.

The onset of the SCSM is closely associated with the thermal state of the tropical western Pacific in the leading winter and spring, especially with the convective activities around the Philippines in spring. As shown in Fig. 5.7, the correlation coefficient between the onset date of the SCSM and the normalized high cloud amount anomaly around the Philippines (i.e., 10°N - 20°N, 110°E - 140°E) in spring (March-May) can reach –0.76, which greatly exceeds the 99% significant level. Therefore, as shown in Fig. 5.7, the onset of the SCSM is early in a spring with a warming state of the tropical western Pacific, while it is late in a spring with a cooling state of the tropical western Pacific.

The impact of the convective activities around the Philippines on the onset of the SCSM may be interpreted from the Walker circulation. Cornejo-Garrido and Stone (1977) and Hartmann *et al.* (1984) showed that the heating caused by strong convective activities over the tropical western Pacific supplies the energy to drive the strong Walker circulation over the

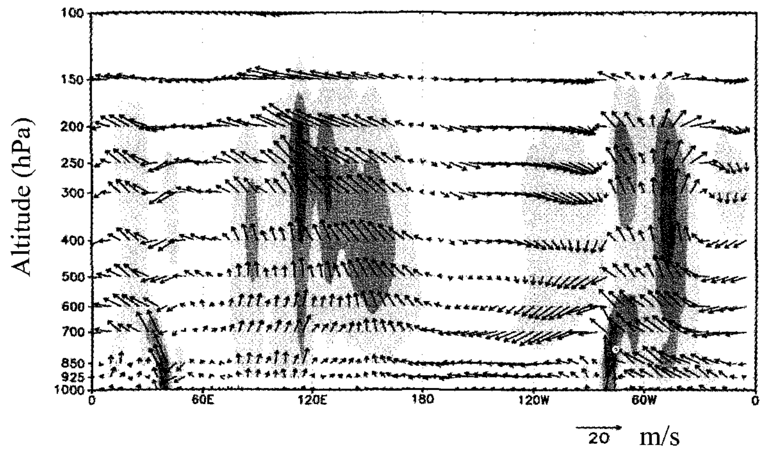


Figure 5.8. The composite distribution of zonal-altitude circulation along the equator (averaged between 5°S - 5°N) for the periods of the preceding November-April during the warming state of the tropical western Pacific (e.g., the positive SST anomaly in the area of NINO-west, i.e., EQ - 14°N, 130°E - 150°E). The vertical velocities in the figure are multiplied by 100. The NCEP/NCAR reanalyzed data are used in this analysis.

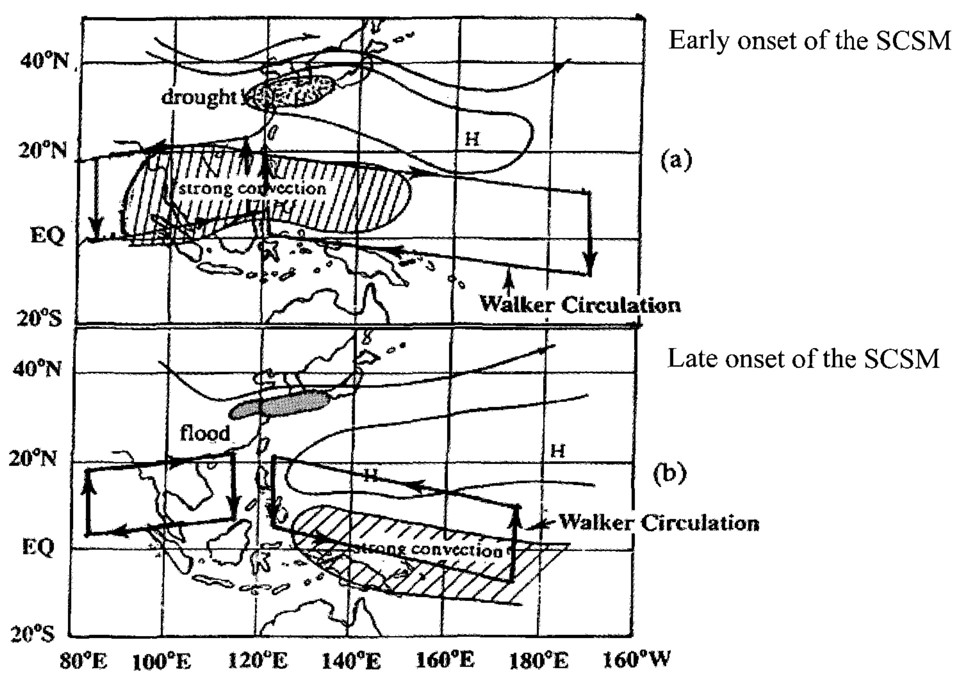


Figure 5.9. Schematic map of the relationships among the SST in the tropical western Pacific (TWP), the convective activities around the Philippines, the western Pacific subtropical high, the onset of the South China Sea summer monsoon (SCSM) and the summer rainfall in China and its surrounding regions. (a) in the warming state of the TWP; (b) in the cooling state of the TWP.

tropical western Pacific. When the tropical western Pacific is in a warming state, convective activities are strong around the Philippines in spring, as shown in Fig. 5.7. Thus, in this case, the ascending branch of the Walker circulation is strong over the tropical western Pacific. Figure 5.8 is the composite distribution of zonal-altitude circulation over the equatorial region (averaged between 5°S - 5°N) for preceding winters and springs (i.e., from the preceding November to April) during the warming state of the tropical western Pacific. It features a strong ascending branch of the Walker circulation over the equatorial western Pacific and the equatorial eastern Indian Ocean. Moreover, in this case, the western Pacific subtropical high shifts eastward and northward, as shown in Fig. 5.2a. Thus, the southwest flow can enter early into the area over the South China Sea from the Indo-China Peninsula because of the strong westerly flow in the lower troposphere over the tropical eastern Indian Ocean, as shown in Fig. 5.8. As a consequence, the onset of the SCSM may be early in a warming case of the tropical western Pacific. Oppositely, when the tropical western Pacific is in a cooling state, the onset of the SCSM may be late.

From the above-mentioned analyses, the impacts of the thermal states of the tropical western Pacific on the western Pacific subtropical high, convective activities over the tropical western Pacific and East Asia, summer monsoon rainfall in East Asia and onset date of the SCSM may be schematically summarized in Fig. 5.9. As shown in Fig. 5.9a, when tropical western Pacific is in a warming state, convective activities are intensified from the Indo-China Peninsula to the area east of the Philippines, and the western Pacific subtropical high may shift unusually northward. In this case, the onset of the SCSM may be early, and the summer monsoon rainfall may be below normal and drought may occur in the Yangtze River and Huaihe River valley of China, South Korea and Japan. On the other hand, as shown in Fig. 5.9b, when the tropical western Pacific is in a cooling state, convective activities are weak around the Philippines and are intensified over the equatorial central Pacific near the dateline, and the western Pacific subtropical high may shift southward. In this case, the onset of the SCSM may be late, and the summer monsoon rainfall may be above normal and flood may occur in the Yangtze River and Huaihe River valley of China, South Korea and Japan.

5.5. *Impact of ENSO Cycle on the Interannual Variability of the Summer Monsoon over China*

It is well known that ENSO cycle is one of the most striking phenomena in the tropics and has a great influence on the Asian monsoon. The weak Indian summer monsoon tends to occur in El Niño years (e.g., Shukla and Paolina 1983; Webster *et al.* 1998). Huang and Wu's (1989) study first showed that the summer monsoon rainfall anomalies in East Asia may depend on the stages of ENSO cycle. Recently, Huang and Zhou (2002) analyzed the composite distributions of the summer monsoon rainfall anomalies during different stages of 14 El Niño events that occurred in the period of 1951-2000. During the 50 summers of 1951-2000, the summers of 1951, 1957, 1963, 1965, 1969, 1972, 1976, 1982, 1987, 1991, 1993 and 1997 were just in the developing stage of the El Niño event, respectively, while the summers of 1953, 1958, 1977, 1983, 1992 and 1998 were just in the decaying state of the El

Niño event, respectively. Figures 5.10a and 5.10b are the composite distributions of the summer monsoon rainfall anomalies for 12 summers in the developing stage of El Niño events and for 6 summers in the decaying stage of El Niño events, respectively. The composite distributions of monsoon rainfall anomalies for the summers in the different stages of ENSO cycles show that during a summer in the developing stage of an El Niño event, the summer monsoon rainfall is strong and flood tends to occur in the Yangtze River and Huaihe River valley of China, but the summer monsoon rainfall is weak and drought may be caused in North China, as shown in Fig. 5.10a. In contrast, during a summer in the decaying stage of an El Niño event, as shown in Fig. 5.10b, the summer monsoon rainfall is weak and drought tends to occur in the Yangtze River and Huaihe River valley of China, but the summer monsoon rainfall may be normal or above normal in North China, and large positive monsoon rainfall anomalies may appear to the south of the Yangtze River and severe flood tends to occur there. For example, in the summer of 1998 when the 1997/98 El Niño event was in its developing stage, a particularly severe flood disaster occurred to the south of the Yangtze River, as shown in Fig. 2.11.

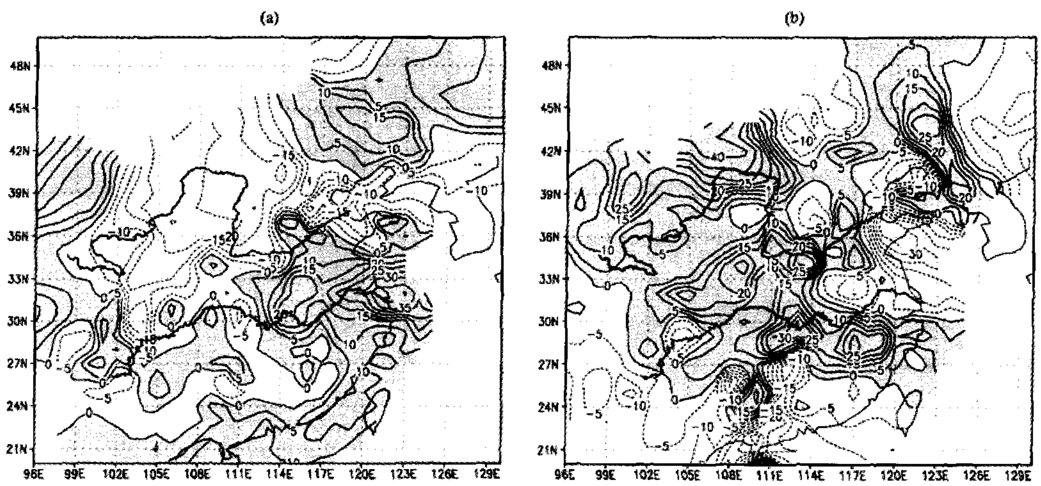


Figure 5.10. The composite distributions of summer rainfall anomalies (in percentage) in China (a) for 12 summers (June-August) when ENSO events were in their developing stage and (b) for 6 summers when ENSO events were in their decaying stage. The solid and dashed contours (contour interval: 5%) in these figures indicate positive and negative anomalies of rainfall, respectively, and the shaded areas denote positive rainfall anomaly regions. The climatological mean monthly precipitation in June, July and August shown in Figs. 2.2c-e are taken as the respective normals.

What causes severe flood disasters to occur in a summer in the decaying stage of an El Niño event in the Yangtze River valley? This may be associated with the water vapor transport from the tropical western Pacific and the Bay of Bengal. Zhang *et al.* (1996) and Zhang (2001) pointed out that since the western Pacific subtropical high shifts southward during and after the mature phase of an El Niño event, the southerly wind anomalies can appear in the lower troposphere along the southeast coast of China. The intensified southerly winds will be favorable for the transport of water vapor from the Bay of Bengal and the tropical western Pacific to the eastern part of China, which can provide the condition of sufficient water vapor for monsoon rainfall in this region.

However, it should be pointed out that there is an interaction between the Asian monsoon and ENSO cycle. Diagnostic and modeling studies have revealed that the variability of the Asian monsoon has a significant effect on the atmosphere/ocean coupled system in the equatorial Pacific (e.g., Yamagata and Matsumoto 1989; Li 1990; Yasunari 1990; Yasunari and Seki 1992; Webster and Yang 1992; Li *et al.* 2001). Because of the interaction between the Asian monsoon and ENSO cycle, the influence of ENSO cycle on the interannual variation of the summer monsoon over China is very complex and needs to be studied further.

5.6. *Impact of the Snow Cover in the Tibetan Plateau on the Interannual Variability of the Summer Monsoon over China*

The interannual variability of the Asian summer monsoon is also influenced by the Eurasian snow cover, especially the snow cover in the Tibetan Plateau. Chen and Yan (1981) and Wei and Luo (1996) pointed out that there is a positive correlation between the snow cover in the Tibetan Plateau and the summer monsoon rainfall in the upper and middle reaches of the Yangtze River.

Recently, the interannual variations of the days and depth of snow cover in the Tibetan Plateau were analyzed using the observed dataset of daily snow cover from the preceding October to May at 72 observational stations located in the Tibetan Plateau during 1960-1999 (Figs. 5.11a and 5.11b). Considering the quality of the observed data, the climatological mean days and depth of snow cover during the period from the preceding October to May averaged for the period of 1965-1999 at the various observational stations are taken as the respective normals. Moreover, both the days and depth anomalies of snow cover are normalized because the variance of days and depth of snow cover are very different at each station. As shown in Figs. 5.11a and 5.11b, both the days and depth of snow cover in the Tibetan Plateau have obvious interannual variations, and their interannual variations are similar. Thus, the normalized depths of snow cover over 2 and below –2 are considered as the criteria measuring a strong snow year and a weak snow year in this study, respectively. By this way, it can be seen from Fig. 5.11a that 1962, 1968, 1978, 1982, 1983, 1989, 1990, 1995 and 1998 were strong snow years, while 1960, 1965, 1967, 1971, 1976, 1984, 1995, 1991 and 1999 were weak snow years in the Tibetan Plateau.

The interannual variations of snow cover in the Tibetan Plateau have an important impact on the summer monsoon rainfall in the Yangtze River valley. Figure 5.12 is the correlation between the summer (June-August) rainfall in China and the normalized depth of snow cover in the period from the preceding October to May averaged for 72 observational stations in the Tibetan Plateau. It may be seen in the figure that there are larger positive correlations in the middle and upper reaches of the Yangtze River and negative correlations in South China and Northeast China. This may explain that if the snowfall in the Tibetan Plateau is heavy in the preceding winter and spring of a year, the following summer rainfall may be strong in the middle and upper reaches of the Yangtze River. For example, in the winter of 1997 and the spring of 1998, a particularly heavy snowfall appeared in the Tibetan Plateau as shown in Figs. 5.11a and 5.11b, and a heavy monsoon rainfall occurred in the middle and upper reaches of the Yangtze River and a particularly severe flood was caused in the Yangtze River valley in the summer of 1998, as shown in Fig. 2.11.

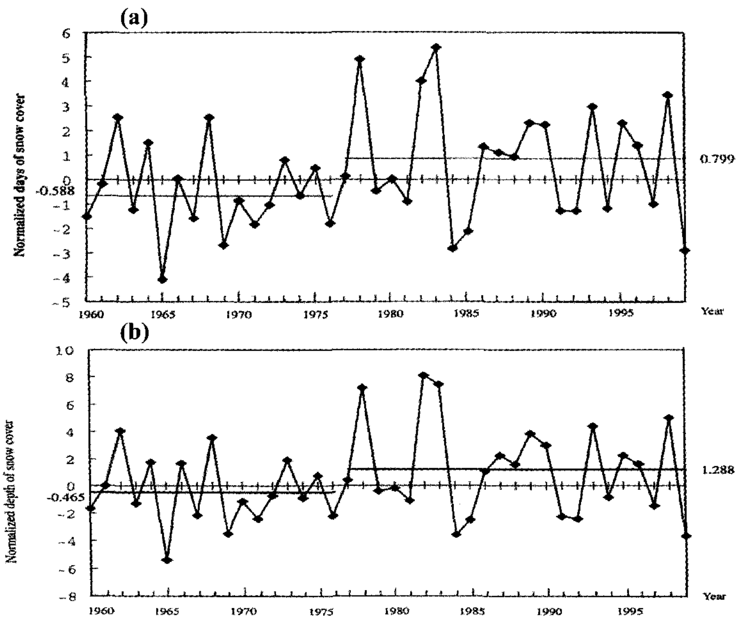


Figure 5.11. Interannual variations of the normalized anomalies of (a) days and (b) depth of snow cover over 0.5 cm during the period of the last October-May averaged for 72 observational stations in the Tibetan Plateau. The climatological mean values of days and depth of snow cover over 0.5cm during the period of the last October-May averaged for 1965-1999 are taken as the respective normals.

5.7. *Impact of the East Asian Winter Monsoon (EAWM) on the Summer Monsoon over China*

East Asia is also a region of strong winter monsoon. The winter monsoon features strong northwesterlies over North China and Northeast China, Korea and Japan and strong northeasterlies along the coast of China (e.g., Staff members of Academia Sinica 1957; Chen *et al.* 1991; Ding 1994). The strong winter monsoon can bring disasters such as low temperature, severe snow storms in Northwest China and Northeast China, North Korea and North Japan in winter, and severe sand dust-storms in North China, Northwest China and Korea in spring. Moreover, the Asian winter monsoon can cause strong convective activities over the maritime continent of Borneo and Indonesia (e.g., Chang *et al.* 1979; Lau and Chang 1987). Besides, strong and frequent activities of cold waves caused by a strong EAWM may trigger the occurrence of an El Niño event (see Li 1988). Tomita and Yasunari (1996) also pointed out that the EAWM might play a key role in the biennial oscillation of the ENSO/monsoon system. Thus, the EAWM may have an important impact on the EASM, and the interannual variability of the EAWM and its impact on the EASM are also an interesting scientific issue.

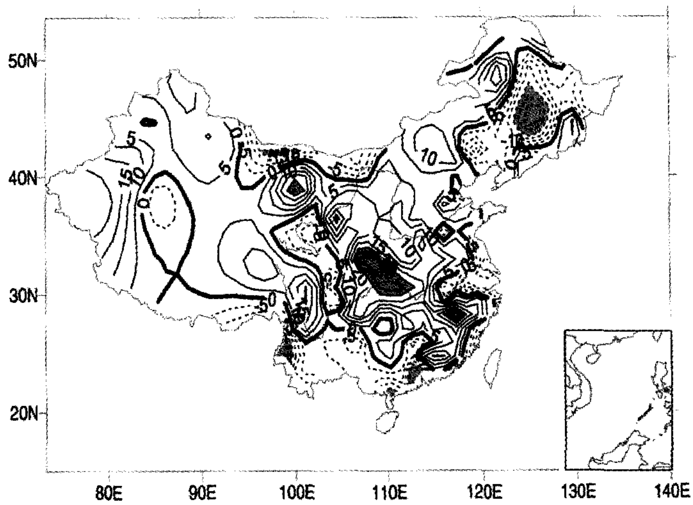


Figure 5.12. The correlation between the summer (June-August) rainfall in China and the normalized depth of snow cover averaged for 72 observational stations in the Tibetan Plateau. The thick black contour (contour interval: 5×10^{-2}) indicates zero correlation, and the solid and dashed contours denote positive and negative correlations, respectively, and the correlations over the 95% confidence level are shaded.

The East Asian summer monsoon and winter monsoon are a phenomenon of the annual cycle in both wind field in the lower troposphere and rainfall over East Asia (e.g., Chen *et al.* 2000). The anomalous EAWM can influence the following EASM (e.g., Sun and Sun 1994; Chen *et al.* 2000).

Chen and Graf (1998) and Chen *et al.* (2000) systematically investigated the interannual variability of the EAWM and its relation to the EASM with a new definition of the EAWM index. Chen *et al.* (2000) pointed out that after a strong EAWM, since the western Pacific subtropical high will shift northward in the following summer, generally a drought summer could occur in the middle and lower reaches of the Yangtze River. On the other hand, the western Pacific subtropical high will shift southward and a flood summer will occur in the Yangtze River and Huaihe River valley following a weak EAWM. For example, a severe flood occurred in the Yangtze River valley in the summer of 1998 following the anomalously weak EAWM in the winter of 1997 and the spring of 1998. This confirms the connection between the EAWM and the following EASM. As for the interannual variability of the EAWM and its impact on the EASM variability, these have been described in another chapter and will be omitted here.

6. Interdecadal Variability of the East Asian Climate System and its Impact on the Summer Monsoon over China

As described in Section 4, the interdecadal variability of the summer monsoon over China is significant, and has caused severe droughts in North China. In order to investigate the causes of the interdecadal variability of the summer monsoon over China and its surrounding regions, the interdecadal variability of the East Asian climate system will be analyzed by using the observed data.

6.1. *Interdecadal Variability of the SST Anomaly in the Tropical Pacific and its Impact on the Summer Monsoon over China*

The SST anomalies in the tropical Pacific can influence not only the interannual variability of the summer monsoon over East Asia, but also its interdecadal variability. In order to study the interdecadal variability of the SST anomaly in the tropical Pacific and its impact on the summer monsoon over China, the nine-year running mean SST anomalies in the equatorial Pacific (averaged between 5°S - 5°N) are analyzed using the database of QISST, provided by the Hadley Center, UK. As shown in Fig. 6.1, the SST anomalies display an obvious interdecadal variability in the equatorial central and eastern Pacific, and the SST in the equatorial central and eastern Pacific were below normal from the mid 1960s to the mid 1970s and remarkably increased during the early and middle 1980s and the 1990s in these regions. In order to show this variability clearly, the difference between the summer SST anomalies averaged for 1977-2000 and those averaged for 1967-1976 in the Pacific is also analyzed (Fig. 6.2). As shown in Fig. 6.2, an obvious El Niño -like SST anomaly pattern appeared in the tropical central and eastern Pacific in the period from 1977 to 2000 (e.g., Huang 2001). This may explain that a "decadal El Niño event" seems to have occurred from the late 1970s to now, while a "decadal La Niña event" occurred in the 1970s. Therefore, there seems to be a "decadal ENSO-like cycle" in the interdecadal variability of SST anomalies of the tropical Pacific.

The interdecadal variability of the summer monsoon over China and its surrounding regions may be in close association with the interdecadal variations of SST anomalies in the tropical Pacific. However, because it is difficult to reproduce the relationship between the monsoon and ENSO obtained from the observed data in climate models (e.g., Webster *et al.* 1998), this close relationship cannot be explained well with numerical simulations so far. Thus, it can only be simply discussed with correlation analysis. Figure 6.3 is the correlation between the summer (June-August) rainfall anomalies over China and the difference of the SST anomalies in the area of NINO.3 between summer and last autumn during 1951-1996 (e.g., Huang *et al.* 1998). From Fig. 6.3, it is seen that the positive correlation appears in the Yangtze River and Huaihe River valley and the lower reaches of the Yangtze River, and the correlation coefficients between them exceed the 95% significance level in these regions. Moreover, the negative correlations are located in North China and the area to the South of the Yangtze River, and these correlations also exceed the 95% significance level in these regions. Although Fig. 6.3 is the correlation between the summer rainfall in China and the

SST anomaly in the equatorial eastern Pacific on the interannual timescale, it may be used to estimate the relationship between them on the interdecadal timescale. The figure also shows that when the equatorial central and eastern Pacific is in a decadal warming episode, the summer monsoon rainfall may be strong in the Yangtze River and Huaihe River valley and the lower reaches of the Yangtze River and weak in North China. From the late 1970s to the

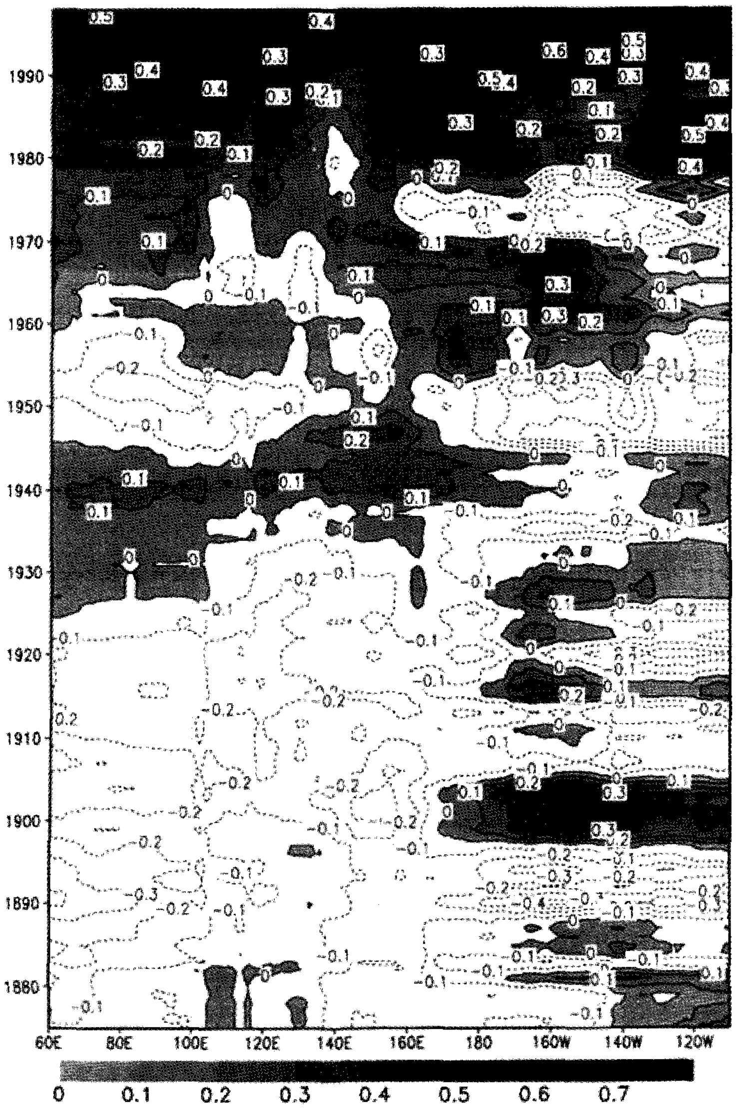


Figure 6.1. The longitude-time cross section of the 9-years running-mean SST anomalies along the equatorial Pacific (averaged between 5°S - 5°N). Units: °C. The solid and dashed contours in the figure denote positive and negative SST anomalies, respectively, and the shaded areas indicate the periods of positive SST anomalies. The dataset of QISST, Hadley Center, U.K. is used in this analysis.

1990s, the SST anomalies in the area of NINO.3 were positive, and according this correlation, the summer monsoon rainfall might be strong in the Yangtze River and Huaihe River valley and weak in North China during this period. Thus, the prolonged droughts that occurred in North China from the late 1970s to the 1990s may be closely associated with the obvious warming trend in the equatorial central and eastern Pacific from the late 1970s to the 1990s. In fact, the prolonged droughts still continue to appear in North China.

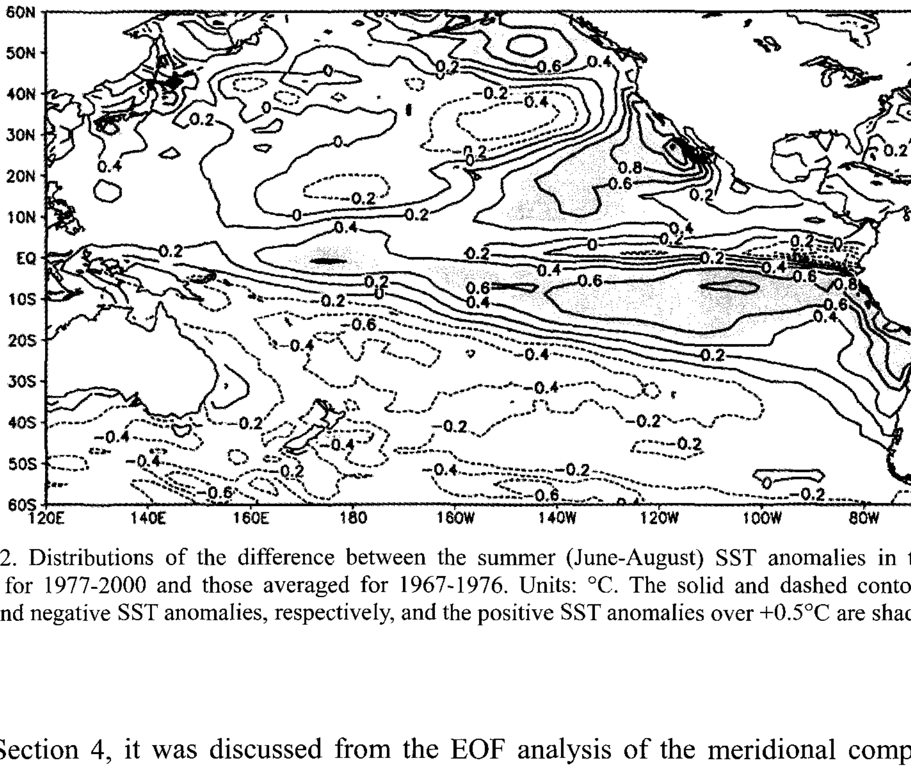


Figure 6.2. Distributions of the difference between the summer (June-August) SST anomalies in the Pacific averaged for 1977-2000 and those averaged for 1967-1976. Units: °C. The solid and dashed contours denote positive and negative SST anomalies, respectively, and the positive SST anomalies over +0.5°C are shaded.

In Section 4, it was discussed from the EOF analysis of the meridional component of water vapor transport that there was a serious decrease of water vapor transport by the EASM in North China from the late 1970s. What caused the decrease of water vapor transport by the summer monsoon in North China from the late 1970s? This may be closely related to the interdecadal variations of SST in the tropical Pacific. Influenced by the interdecadal variations of SST in the tropical Pacific, the interdecadal anomaly distribution of water vapor transport flux over East Asia and the tropical western Pacific after the late 1970s is different from that before the late 1970s. Figures 6.4a and 6.4b are the anomaly distributions of water vapor transport flux averaged for the period of 1967-1976 and for the period of 1977-1999, respectively. In the figures, the climatological mean distribution of water vapor transport averaged for 1961-1990 is taken as the normal. From Fig. 6.4a, it may be seen that during the period of 1967-1976, the westward transport flux anomalies of water vapor by the trade winds along the equator were strong, and the northward transport flux anomalies of water vapor appeared over the South China Sea and around the Philippines. Moreover, the water vapor anomaly fluxes were transported from the Bay of Bengal toward the Indo-China Peninsula and South China. Therefore, a large amount of water vapor was transported into North China from the tropical western Pacific, the South China Sea and the Bay of Bengal in the period of 1967-1976. This provided the condition of sufficient water vapor transport for the strong summer rainfall in North China during 1967-1976.

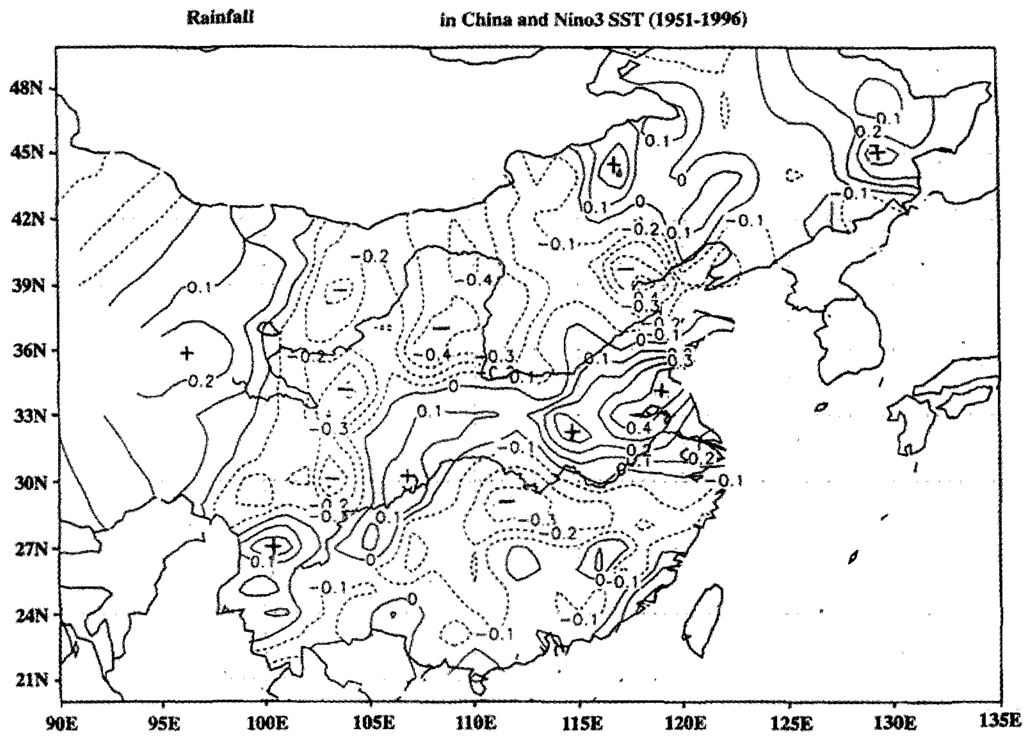


Figure 6.3. Correlation between the summer (June-August) rainfall anomalies in China and the differences of SST anomaly in the area of NINO.3 (i.e., 150°W - 90°W, 4°S - 4°N) between summer and the preceding autumn from 1951 to 1996. The areas of solid and dashed contours in the figure denote the positive and negative correlations, respectively.

However, in the period of 1977-1999, influenced by the warming of the tropical Pacific as shown in Fig. 6.2, the trade winds became weak over the tropical Pacific, and the distribution of water vapor transport anomalies shown in Fig. 6.4b is opposite to that during 1967-1976 as shown in Fig. 6.4a. Since the trade winds became weak over the tropical Pacific from the late 1970s, the eastward transport flux anomalies of water vapor appeared over the tropical Pacific, and the southward transport flux anomalies of water vapor appeared over the South China Sea and around the Philippines. Moreover, the water vapor anomaly fluxes as shown in Fig. 6.4b show the weak water vapor transport from the Bay of Bengal to the Indo-China Peninsula and South China. Because the southward transport flux anomalies of water vapor appeared over North China and East China, as shown in Fig. 6.4b, the water vapor transported into the eastern part of China from the tropical western Pacific, the South China Sea and the Bay of Bengal became weak from the late 1970s. As a consequence, the water vapor transport by the summer monsoon flow became weak in North China, and a large amount of water vapor converged in the Yangtze River valley. This caused the remarkable decrease of summer precipitation in North China and the significant increase of summer precipitation in the Yangtze River valley from the late 1970s to the 1990s.

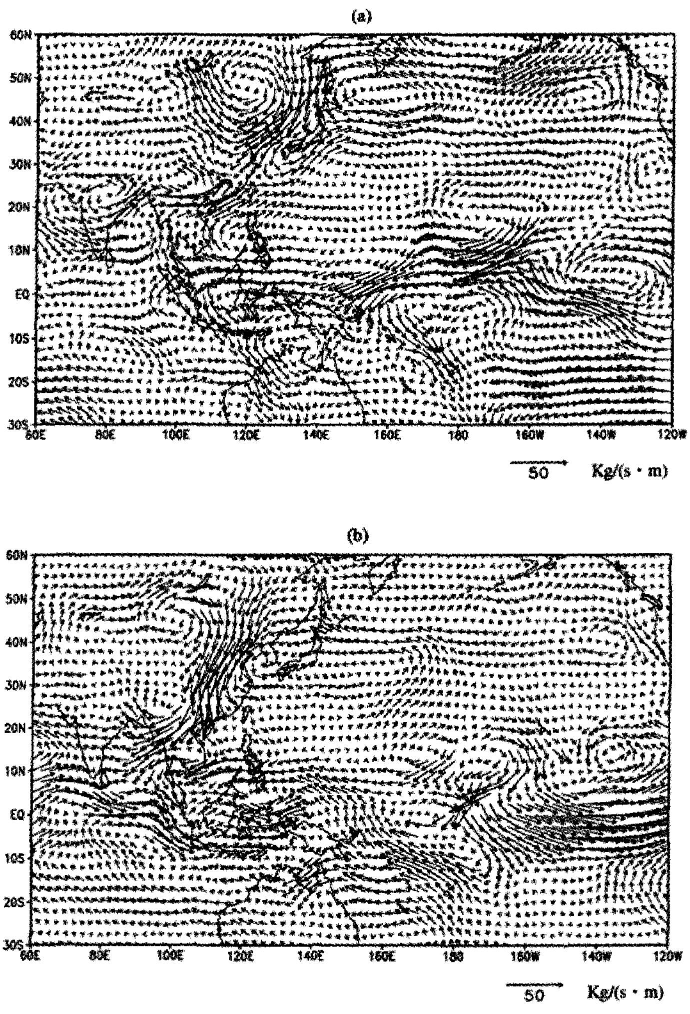


Figure 6.4. The anomaly distributions of water vapor transport fluxes averaged (a) for the summers of 1967-1976, and (b) for the summers of 1977-1999. The climatological mean distribution of water vapor transport fluxes for 30 years from 1961 to 1990 are taken as the normal.

6.2. *Interdecadal Variability of Snow Cover in the Tibetan Plateau and its Impact on the Summer Monsoon over China*

As shown by Chen *et al.* (2000), the EAWM became weak from the late 1980s. According to Tao and Zhang's (1998) study, if the EAWM is weak in a winter or spring, the disturbances over the south side of the Tibetan Plateau tend to be more active, which cause stronger rainfall or snowfall in the Tibetan Plateau. In fact, in comparing the time and depth of snow cover in the Tibetan Plateau in the period from the late 1970s to the late 1990s with those in the period from the 1960s to the mid 1970s as shown in Figs. 5.11a and 5.11b, the time of snow cover has become longer and the snow depth has become deeper in the Tibetan Plateau from the late 1970s.

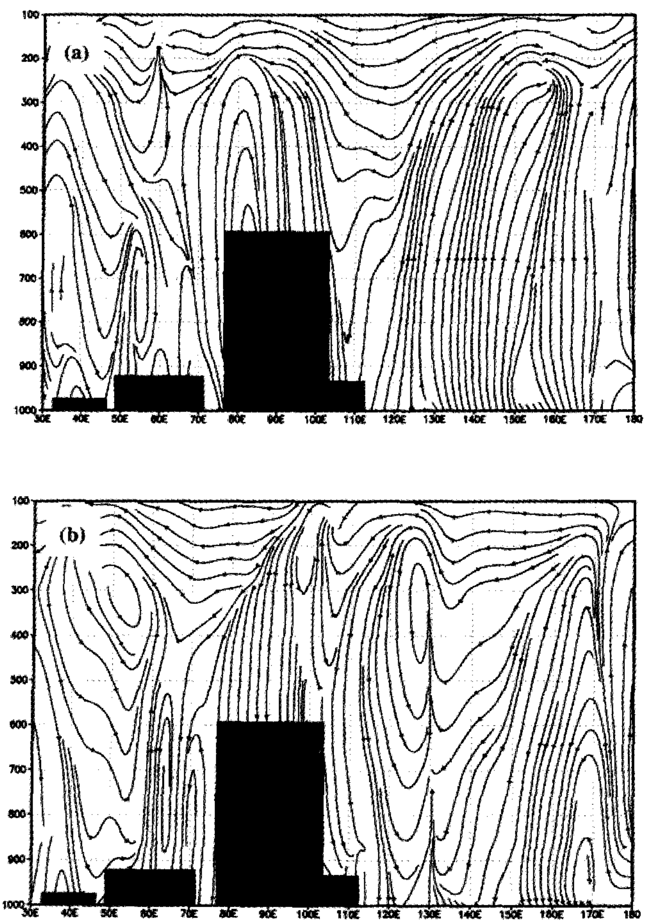


Figure 6.5. The composite anomaly distributions of the vertical-zonal circulation circle along 30°N for (a) the summers of strong snow years and (b) for the summers of weak snow years in the Tibetan Plateau. The climatological mean distribution of vertical-zonal circulation circle along 30°N averaged for 35 summers from 1965 to 1999 is taken as the normal.

The interdecadal variability of snow cover in the Tibetan Plateau in winter and spring may have a significant effect on the interdecadal variations of summer monsoon rainfall in the central and eastern parts of China, especially in the middle and upper reaches of the Yangtze River. As shown in Fig.5.12, the correlations between the summer rainfall and the normalized depth of snow cover in the Tibetan Plateau in winter and spring exceeded the 95% confidence level in these regions. Since the time of snow cover became longer and the depth of snow cover became deeper in the Tibetan Plateau from the late 1970s to the late 1990s, it may be also explained from Fig. 5.12 that the summer rainfall became stronger in the middle and upper reaches of the Yangtze River.

Why can the snow cover in the Tibetan Plateau in winter and spring influence the following summer monsoon rainfall in the Yangtze River valley on the interdecadal timescale? This problem has not been well studied up to now, especially using numerical simulations. Thus, it can only be simply explained from the observed data. Based on a preliminary investigation, the snow cover in the Tibetan Plateau has an important influence on the vertical-zonal circulation circle in the east and west sides of the Tibetan Plateau. Figures 6.5a and 6.5b are the composite anomaly distributions of the vertical-zonal circulation circle along 30°N for the summers of strong snow years and for the summers of weak snow years as defined in Section 5, respectively. In the figures, the vertical-zonal circulation along 30°N averaged for the period of 1965-1999 is taken as the climatological mean. From Fig. 6.5a, it may be seen that when the depth of snow cover in the Tibetan Plateau is deeper from the preceding winters to springs during a period, a stronger ascending flow may appear in the area from 110°E to 160°E in the following summers during the period. This may show that when the snowfall in the Tibetan Plateau is heavy in winters or springs during a period, a stronger ascending flow will appear in the east side of the Tibetan Plateau during the following summers. This may cause strong summer rainfall in the middle and upper reaches of the Yangtze River during the period. Since the time of snow cover became longer and the depth of snow cover became deeper in the Tibetan Plateau from the late 1970s, according to the composite distribution of vertical-zonal circulation anomalies shown in Fig. 6.5a, the stronger ascending flow might appear in the area from the middle and upper reaches of the Yangtze River to the subtropical western Pacific from the late 1970s. This may cause strong convective activities and strong monsoon rainfall in these regions. Therefore, the interdecadal variability of snow cover in the Tibetan Plateau also has a significant influence on the interdecadal variability of summer monsoon rainfall in the Yangtze River valley.

7. Summary and Discussion

It is seen from the above review that significant progress in recent research on the intraseasonal, interannual and interdecadal variations of the summer monsoon over China and its surrounding regions has been achieved. Moreover, the causes and physical processes of these variabilities have been preliminarily sought. Through these studies, the seasonal prediction of the summer monsoon over China has been improved to some extent and the predictability of the interannual and interdecadal variabilities of the summer monsoon over

East Asia has been understood further. However, since climate variations of the summer monsoon over East Asia and their physical mechanism are very complex, there are many unknown phenomena in the intraseasonal, interannual and interdecadal variabilities of the summer monsoon over China and its surrounding regions, and the causes and physical processes of these variabilities are still unclear so far. Besides, the summer monsoon rainband can not be simulated well in GCMs or coupled models. Therefore, the following problems need to be studied further:

(1) Interaction between the monsoon activity and the low-frequency oscillation in East Asia.
(2) Internal-dynamic process of monsoon variability in East Asia.
(3) Physical mechanism of the interannual variability of the EASM and its linkage with the EAWM as a phenomenon of the annual cycle.
(4) Processes of the interaction between the Asian monsoon and ENSO cycle and its reproduction in coupled models.
(5) Modeling of the EASM, especially the summer monsoon rainband.
(6) Causes of the interdecadal variability of the EASM and its impact on interannual variability.

The authors believe that with the implementation of the CLIVAR Program, the above-mentioned problems related to climate variations of the summer monsoon over China will be well investigated from observational, theoretical and modeling aspects in the early part of the 21st century.

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