The Characteristics of Interannual Variations on the East Asian Monsoon

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Abstract

By utilizing the meteorological observational data over the mainland of China and surrounding oceans from 1951 to 1990, the characteristics of the climatic trend of East Asian Monsoon and the interannual variation of monsoon overlapped on the general trend have been studied. The results show that during the past 40 years, especially in the 1980s, China has generally been drying and the areas south of 35°N in China are getting cooler. This may be attributed to the weakening of the winter and summer monsoon in China. Mei-Yu in middle and lower reaches of Yangtze River and the precipitation in other areas of China are affected by the climatic change trend.

It has long been believed that the Indian Monsoon is correlative to the summer precipitation of China. But in our studies, it can be concluded that there is a significant correlation only between the summer precipitation over India and that over the western part of North China, and for other areas of China the correlation is not obvious.

The interannual anomaly of the summer monsoon in China is mainly attributed to the variation of each member of the circulation system of the East Asian summer monsoon. The interactions of air-sea and air-land have also been discussed. Although a significant correlation can not be found between the SST of equatorial eastern Pacific/Polar ice and the winter/summer monsoon of China, they all have similar significant interannual oscillations and the correlation between these interannual oscillation components of the same frequency is very high.

1. Introduction

The East Asian Monsoon circulation has not only the intraseasonal and seasonal oscillations but also obvious interannual variations. The interannual variations of monsoon are often shown in the variation of the climatic trend. In these climatic trends, there exist the year-to-year variation of the seasonal transition, the interannual variation of the amplitudes of the intraseasonal oscillations. The factors which caused the interannual variations of the monsoon are very complicated and varied. Besides the air-sea interaction, land surface effects and other external forcing (*e.g.*, volcano and so on) should also be considered. Before discussing the interannual variation of the East Asian Monsoon, we need to study the climatic trend for the last several decades. For example, the summer monsoon rainfall of North China has been decreasing since the 1950's. There is an apparent trend toward a more arid climate. This indicates that the summer monsoon of North China is weakening. The summer rainfall in the middle and lower Yangtze River basin mainly comes from the Mei-Yu (plum rains, Bai-u in Japanese). The in-

terannual variation of the beginning and ending of the Mei-Yu is a major factor determining the variation of the yearly Mei-Yu amount. To discuss the interannual variation of the East Asian Monsoon, we must consider the interannual variability of the seasonal transition. The interannual variation of the amplitudes and phases of the seasonal and intraseasonal oscillation can determine the interannual variability of summer rainfall in various places.

It is well known that there are not only apparent seasonal and intraseasonal oscillations but also interannual oscillations. For example, there is a quasi-biennial oscillation (QBO) in the tropospheric wind field (Yasunari, 1985), interannual oscillations in the SST of the Pacific Ocean (Yan *et al.*, 1989a, Chen *et al.*, 1989b), interannual oscillation in precipitation and temperature of the mainland of China (Chen L. and Chen D., 1990), and interannual oscillations in areas of Arctic ice (Chen Y. and Chen L., 1991). This implies that the interannual oscillations may be the important processes of the interannual variability of the monsoon, and that it is important to discuss the mechanism of the interannual variability of the monsoon through the study of the interactions between the above oscillations. In addition, the in-

Fig. 1. The outline map of China. The numbers represent the names of some administrative areas. 1-Heilongjiang; 2-Nei Menggu; 3-Jilin; 4-Liaoning; 5-Beijing; 6-Tianjin; 7-Hebei; 8-Shandong; 9-Henan; 10-Jiangsu; 11-Anhui; 12-Shanghai; 13-Zhejiang; 14-Jiangxi; 15-Hubei; 16-Hunan; 17-Fujian; 18-Guangdong; 19-Hainan; 20-Guangxi; 21-Guizhou; 22-Sichuan; 23-Yunnan; 24-Xizang; 25-Shaanxi; 26-Gansu; 27-Shanxi; 28-Ningxia; 29-Qinghai; 30-Xinjiang.

teraction between the interannual oscillation and the seasonal and intraseasonal oscillations is also very important in determining the interannual variation of the monsoon circulation. This interaction is also very important in detecting the formation mechanism of the seasonal and interannual oscillations of the monsoon. Murakami *et al.* (1986) have studied the relationship among the transient oscillation, the 30–60 day oscillation and the seasonal oscillation and found that there is a close linkage among them. So we should study not only the interannual variation itself but also its interaction with the seasonal variation and intraseasonal (30–60 day) oscillation.

2. The climatic trend of the East Asia Monsoon

There is no single good physical parameter to represent the intensity and interannual variability of the East Asian Monsoon. Most Chinese meteorologists use the following four physical quantities to represent the intensity of a monsoon:

- (1) Summer rainfall: representing the intensity of the summer monsoon.
- (2) Winter temperature: representing the intensity of the winter monsoon.
- (3) Monsoon index (MI): the mean pressure differences between 110°E and 160°E in the area from 10°N to 50°N (Guo, 1988a). The MI represents the pressure difference between the ocean and the continent.
- (4) Zonal wind index (ZWI): the mean difference of geopotential between certain two

latitude lines in the East Asian region at 500 hPa or 200 hPa. This is equivalent to the zonal wind intensity over the East Asian region.

Chen *et al.* (1991b) have studied the characteristics of the interannual variation of monthly rainfall and temperature of China using the data of 160 stations from Jan. 1951 to Dec. 1989 (among them 34 stations have data earlier than 1950). The results for the yearly temperature show that, since the beginning of this century, for the whole of China it was warmest in the 1940's. Subsequently it became cooler, they slightly warmer (since the 1970's). But the temperature in the 1980's has not yet exceeded the level of the 1940's. In more detail, the temperature in Northeast China, North and Northwest China (see Fig. 1), which are in the areas north of 35°N, was higher in the 1980's than in the 1950's. But the area south of 35°N, north of 25°N and east of the Tibetan Plateau was cooler in the 1980's than in the 1950's. The center of this cooling region is located in the Sichuan province and the south part of Shaanxi. This cooling is not consistent with the global warming. Table 1 shows the ten-year mean temperature, climatic mean temperature during the last 39 years, and the difference of temperature between the 1980's and the 1950's for most administrative areas of China. Table 2 is the same as Table 1 except for precipitation. It can be seen that the temperature decreased in the all provinces of the Yangtze River basin. Sichuan province has cooled by about 0.48°C.

Table 1. 10-year mean temperature (°C) in China.

1	2	3	4	5	6	7
N	51-60	61-70	71-80	81-89	39YM	80S-50S
Heilongjiang	1.89	1.93	2.18	2.67	2.16	0.78
Nei Menggu	2.85	3.07	3.37	3.67	3.23	0.82
Jilin	4.77	4.90	5.11	5.42	5.04	0.65
Liaoning	8.62	8.76	8.93	9.26	8.88	0.64
Beijing	11.55	11.64	11.34	12.43	11.72	0.88
Tianjin	12.28	12.26	12.30	12.64	12.36	0.36
Hebei	10.36	10.67	10.85	11.20	10.76	0.84
Shandong	12.75	12.76	12.90	13.24	12.90	0.49
Henan	14.30	14.47	14.46	14.44	14.42	0.14
Jiangsu	14.31	14.45	14.36	14.30	14.36	-0.01
Anhui	15.62	15.76	15.69	15.67	15.68	0.05
Shanghai	15.67	15.68	15.65	15.90	15.72	0.23
Zhejiang	16.89	16.90	16.85	16.80	16.86	-0.09
Jiangxi	17.94	18.19	18.03	18.03	18.05	0.09
Hubei	16.27	16.27	16.25	16.07	16.22	-0.20
Hunan	17.18	17.35	17.20	17.12	17.22	-0.06
Fujian	19.21	19.28	19.24	19.05	19.20	-0.16
Guangdong	21.63	21.59	21.57	21.68	21.61	0.05
Hainan	23.95	23.63	23.81	23.94	23.83	-0.01
Guangxi	21.10	21.09	21.06	21.08	21.09	-0.02
Guizhou	15.36	15.30	15.29	15.20	15.29	-0.16
Sichuan	15.34	15.12	15.07	14.86	15.10	-0.48
Yunnan	15.27	14.95	14.95	15.26	15.10	-0.01
Xizang	7.70	7.24	7.58	7.65	7.54	-0.05
Shaanxi	12.22	12.08	12.21	12.03	12.14	-0.19
Gansu	8.19	7.83	7.98	8.05	8.01	-0.14
Shanxi	10.05	10.20	10.45	10.70	10.24	0.65
Ningxia	8.87	8.73	8.86	9.01	8.86	0.14
Qinghai	1.52	1.36	1.52	1.83	1.55	0.31
Xinjiang	9.14	9.50	9.53	9.73	9.47	0.59

Column 1: name of administrative area;
Column 2-4: 10-year mean value;
Column 5: 9-year mean value; Column 6: 39-year mean value;
Column 7: difference between 1980s and 1950s.

Figure 2a and 2b show the temperature anomalies of winter (DJF) and summer (JJA) in the 1980's in China. It can be seen that in winter (Fig. 2a) there are negative anomalies in the mainland of China south of 35°N and east of 102°E with several maximum negative anomaly centers of about 0.4°C in Sichuan, Henan, Guangdong and Zhejiang. Outside of the above-mentioned regions are positive anomalies. There are warming regions in Northeast China, North China and Xingjiang with an amplitude of more than 1°C. In summer, there are negative anomalies in the region east of 102°E, between 26°N and 40°N and in Northeast China and part of Xingjiang east of the Tianshan mountain. The cooling amplitude in these regions is more than 1.2°C.

The precipitation of China as a whole was large in the 1940's-1950's and then decreased gradually.

Compared with the 1950's, the precipitation of the 1980's decreased in all parts of China except the middle and lower Yangtze River basin and areas very close to the north of it. Figure 3 gives the difference of the ten-year mean of yearly precipitation between the 1980's and the 1950's. It is obvious that in the region north of the Yangtze River basin and south of the Yellow River basin, the precipitation increased. In other regions the precipitation decreased. The maximum of the precipitation decrease is located in North China (Shandong and Hebei provinces) and regions south of the Yangtze River (Hunan, south part of Zhejiang and Jiangxi). The precipitation decrease in North China is about 20 % or more of the 39-year mean value. This indicates that most parts of China are becoming drier. The precipitation distribution in summer is consistent with that for a

Table 2. 10-year means of yearly precipitation amount (mm) in China.

1	2	3	4	5	6	7
N	51–60	61–70	71–80	81–89	39YM	80S–50S
Heilongjiang	555	507	456	545	515	–10
Nei Menggu	377	340	338	348	351	–29
Jilin	690	657	647	678	668	–12
Liaoning	747	717	674	654	699	–93
Beijing	782	584	568	568	626	–214
Tianjin	562	558	595	520	559	–42
Hebei	550	535	493	440	506	–110
Shandong	699	725	718	560	678	–139
Henan	803	799	774	769	786	–34
Jiangsu	992	976	943	903	955	–89
Anhui	1234	1081	1180	1204	1174	–30
Shanghai	1218	1040	1114	1176	1136	–42
Zhejiang	1645	1420	1540	1572	1544	–73
Jiangxi	1624	1551	1564	1564	1576	–60
Hubei	1123	1139	1079	1192	1131	69
Hunan	1386	1384	1326	1296	1349	–90
Fujian	1516	1395	1468	1457	1459	–59
Guangdong	1713	1607	1775	1771	1715	58
Hainan	1776	1603	1675	1632	1672	–144
Guangxi	1474	1467	1520	1442	1477	–32
Guizhou	1123	1154	1175	1052	1128	–71
Sichuan	1116	1096	1055	1085	1088	–31
Yunnan	960	1008	981	941	973	–19
Xizang	478	487	436	413	455	–65
Shaanxi	651	666	619	689	655	38
Gansu	314	341	317	319	323	5
Shanxi	522	551	516	491	521	–31
Ningxia	216	216	208	194	209	–22
Qinghai	384	380	390	398	388	14
Xinjiang	129	105	109	125	117	–4

Column 1: name of administrative area;
Column 2–4: 10-year mean value;
Column 5: 9-year mean value; Column 6: 39-year mean value;
Column 7: difference between 1980s and 1950s.

whole year. Thus, in general, the summer monsoon in China is getting weaker.

The correlations between yearly precipitation and temperature in the last 39 years are negative for most of the stations, *i.e.* a warm year has less precipitation and cool year has more precipitation (Fig. 4). There are two negative correlation centers, one is in North China and Northeast China and the other lies between the Yangtze and Yellow Rivers. The significance level of the above negative correlation centers is more than 99 %.

It is generally considered that the thermal contrast between the continent and ocean is the main factor causing the monsoon phenomena. The contrast between the air temperature over the continent and the sea surface temperature can be used to approximately represent the thermal contrast between

the land and the sea. The temperature anomaly in the 1980s over continent has been shown in Fig. 2, but what about the SST anomaly in the surrounding sea area? Figure 5 gives the mean air temperature curves of winter and summer for regions east of 100°E, south of 35°N and north of 35°N over the continent, and the SST curves of winter and summer for regions north and south of 35°N in the Pacific west of 180°E. It can be seen that the sea surface temperature in regions north of 35°N decreased while for the regions south of 35°N in the western Pacific the sea surface temperature increased. This makes obvious contrast with the temperature evolution over the continent of China.

For summer, since the temperature over the eastern part south of 35°N of China obviously decreased, and at the same time the sea surface temperature

Fig. 2. The distribution of winter (Dec.-Feb.) (a) and summer (June-Aug.) (b) temperature anomalies in China in the 1980s.

in the sea east of the continent increased slightly, the difference of temperature between the sea and continent reduced. This is unfavorable for summer monsoon formation and may be the reason for the decrease of the summer monsoon rainfall in the areas south of Yangtze River basin. The warming of the continent and cooling of the sea in the area north of 35°N makes the thermal contrast of sea-land stronger. This is unfavorable for the development of a cold air mass over the continent. The summer rainfall in North China is determined mainly by the southerly monsoon current and northerly cold air, so the weakening of the southerly summer monsoon flow and northerly cold air is favorable for a decrease in summer precipitation in North China.

For winter, the warming over the continent north of the Yangtze River and cooling in south, combined

with the cooling of the ocean in north and warming in south, make the winter monsoon of North China weak. In general, during the last 40 years, both the winter and the summer monsoon in East Asia have weakened. The interannual variability of the East Asian Monsoon during last 40 years is superimposed over the above climatic trend.

Guo (1988a) calculated the summer Monsoon Index (SMI) for the period from 1871 to 1981. The results show that the SMI during 1951–1959 was close to the normal, and the SMI of 1960–1967 slightly increased and after 1967 the SMI decreased. This indicates a weakening tendency of the East Asian Monsoon and is consistent with the above analysis shown by Fig. 5.

The above climatic trend of the East Asian Monsoon is unique. Shukla *et al.* (1987) gave the anomaly

Fig. 3. The difference of ten-year means of yearly precipitation(mm) between the 1980s and the 1950s.

Fig. 4. The correlation distribution between yearly mean temperature and yearly precipitation amount during the period of 1951–1989 in China. The contour is the confidence line: the values of 0.05, 0.01 and 0.001 correspond to a confidence of 95 %, 99 % and 99.9 %, respectively. Plus and minus signs indicate positive and negative correlations, respectively.

lies of the summer monsoon rainfall in various places of India up to 1984. There is no such obvious climatic trend as in the East Asian Monsoon area. Many scientists have studied the relationship between the Indian Monsoon and the East Asian Monsoon, but the conclusions are contradictory. Guo (1988b) calculated the correlation coefficient between the summer monsoon rainfall over the whole of India and various stations over China. It is pointed out that there are no significant correlations between India and most of the stations in China, ex-

cept some North China stations. Figure 6 is a map of the correlation coefficient between the summer monsoon rainfalls over the middle part of India and 200 stations of China. The contours in the map indicate the significance level of the correlation coefficients. The real correlation coefficient, corresponding to a 95 % significance level is 0.36, while one corresponding to a 99 % significance level is 0.48 for the total of 30 samples. It can be seen that only in the western part of China, from Shaanxi to Sichuan province there is a positive correlation maximum. In Hebei

Fig. 5. The curves of air temperature over mainland China and the sea surface temperature in the western Pacific during the period 1951–1989. The abscissa and ordinate are years and the anomalies ($^{\circ}\text{C}$) of air temperature or SST. The solid line represents the yearly anomalies while the dashed line is its linear trend.

- (a) Summer temperature north of 35°N and east of 100°E in China;
- (b) Summer temperature south of 35°N and east of 100°E in China;
- (c) Summer SST north of 35°N and west of 180°E in the western Pacific;
- (d) Summer SST south of 35°N and west of 180°E in the eastern Pacific;
- (e) Winter temperature north of 35°N and east of 100°E in China;
- (f) Winter temperature south of 35°N and east of 100°E in China;
- (g) Winter SST north of 35°N and west of 180°E in the western Pacific;
- (h) Winter SST south of 35°N and west of 180°E in the western Pacific.

province there is a small correlation center. And outside the above region there is no significant correlation. It is known that Shaanxi province is very close to the western boundary of the regions in which the summer monsoon prevails. The correlations for the regions with high summer monsoon rainfall,

such as South China, the Yangtze River basin and the southern part of North China, are not significant.

Fig. 6. The correlation distribution of summer precipitation amounts between the middle part of India and China.

3. The interannual variation of the summer monsoon in China

As described in the last section, the winter and summer monsoon of China can be expressed by the winter temperature and summer rainfall respectively. In summer, the circulation systems constituting the East Asian Monsoon are very different from those constituting the Indian Monsoon (Chen *et al.*, 1991a). For the Indian Monsoon system, there is only one rainbelt corresponding to the monsoon trough which is of a tropical nature, while in the East Asia Monsoon system there are two rainbelts. One is linked with a monsoon trough in the South China Sea and the ITCZ in the western Pacific (it may move to the south coast of the mainland of China in July and August) and has a tropical nature, and the other is formed by the joint contribution of the cold air mass from the polar continent and warm and wet flow on the north side of the subtropical high. This rainbelt is known as the Mei-Yu (plum rain, or Bai-u in Japan). The mainland of China is mainly affected by the second rainbelt which has a subtropical nature. To the south of this rainbelt there is a prevailing southwesterly flow which may come from the south side of the subtropical high and turns to a southwesterly on the west side and northwest side of the subtropical high, or may come from the South China Sea or the Bay of Bengal. This means that the southwesterly flow on the south side of the Mei-Yu rain belt may have a subtropical or the tropical nature. Chen *et al.* (1991a) have discussed the circulation pattern of the tropical monsoon influencing the continent of China. The onset of Mei-Yu rain season begins in South China in mid-May in general. This is referred to by Chinese

meteorologists as the early rainy season of South China and is subtropical in nature. There is also another rainy season in the mid-summer in South China which is mainly caused by typhoons and other tropical systems, and is referred to as the late rainy season in order to distinguish it from the former. In the second dekad (10 days) of June, the rainbelt jumps to the Yangtze River basin. This rain is called the plum rain by Chinese scientists. In the first or second dekad of July the rainbelt moves northward to the Huaihe River basin, and finally in the second dekad of July the rainy season begins in North China and the Haihe River basin. The position of the second rainbelt at every dekad from spring to summer was shown in Tao and Chen (1987). This chart is constructed using the climatic mean value of dekad rainfall for 1951–1983.

Here we give the date of onset and ending of the Mei-Yu for the years since 1901, shown in Table 3, and the dekad date of the beginning and ending of the rainy season for Haihe River basin (Table 4). Table 3 comes from the Beijing Meteorological Center and is determined on the basis of the rainfall of five stations in the Yangtze River basin (Shanghai, Nanjing, Wuhu, Jiujiang and Hankou). Table 4 is formulated according to the precipitation over 40 stations in the Haihe River basin. It can be seen that, on the average, the Mei-Yu in the middle and lower Yangtze River basin begins on June 16 and ends on July 10.

The rainy season of the Haihe River basin area begins in the second dekad of July and ends in the second dekad of August and lasts about 3–4 dekads.

Since the summer rainbelt of China moves northward during the period from spring to summer and southward during the period from summer to au-

Table 3. The Date of Onset and Termination of Mei-Yu in the Middle and Lower Reaches of Yangtze River.

Year	Date of Onset	Date of Termination	Length of Mei-Yu (No. of Rainy Days)	Intensity of Rainfall*
1901	June 15	July 21	27	2.40
1902	—		Empty	0
1903	May 27	July 15	33	1.08
1904	—		Empty	0
1905	June 26	July 5	9	0.41
1906	June 3	July 13	40	1.34
1907	June 10	July 20	40	1.17
1908	June 8	July 16	38	1.72
1909	June 2	July 15	43	2.72
1910	June 4	July 21	47	1.86
1911	May 28	July 14	47	1.75
1912	June 24	July 20	26	1.35
1913	June 13	July 12	29	0.35
1914	June 14	June 28	14	0.27
1915	June 16	July 8	23	1.40
1916	June 13	July 18	35	1.69
1917	June 3	June 22	19	1.08
1918	June 7	July 10	33	1.31
1919	June 10	July 14	34	1.84
1920	June 8	July 9	21	0.86
1921	June 10	July 17	22	1.70
1922	June 24	July 11	17	0.68
1923	June 25	July 21	26	1.20
1924	June 16	July 8	22	1.14
1925	—		Empty	0
1926	June 21	July 17	26	1.06
1927	June 13	July 8	25	1.18
1928	June 19	June 30	11	0.56
1929	June 14	July 9	25	1.06
1930	June 12	July 3	21	1.72
1931	June 13	July 30	47	2.23
1932	June 11	July 11	30	0.96
1933	June 3	June 23	20	0.69
1934	—		Empty	0
1935	June 15	July 7	22	0.97
1936	June 14	July 6	22	0.51
1937	June 6	July 2	26	0.25
1938	June 1	July 6	35	0.32
1939	June 15	July 4	19	1.00
1940	June 14	June 30	16	0.78
1941	June 22	July 25	33	1.14
1942	June 16	July 9	23	1.21
1943	June 23	July 18	25	0.83
1944	June 23	July 5	13	0.61
1945	June 12	June 30	18	1.02
1946	June 10	July 10	30	1.03
1947	July 4	July 19	15	0.63
1948	June 20	July 24	32	1.16
1949	June 22	July 18	26	0.95
1950	June 21	July 9	18	0.45
1951	June 22	July 23	31	0.94

Table 3. (Continued)

Year	Date of Onset	Date of Termination	Length of Mei-Yu (No. of Rainy Days)	Intensity of Rainfall*
1952	July 2	July 16	14	0.27
1953	June 19	June 29	10	0.93
1954	June 12	Aug. 1	50	3.02
1955	June 21	July 9	18	0.98
1956	June 4	July 15	41	1.26
1957	June 19	July 10	21	0.93
1958	—		Empty	0
1959	June 27	July 6	9	0.36
1960	June 8	June 26	18	0.63
1961	June 6	June 16	10	0.60
1962	June 17	July 9	22	1.49
1963	June 23	June 30	7	0.38
1964	June 23	June 30	7	0.91
1965	—		Empty	0
1966	June 25	July 13	18	0.77
1967	June 23	July 10	17	0.52
1968	June 23	July 20	26	0.51
1969	June 23	July 21	28	2.36
1970	June 18	July 22	35	1.09
1971	June 9	June 15	6	0.21
1972	June 20	June 30	10	0.62
1973	June 16	June 26	10	0.48
1974	June 10	July 19	39	1.03
1975	June 16	July 17	31	1.21
1976	June 21	July 16	25	0.74
1977	June 9	June 18	9	0.36
1978	June 8	June 15	7	0.35
1979	June 19	July 23	35	1.23
1980	June 9	July 22	43	1.61
1981	June 25	July 2	7	0.61
1982	July 9	July 26	17	1.02
1983	June 19	July 19	30	1.60
1984	June 7	July 8	31	1.43
1985	June 22	July 7	15	0.44
1986	June 20	July 8	18	0.82
1987	July 1	Aug. 1	31	1.18
1988	June 10	June 23	13	0.45
1989	June 14	July 14	30	0.89
1990	June 19	July 3	14	0.54
Mean	June 16	July 10	24	

* : Intensity of rainfall = $\frac{\text{rainfall amount during active period of Mei-Yu}}{\text{normal rainfall amount during Mei-Yu}}$

tumn, the summer monsoon rainfall has strong regional-
ity. So we divide the whole of China into six
divisions and discuss the interannual variations of
precipitation by using their spatial mean rainfall of
summer (for South China we take the rainfall of
April–June which is the main rainy season and for
other divisions take the rainfall of June–August).
The six divisions are the following: South China
(12 stations in Fujian, Guangdong and Guangxi
provinces are taken), the Middle and lower Yangtze
River basin (8 stations in Hubei and Jiangsu
provinces), the Huaihe River basin (6 stations in
Anhui province), North China (6 stations in Bei-
jing, Tianjin and Hebei province), Northeast China
(10 stations in Heilongjiang, Jilin and Liaoning
provinces) and Southwest China (7 stations in
Sichuan and Guizhou provinces). Table 5 and Fig. 7
(right panel) give the standardized anomaly of sum-

Table 4. The Dekad Date of Onset and Termination of Rainy Season (≥ 50 mm) in Haihe River Valley.

Year	Onset	Termination	Anomalies during summer (June–Sept.)
	Month Date	Month Date	
1951	Aug. S	Aug. S	3 –
1952	July F	July T	2 –
1953	July S	Aug. T	0
1954	June F	Aug. S	2 +
1955	July F	Sept S	2 +
1956	June F	Aug. F	3 +
1957	June F	July S	2 –
1958	July S	Aug. F	2 +
1959	July F	Aug. T	2 +
1960	July S	July T	2 –
1961	July S	Aug. S	1 –
1962	July T	Aug. S	2 +
1963	July F	Aug. F	4 +
1964	July F	Sept S	1 +
1965	—		3 –
1966	July S	Aug. T	0
1967	July F	Aug. S	0
1968	July S	July S	3 –
1969	July F	Aug. S	1 +
1970	July S	Aug. F	1 –
1971	June T	Aug. T	0
1972	July S	July T	2 –
1973	July F	Aug. T	2 +
1974	July S	Aug. F	1 +
1975	July F	Aug. F	0
1976	July S	Aug. T	1 +
1977	June T	Aug. F	1 +
1978	July T	Aug. T	2 +
1979	July S	Aug. S	0
1980	Aug. S	Aug. S	2 –
1981	July F	Aug. S	1 –
1982	July T	Aug. F	2 –
1983	—		2 –
1984	July F	Aug. F	0
1985	July F	Aug. T	0
1986	June T	June T	2 –
1987	July F	Aug. T	0
1988	July F	Aug. S	0
1989	July S	July T	2 –
1990	July S	Aug. T	0
Mean	July S	Aug. S	

F–First ten day of the month (1–10); +:flood; –:drought. 1:light;
S–Second ten day of the month (11–20); 2: moderate; 3: strong;
T–Third ten day of the month (21–31); 4: extremely strong.

mer rainfall of the above 6 divisions for 1951–1989. In the process of its jumping movement over the Chinese continent from spring to summer, the rainbelt often dwells in South China, the Yangtze River basin, the Huaihe River basin, North China and Northeast China. Hence, Chinese meteorologists generally divide the whole of China into these six

regions to investigate the rainbelt. It is obvious that during the last 39 years there are 20 years in which drought (standardized rainfall anomaly < -1) or flood (standardized rainfall anomaly > 1) occurred in two of the 5 divisions of the eastern part, and there are 10 years in which the drought/flood occurred in 3 divisions. So it can be concluded that

Fig. 7. The standardized anomalies of winter temperature (left panel) and summer precipitation (right panel) in six divisions (SW-Southwest China, NE-northeast China, N-North China, H-Huaihe River Basin, Y-Yangtze River Basin, S-South China) of China during the period 1951–1989.

there is a drought/flood in the summer monsoon period every two years and a severe drought/flood every four years. The severe drought/flood years are 1954, 1956, 1958, 1963, 1965, 1968, 1973, 1977, 1980 and 1985. 1953, 1955, 1957, 1961, 1966, 1967, 1968, 1972, 1975, 1978, 1983, and 1988 are also drought/flood years. Their temporal distribution is rather even and there is an obvious periodicity. For the region of southeastern China there are 7 drought years and 7 flood years during the last 39 years *i.e.* one occurrence for every 3.5 years.

For the winter monsoon, the winter mean temperature were calculated and the standardized temperature anomalies for the six divisions are given in Table 6 and Fig. 7 (left panel). We can see that for the 5 divisions of the eastern part of China there are 10 years of severe cold winter (standardized temperature anomaly < -1 in at least 3 divisions) or very warm winter (standardized temperature anomaly > 1 in at least 3 divisions). There are

14 ordinary warm/cold years (standardized temperature anomaly > 1 or < -1 in two divisions) during the last 39 years. 1956, 1963, 1967, 1968 and 1983 are severe cold years in three divisions, while 1978, 1986 and 1989 are very warm years in three divisions. There are two other years in which a cold or warm winter occurred in more than 3 divisions. Meanwhile, it also can be seen that there has been no severe cold year in North China and Northeast China since 1980. The winter temperature in above two divisions is slightly higher than normal since 1973. On the other hand, Southwest China has been getting cooler since 1970 and has a cooling climatic trend.

In some regions of China the winter temperature has a significant correlation with the summer precipitation of the coming year. Figure 8 shows the correlation of the winter (December–February) temperature with the precipitation of the coming summer (June–August) over 200 stations during the period

Table 5. Seasonal pricipitation amount (mm) in six divisions of China.

Year	1		2		3		4		5		6	
	Apr.-June		June-Aug.		June-Aug.		June-Aug.		June-Aug.		June-Aug.	
	a	b	a	b	a	b	a	b	a	b	a	b
1951	710	.3	528	.3	473	.0	348	-.5	320	.1	621	.0
1952	673	.0	378	-.7	433	-.3	310	-.8	275	-.8	542	-.9
1953	697	.2	508	.2	318	-1.2	565	1.5	377	1.2	555	-.7
1954	715	.4	954	3.1	702	1.6	630	2.1	343	.6	865	2.6
1955	517	-1.3	563	.5	512	.3	391	-.1	239	-1.6	658	.4
1956	536	-1.1	528	.3	846	2.7	600	1.8	349	.7	626	.0
1957	864	1.6	549	.4	506	.2	347	-.5	423	2.2	582	-.4
1958	414	-2.1	307	-1.2	565	.6	362	-.4	259	-1.1	690	.7
1959	834	1.4	387	-.6	375	-.7	481	.7	343	.6	739	1.2
1960	559	-.9	397	-.6	476	.0	345	-.5	356	.8	716	1.0
1961	563	-.9	302	-1.2	369	-.8	496	.8	329	.3	792	1.8
1962	698	.2	613	.9	570	.7	382	-.2	330	.3	735	1.2
1963	363	-2.6	444	-.3	686	1.5	660	2.3	346	.6	504	-1.3
1964	597	-.6	378	-.7	370	-.8	541	1.3	357	.8	624	.0
1965	844	1.5	431	-.3	796	2.3	233	-1.5	339	.5	604	-.2
1966	786	1.0	314	-1.1	242	-1.7	431	.3	342	.5	715	1.0
1967	581	-.7	252	-1.5	382	-.7	398	.0	244	-1.5	591	-.4
1968	666	.0	272	-1.4	600	.9	157	-2.2	256	-1.2	610	-.1
1969	613	-.5	780	2.0	417	-.4	431	.3	404	1.8	542	-.9
1970	617	-.4	512	.2	452	-.2	371	-.3	256	-1.2	623	.0
1971	739	.6	401	-.5	580	.8	436	.3	306	-.2	580	-.5
1972	735	.6	423	-.4	577	.7	196	-1.9	253	-1.3	406	-2.3
1973	929	2.2	445	-.2	332	-1.1	528	1.1	362	.9	802	1.9
1974	672	.0	522	.3	470	-.1	369	-.3	291	-.5	658	.4
1975	827	1.3	627	1.0	460	-.1	378	-.2	275	-.8	524	-1.1
1976	605	-.5	393	-.6	339	-1.0	472	.6	273	-.9	541	-.9
1977	564	-.9	528	.3	315	-1.2	520	1.1	297	-.4	601	-.2
1978	742	.6	224	-1.7	302	-1.3	453	.5	272	-.9	548	-.8
1979	743	.6	443	-.3	530	.4	366	-.3	261	-1.1	573	-.5
1980	742	.6	844	2.4	568	.7	275	-1.1	256	-1.2	486	-1.5
1981	787	1.0	439	-.3	341	-1.0	359	-.4	323	.1	627	.0
1982	546	-1.0	567	.6	567	.7	395	-.1	269	-.9	481	-1.5
1983	666	.0	591	.7	561	.6	242	-1.4	327	.2	679	.6
1984	797	1.1	480	.0	501	.2	415	.1	388	1.5	738	1.2
1985	550	-1.0	343	-.9	287	-1.4	340	-.6	395	1.6	631	.1
1986	697	.2	548	.4	483	.0	311	-.8	388	1.5	591	-.4
1987	642	-.2	570	.6	505	.2	404	.0	329	.3	592	-.3
1988	630	-.3	410	-.5	290	-1.4	434	.3	292	-.5	673	.5
1989	628	-.3	609	.8	508	.2	306	-.9	272	-.9	660	.4
mean	669	119	482	151	477	137	402	110	316	49	624	93

1-South China; 2-Middle and Lower Reaches of Yangtze River;
3-Huaihe River Basin; 4-North China;
5-Northeast China; 6-Southwest China;
a-pricipitation amount; b-standard deviation;
mean-mean value and variance.

of 1951-1989. The contours in the map indicate the different significance levels corresponding to different correlation coefficients (For a significance level of 0.95, the correlation coefficient is 0.28, while for a 0.99 significance level it is 0.42). It can be seen that in the eastern part of Northwest China, Shaanxi and Gansu provinces there is a positive correlation region with a significance level of 99 %. Moreover, there is also a positive correlation area in South China with a maximum center of 99 % significance

Table 6. Winter temperature (Dec.–Feb.) (°C) in six divisions of China

Year	1		2		3		4		5		6	
	a	b	a	b	a	b	a	b	a	b	a	b
1951	14.5	.7	5.2	.6	2.2	.2	−1.0	.5	−14.9	.1	8.0	1.5
1952	13.8	.0	4.6	−.1	1.6	−.4	−2.7	−1.0	−16.8	−1.1	6.9	.3
1953	15.1	1.3	5.3	.6	2.5	.5	−.5	.9	−14.7	.3	7.7	1.2
1954	13.2	−.6	3.1	−1.6	.1	−1.7	−2.7	−1.0	−16.3	−.8	5.6	−1.3
1955	13.6	−.2	5.4	.7	2.3	.3	−1.5	.1	−15.1	.0	7.4	.8
1956	12.8	−1.0	3.0	−1.7	−.8	−2.6	−4.5	−2.5	−18.3	−2.0	5.7	−1.1
1957	13.7	−.1	4.9	.2	2.3	.3	−2.0	−.3	−15.5	−.3	6.7	.1
1958	14.2	.4	5.1	.4	2.6	.6	−.2	1.2	−12.9	1.4	6.6	−.1
1959	15.0	1.1	5.7	1.1	3.0	.9	−.7	.8	−14.1	.6	7.5	1.0
1960	13.3	−.5	4.7	.0	2.3	.3	−1.3	.2	−15.6	−.3	6.4	−.4
1961	14.0	.2	5.4	.7	2.4	.4	−1.3	.3	−13.9	.7	7.2	.6
1962	12.9	−.9	4.8	.1	1.8	−.1	−1.1	.4	−14.5	.4	6.8	.1
1963	13.2	−.6	3.7	−1.0	.7	−1.2	−3.4	−1.5	−15.5	−.2	5.7	−1.1
1964	14.9	1.1	5.9	1.3	3.1	1.1	−.8	.7	−16.2	−.7	7.3	.7
1965	15.2	1.4	5.4	.7	3.0	1.0	−1.3	.3	−15.9	−.5	7.6	1.0
1966	13.0	−.8	3.8	−.9	1.0	−.9	−2.7	−1.0	−15.6	−.3	5.6	−1.3
1967	11.6	−2.2	2.8	−1.9	−.1	−2.0	−4.1	−2.1	−16.6	−.9	5.1	−1.8
1968	14.6	.8	3.9	−.8	.5	−1.4	−3.1	−1.3	−16.8	−1.0	6.8	.2
1969	13.8	.0	4.6	−.1	2.1	.1	−2.3	−.6	−16.8	−1.1	6.9	.2
1970	13.7	−.1	4.7	.0	2.0	.0	−2.0	−.4	−15.2	−.1	6.8	.2
1971	13.5	−.3	3.6	−1.1	.8	−1.1	−3.4	−1.6	−14.8	.2	6.4	−.3
1972	15.0	1.2	5.5	.9	2.7	.7	−1.0	.5	−14.2	.6	7.6	1.1
1973	12.6	−1.2	4.0	−.7	1.4	−.5	−1.5	.1	−15.0	.1	6.3	−.4
1974	14.6	.8	5.3	.7	2.5	.5	−1.2	.3	−14.4	.4	6.6	−.1
1975	13.4	−.4	4.8	.2	2.3	.3	−1.1	.5	−13.3	1.1	6.5	−.2
1976	12.2	−1.6	3.2	−1.5	1.5	−.5	−2.3	−.6	−17.8	−1.7	4.7	−2.3
1977	14.6	.8	5.7	1.1	3.2	1.1	−.4	1.1	−16.0	−.6	7.3	.7
1978	15.7	1.9	6.3	1.7	3.7	1.6	−.6	.9	−13.9	.7	8.3	1.9
1979	14.0	.2	5.2	.6	2.6	.6	−1.5	.1	−14.8	.1	6.7	.0
1980	14.5	.7	4.4	−.3	1.9	−.1	−2.0	−.3	−15.7	−.4	7.3	.7
1981	14.0	.2	4.8	.1	2.2	.2	−.9	.6	−12.7	1.5	6.6	−.1
1982	12.4	−1.4	4.8	.1	2.5	.5	−.7	.8	−13.8	.8	5.8	−1.1
1983	11.7	−2.1	3.3	−1.4	1.4	−.5	−1.3	.3	−15.2	−.1	5.3	−1.6
1984	13.2	−.6	4.2	−.4	1.4	−.5	−2.0	−.3	−15.5	−.3	6.0	−.7
1985	13.4	−.4	4.2	−.4	1.7	−.2	−1.6	.0	−15.2	−.1	6.5	−.2
1986	16.0	2.2	6.0	1.4	3.2	1.2	−.3	1.1	−14.4	.4	8.6	2.2
1987	14.0	.1	4.7	.0	2.5	.5	−.4	1.1	−13.3	1.1	6.5	−.2
1988	13.8	.0	5.5	.9	2.9	.9	.6	1.9	−11.9	1.9	6.3	−.4
1989	14.8	1.0	7.2	2.6	4.6	2.5	.9	2.1	−9.3	3.6	7.8	1.3
mean	13.8	1.0	4.7	1.0	2.0	1.1	−1.6	1.2	−15.1	1.6	6.7	.9

1–South China; 2–Middle and Lower Reaches of Yangtze River;
3–Huaihe River Basin; 4–North China;
5–Northeast China; 6–Southwest China;
a–mean temperature; b–standard deviation;
mean–mean value and variance.

level located at the juncture of Fujian and Guangdong. On the other hand, in the region from Northeast China through the east part of North China to the north side of Yangtze River basin, there is a negative correlation. The maximum correlation centers are in the western part of Northeast China.

Shandong and Sichuan province have a 95 % significance level. The correlation between winter temperature and summer rainfall implies that there is a certain relationship between the winter monsoon and the summer monsoon. This is worthy of further investigation.

Fig. 8. The correlation distribution between the winter (Dec.-Feb.) temperature and precipitation of the next summer (June-Aug.) over 200 stations of China during the period 1951-1989.

4. The oscillation characteristics of the inter-annual variation of East Asian Monsoon

According to the studies of Chen (1984) and Tao and Chen (1987), the East Asian Monsoon system is different from the Indian Monsoon system and has an interaction with the Indian Monsoon. The members of the East Asian Monsoon mainly include the surface Australian High, the cross-equatorial currents in the South China Sea and the western Pacific, the monsoon trough in the South China Sea, the ITCZ in the western Pacific, the subtropical High in the western Pacific, the Mei-Yu front, and the troughs in extratropical westerly. Having studied the circulation pattern for drought/flood years of the middle and lower Yangtze River basin, Tao *et al.* (1988) pointed out that the important reason for the drought or flood there is the abnormal position of the main elements of the East Asian Monsoon circulation systems. In flood years, the surface Australian cold high, the ITCZ at 850 hPa and subtropical high at 500 hPa are all located slightly southward of their normal position, while in drought years all the circulation systems move northward from their normal position. Hence, it is very important to investigate the possible periodicity in the rainfall data and to investigate the interannual variation of the associated monsoon circulation systems.

One major characteristic of the East Asian Monsoon circulation systems is their oscillations with an interannual time scale. Beijing has long-term precipitation records beginning from 1724. According to Lu (1984) who improved the methods of Jenkins and Watts (1968), we made a spectra analysis of precipitation data for 261 years (1724-1985) at Beijing. The result shows that there are 140 year, 70

year, 3.7 year and 2.5 year periods with a significance level exceeding 95 % for all the periods above. Chen L. and Chen D. (1990) has done the power spectra analysis for 48 Chinese stations using the monthly mean temperature and precipitation data during 1933-1989, and found that there are obvious quasi-biennial and quasi-triennial oscillation periods (referred to as QBO and QTO, respectively). Figure 9 gives the power spectrum of precipitation (Fig. 9A) and temperature (Fig. 9B) for some Chinese stations. It is clear that for some stations the QBO is their principal oscillation and for other stations the QTO is dominant. There are still some stations where both the QBO and the QTO are important and have a 95 % significance level. At most of the stations, the density of the power spectrum of the QBO of temperature can pass the 95 % significance test. These stations mainly lie in North China, the Yangtze River basin, South China, Northeast China and the east part of Northwest China. There are 15 stations (about 1/3 of the total of 48) in which the QTO of temperature can pass the 95 % level significance test. These stations are mainly located in Northeast China, Northwest China and South China. For precipitation, the QBO at most of the stations is significant at the 95 % level and the QTO at 23 stations can pass the same significance test. The Maximum amplitudes of the QBO mainly occur in Northeast China, North China, the eastern part of Northwest China, the middle and lower Yangtze River basin and Guangdong province, *etc.* and the QTO is active in North China, Northeast China and the middle and lower Yangtze River basin.

The QBO and QTO of temperature and precipitation have been analyzed by using the complex

Fig. 9. The power spectrum of precipitation (A) and temperature (B) for some Chinese stations. A: The stations are Tianjin (a), Huaian (b), Guiyang (c), Changchun (d), Guangzhou (e) and Shanghai (f), respectively. B: The stations are Lanzhou (a), Xian (b), Youyang (c), Jinan (d), Linxia (e), Guiyang (f), respectively. The unit of abscissa is month.

EOF (Huang, 1988). First we decomposed the real time series of the QBO or QTO. Then we composed a reconstructed field using certain principal components. Fig. 10 shows a composed precipitation field for the first principal component with an explicable percentage of 18.0%. The data used in this analysis are monthly mean rainfall data at 160 stations during 1951–1987. When the phase is equal to zero, there are negative areas in Northeast China, the Huaihe River basin and South China, and positive areas in North China and the middle and lower Yangtze River basin. The northern zero line is located in the northern part of Northeast China. At the $\pi/4$ phases, the zero line moves southward to the middle part of Northeast China. The negative areas over Northeast China expand southward, while

the negative area over South China spread northward. At a phase of $\pi/2$, the former reaches North China and the latter reaches the middle and lower Yangtze River basin; between these two negative regions there is a positive area. The northern zero line moves to North China. When the phase equals $3/4\pi$, the northern zero line arrives at the Yangtze River basin. The two negative centers are formed in North China and the middle and lower Yangtze River basin respectively and these two centers spread westward. Meanwhile, a new positive area forms over Northeast China and South China. In this process, the southern zero line moves northward from South China with phase 0 to Yangtze River basin with phase $3/4\pi$. Hence, there are two sources of QBO of precipitation over China, one comes from the high latitude

Fig. 10. The composite temporal sequences of amplitude of the first main component of CEOF for QBO (A) and QTO (B) of rainfall in China. The phases of Fig. 10A (a) to Fig. 10A (d) and Fig. 10B (a) to Fig. 10B (d) indicate half cycles of the QBO and the QTO, respectively.

and propagates to North China through Northeast China. The other comes from the South China Sea and western Pacific and propagates through South China to the middle and lower Yangtze River basin. During the northward propagation, the QBO also spreads westward. The characteristics of propagation of the QTO of precipitation are basically similar to those of the QBO. Namely, the QTO propagates from Northeast China to North China while the southern one spreads westward and propagates from South China to the middle and lower Yangtze River basin. But the propagation of the QTO is not as clear as that of the QBO.

The propagation characteristics of a recomposed field of temperature for the QBO and QTO are similar to those of precipitation (figures omitted). But the oscillation takes a positive sign over whole country at some phases and takes a negative sign at other phases. During the transition between positive and negative sign, the positive or negative centers show an apparent movement from Northeast China to North China and from South China to the middle and lower Yangtze River basin respectively. This indicates that for the short-range oscillation, the oscillations have the same sign over the whole of China. The above-mentioned features can also be seen in Table 6 and Fig. 8.

The members of the East Asian Monsoon system all have an obvious interannual oscillation. Guo (1988a) analyzed the power spectrum of the monsoon index series during 1871–1980, and found that the winter monsoon index has 2.06–2.1 year oscillations at a 95 % significance level. The summer monsoon index (SMI) has oscillation of 35 year period at the 95 % significance level and of 2.5-, 3.5- and 7-year period at a significance level close to 95 %. This indicates that generally the East Asian Monsoon system has a QBO, QTO and other oscillations of longer periods.

The subtropical high is a very important member of the East Asian Monsoon system. Wang and Zhao (1987) took the mean anomalies of the 500 hPa height at 9 points in the area between 15°–25°N and 105°–125°E to represent the strength of the subtropical high over the South China sea, and found that it has oscillations with a 2.2-, 3.5-, 5- and 11.3-year period at the 95 % significance level. The Australian High at surface level during boreal summer is another important member of the East Asian Monsoon system. Wang and Zhao (1987) used the monthly SLP averaged in the region of 20°–30°S, 120°–150°E in July to express the strength of this high and found, through spectral analysis, that there are also oscillations of period 3.6 years and 15.7 years.

In conclusion, the strength and the members of the East Asian Monsoon have apparent QBO, QTO and quasi-5–7 years oscillations (QFO). These short-

range climatic oscillations have characteristic propagation and are influenced by the systems both from high latitudes and from the tropical oceans. Hence, to understand the mechanism of the interannual oscillations of the East Asia Monsoon, we must investigate the interannual variation of the climatic parameters in other regions of the world.

Chen Y. and Chen L. (1991) have analyzed the monthly sea ice area data in the north polar region from 1953 to 1989, which were supplied by NMC of USA. It is found that by using spectral analysis, there are the QBO and QTO. Neither pass the 95 % level significance test, though the QBO's significance level is close to the 95 % level. It is also found for some particular months, *e.g.* May, June, September and November, that the QBO can pass the 95 % level significance tests. These facts show that the oscillation is obvious during the melting and freezing seasons.

5. The features of oscillations and their interactions in air-sea-land system

The interannual oscillations also exist in the Pacific SST. By using COADS data, we analyzed the power spectrum of the SST in the equatorial regions (from 1950 to 1987) and found that there exist QBO, QTO and QFO both in the western Pacific and in the eastern Pacific. Yan *et al.* (1989) pointed out that for the SST at the equator of the eastern Pacific, the sum of the QBO and QTO components approximately approaches the actual value, and several SST peaks in the El Niño years can be displayed clearly. But, these two components can not show all the SST valley points in the La Niña years unless the QFO component is added. The sum of the three components of QBO, QTO and QFO accounts for 75 % of the variance of the actual value.

Figures 11a and 11b are the time-longitude cross sections of the QBO and QTO of SST anomalies (SSTA) along the equator. It can be found that in the period of 1950–1979, the QBO and QTO of SSTA propagate eastward at the equator of the western Pacific and westward at the equator of the eastern Pacific. The oscillations in the western Pacific are weaker than those in the eastern Pacific. For both the QBO and the QTO, there is an oscillation center in the central and eastern Pacific, respectively. We also find that an oscillation with a longer period exists in the QBO and the QTO of SSTA. The QBO is stronger in the periods 1950–1954, 1962–1967, and 1971–1976 and is weaker in the periods 1955–1961, 1968–1970. Namely, it has a cycle of 11 years. The QTO is stronger in the periods of 1966–1975 and 1952–1958. From the time-longitude sections of the QBO and QTO of the upper zonal wind along the equator which were made by Yasunari (1985), we are surprised to find that the periods of strong and weak QBO and QTO of zonal wind and SSTA are

Fig. 11. The time-longitude cross sections of the QBO (a) and QTO (b) of SSTA along the equator. The interval of contours is 0.3°C and the negative values are shaded.

very similar. This indicates that the interannual oscillations in the air-sea system are well coupled.

Yasunari (1987) investigated the QTO of the southern oscillation index (SOI) (named SO in his paper) and the evolution of atmospheric and oceanic circulations. From COADS data, we simplified the data sets of sea surface temperature (SST), zonal wind (U) at the sea surface and sea level pressure (SLP) to a $4^{\circ}\times 4^{\circ}$ mesh, analyzed the QBO and QTO using the method of CEOF (the complex empirical orthogonal function). The composite fields for the first main component at different phases have been made. Using SST in the eastern equatorial Pacific as the reference for time series, the lag-correlation with the SST, U and SLP in the whole field was also analyzed. Figure 12 shows the patterns of lag-correlation between the averaged SST in the eastern equatorial Pacific and the QBO component of SST, U and SLP. The data sets are from 1950 to 1979. The correlative coefficient with a confidence of 95 % is 0.12. In Fig. 12, The SSTA in the eastern equatorial Pacific lagged in negative months and led in positive months. It is found that at zero month (*i.e.* the simultaneous correlation) the negative correlative center is located in the western equatorial Pacific while the positive center is situated in the equatorial Indian Ocean. This shows a seesaw pattern of SSTA. At 12 month (the first half cycle) accompanying cold water in the eastern equatorial Pacific and warm water in the western equatorial Pacific, a warm water center appears in

Fig. 12. The lag-correlation of QBO with SSTA in the eastern equatorial Pacific. (a) SST (0.1°C), (b) U (0.1 m/s) and (c) SLP (0.1 hPa). Negative values are shaded. The SSTA in the eastern equatorial Pacific is behind in negative months.

the western subtropics of the southern Pacific. It travels eastward to the coast of South America, then moves northward to the equator and finally propagates westward to form a warm water center in the equatorial region of the central Pacific at 3 months. Meanwhile, the warm water center in the western equatorial Pacific at 12 months extends eastward and establishes a positive center in the equatorial region of central Pacific at 3 months. The source of warm water in the western equatorial Pacific may come from the subtropics of the northern Pacific from which it propagates to the western Pacific to affect the western equatorial Pacific. From the above discussion, we can see that for the QBO the activity of SSTA at the equator has a close relationship with the ocean activity of both subtropics in the southern and the northern Pacific. The equatorial zonal wind center of the QBO moves eastward regularly from the Indian Ocean; it is enhanced when it arrives at the western equatorial Pacific and then continues to shift eastward to the central equatorial Pacific. According to the time resolution in the figures, in the western equatorial Pacific the phase of the warm water center is the same as that of the center of the west wind anomaly. At 0 month, the strongest warm water center locates in the eastern equatorial Pacific, the west wind anomaly center only in the western equatorial Pacific, and the east part of the eastern equatorial Pacific is still dominated by an east wind anomaly center.

The correlation of SLP anomalies demonstrates that they travel eastward with the same anomalous signs in both hemispheres. It indicates that during the occurrence of the southern oscillation (SO) the SLP distribution in the east-west direction in the northern Pacific has the same variation as the southern Pacific. Chen (1984) studied this phenomenon and pointed out that, while paying attention to SO, we should also give an eye to the northern oscillation (NO). He also indicated that only when the SO and NO have the same variation signs simultaneously does strong El Niño event occur. The conclusion is the same as that shown in the SLP of Fig. 12.

Figure 13 is the composite field of each phase for the first component of the QBO of SSTA with the method of CEOF. It is calculated from COADS which has a time series from 1950–1987 and a grid of $4^\circ \times 4^\circ$. It is a half cycle from phase one to phase six. The interval between two adjacent phases is 2 months. In phase 6 there is a warm water center in the eastern equatorial Pacific and the central equatorial Pacific. The former originates from the subtropics of the southern Pacific in phase one and travels anticlockwise to the eastern equatorial Pacific. The latter locates at the western equatorial Pacific in phase 1, which is affected by the warm water in the subtropics of the northern Pacific. The characteristics of the QBO are just the same as those

in Fig. 12. The evolutionary features of U and SLP of the QBO are also consistent with those in Fig. 12, though the analysis methods are somewhat different. Figure 14 is the synthesized scalar speed $\bar{v}$ QBO field of each phase corresponding to Fig. 13. The features of $\bar{v}$ in Fig. 14 are very similar to those in Fig. 12. The characteristics of the QTO for SST, U and SLP are very similar to those of the QBO.

There are many works studying the influence of the SST in different areas on the summer rainfall in China. The group for long-range weather prediction of the Institute of Atmospheric Physics of Academia Sinica (1978) found that the temperature of the Kuroshio current in winter (December–February) has a high positive correlation with the summer rainfall in the middle and lower reaches of Yangtze River and that of Hebei province (the coefficient is above 0.53), the southward deviation of the axis of the Kuroshio current is conducive to precipitation in the middle and lower reaches of Yangtze River. Tao *et al.* (1988) discovered that the correlation of the standard deviation between SSTA in the warm pool of the western Pacific in early summer and anomalous precipitation of the middle and lower reaches of Yangtze River in June and July has a large negative value. For instance, the coefficient from 1954–1984 is about -0.45 . This shows that if the SST of the warm pool is larger in early summer, the precipitation in the middle and lower reaches of Yangtze River is less in summer. Luo *et al.* (1987) pointed out that the large (small) amount of rainfall of the middle and lower reaches of Yangtze River coincides with the warm (cold) SSTA in the Indian Ocean and the Bay of Bengal in the same period. In other words, they have a positive correlation. The contrast of the results between Tao *et al.* (1988) and Luo *et al.* (1988) displays a negative correlation of the SST in the warm pool of the western Pacific and in the equatorial Indian Ocean.

Chinese meteorologists have noticed the possible influences of SST in the eastern equatorial Pacific with El Niño on the summer precipitation in China. Chen *et al.* (1990) calculated the correlation between SST in the eastern equatorial Pacific and the rainfall in different regions of China. No obvious relationship has been found, which is in agreement with the result that there is no evident correlation existing between the pressure at Darwin and pressure over China. From Table 5 and Fig. 8 which shows the standardized anomalies of the summer rainfall in each region of China, we get the positive and negative anomalous numbers of summer precipitation in corresponding regions in an El Niño year and the next year from 1951–1989 (Table 7). In ten summers of an El Niño year from 1951–1989 (1951, 1953, 1957, 1963, 1965, 1969, 1972, 1976, 1982 and 1986) North China and Southwest China were drier (occurred seven times and eight times, respectively)

Fig. 13. The composite phase field reconstructed from the corresponding field of the first eigenvector of SSTA's QBO. The negative regions are shaded. Interval of contours is 0.2°C . Unit: 0.1°C . (a) phase 1 (0π); (b) phase 2 ($\pi/6$); (c) phase 3 ($\pi/3$); (d) phase 4 ($\pi/2$); (e) phase 5 ($\pi/3$); (f) phase 6 ($\pi/6$).

Fig. 14 As in Fig. 13, except for scalar speed $\bar{V}$. Units: 0.01 m/s . Interval: 0.02 m/s .

and Northeast China was wetter (happened seven times). The probabilities are only 0.7 and 0.8 during the forty years. But there are no apparent correlations in the middle and lower reaches of Yangtze River and South China in Table.7, which is consistent with the fact that in an El Niño year India is

drier in summer, and the summer precipitation of India has a positive correlation with that of North China and Sichuan province (Fig. 8). During an El Niño year, the subtropical high of the western Pacific displaces to the east and south; the ITCZ also displaces to the south and the occurrence of typhoons is less (Li, 1990) because, when the eastern equatorial Pacific is warm, the western equatorial Pacific tends to be cold and the strong convection area moves eastward, thus the subtropical high is displaced to the south. However, this means that El Niño affects the summer rainfall first in South China and the middle and lower reaches of the Yangtze River, but not in North China. There are several interpretations of this observation:

(1) During an El Niño event, there may be two warm water centers in the eastern equatorial Pacific. One is near the equator of the central Pacific and the other is located at the equator of the eastern Pacific. When the former center is stronger, then there is less rainfall in the middle and lower reaches of the Yangtze River (Fu, 1988).

(2) The commencement of an El Niño event can be in spring and summer or in autumn and winter. These two kinds of El Niño have different impacts on the summer rainfall of China. If an El Niño begins in the fall of the previous year, the precipitation will be more in the middle and lower reaches of Yangtze River in the current summer. It is less if an El Niño begins in the current spring and summer

Table 7. The number of positive and negative summer precipitation anomalies in five divisions of China in El Nino and the next year during the period of 1951–1989.

	1		2		3		4		5	
	+	–	+	–	+	–	+	–	+	–
El Nino Year	6	4	6	4	3	7	7	3	2	8
Next Year	4	6	5	5	5	5	6	4	6	4

1–South China; 2–Middle and Lower Reaches of Yangtze River;
3–North China; 4–Northeast China; 5–Southwest China.

(Zhao, 1990). Since the result is obtained from only ten cases since 1950 and they have been divided into two groups, more examples are needed to verify this conclusion.

(3) Though the anomaly of the SST in the eastern equatorial Pacific has no good correlation with precipitation over China, their QBO and QTO components correlate well (Chen *et al.*, 1990). That the two components superpose when in phase and compensate when opposite in phase results in an irregular variation of the SST in the eastern equatorial Pacific and of rainfall. Figure 15 is the lag correlation patterns of the QBO and QTO between an anomalous SST in the eastern equatorial Pacific and the precipitation at 160 stations in China. It demonstrates an apparent relation in the middle and lower reaches of the Yangtze River. Hence the relation between ENSO and the summer monsoon in China can be exposed from different time-scale interannual oscillations.

The another component of the climate system which may influence the interannual variation is the polar ice area. Figure 16 is the yearly curve of the sea ice area in the Arctic, in which “E” represents an El Nino year. It can be found that El Niño events occur in the years with peak values of ice area at the north pole, though not all years of peak values have El Niño events. This shows that there is a relationship between the occurrence of an El Niño and the area of sea ice in the Arctic. It needs further investigation in our future research.

There have also some studies of the relationship between the snow cover of the Tibetan Plateau in winter and spring and the summer precipitation over China. Chen and Yan(1978) revealed that with a large (less) amount of snow cover over the Plateau in winter and spring, the South Asian High at 100 hPa is weak (strong) in spring of the same year, and rainfall in June in South China is greater (small). Li *et al.* (1991) also obtained the same conclusion using a two-dimensional five-layer atmospheric model which contains an ocean mixed-layer as a boundary condition. However, some other researches have obtained a similar result using the snow cover in Europe and Asia (Mooley and Shukla, 1987). It needs to be verified whether there are strong correlations

between the snow cover of Europe and Asia and that of the Tibetan Plateau, and which of those regions is more effective in influencing the monsoon circulation.

6. Summary and concluding remarks on interannual variation of the East Asian Monsoon

In this section, we will summarize the factors affecting the interannual fluctuation of the East Asian Monsoon.

1. The effect of land-sea thermal contrast on a decade time-scale: It is observed that in the last forty years, the temperature increased on the continent of northern China in summer and winter. The SST decreased in the Pacific Ocean in the same latitudes. The temperature showed a cooling trend in southern China but showed a warming in the adjacent Pacific Ocean south of 35°N. Meanwhile, precipitation is reduced on the continent in summer, especially in the north. These facts show that the summer monsoon and the winter monsoon weakened on the continent. Therefore, the thermal contrast between the continent and the adjacent Pacific Ocean seems to be the primary factor in decade-scale trend of monsoon activity. The interannual variation of the monsoon is a fluctuation superimposed on this long-term trend of climatic change. Hence, there were no continuous flood years in northern China in recent years.

2. The effect of Ocean: The effective regions are the Kuroshio current, the warm pool of the western Pacific, and SST of the eastern Pacific, South China Sea and Indian Ocean. From Figs. 11–13 it is obvious that SST in the western and eastern equatorial Pacific and the equatorial Indian ocean have an intrinsic relationship and they can not be regarded as independent systems. The variation of the East Asian Monsoon is influenced by these change of ocean temperature.

3. The effect of ice and snow: Both the sea ice area in the Arctic and the snow cover on the Tibetan Plateau influence the climate system. There have been considerable investigations into the influence of the polar ice on drought and floods in China. It has been discovered that, when the sea ice area from

Fig. 15. The distribution of the lag-correlation coefficient of the QBO and QTO between SST in the eastern equatorial Pacific and rainfall in China. Negative months mean that SST in the eastern equatorial Pacific is behind.

Fig. 16. The curve of sea ice area in the Arctic during the period 1953–1989. E represents an El Niño year. The curves of monthly and 12-month moving mean are indicated by thin and thick with units of 10^{25} km^2 .

the Bering sea to Alaska is large (small) in spring, the subtropical high of the western Pacific is weak (strong) in the same season. Furthermore, the temperature in the eastern China is lower (higher) and precipitation is more (less).

Eventually, the mechanism of interannual variation of the East Asian Monsoon should be analyzed mainly from the interaction of the air-sea-land system, except for external forcing such as solar influences and volcanoes. A study in this field is under development presently. It is anticipated that the main research will focus on the SST of the warm pool in the western Pacific and the snow cover of Eurasia. These factors probably influence the interannual variations of East Asian Monsoon more directly.

In conclusion, the interannual variations of the East Asian Monsoon are superposed on the decade time-scale climatic trend. There apparently exist several different time-scales of interannual oscillations. They are the quasi-biennial oscillation (2 years), quasi triennial oscillation (3.5 years) and

quasi-5–7 year oscillation (5–7 years). These oscillations also exist in the air-sea-ice system. For each time-scale of the oscillation, the correlations are evident among the East Asian Monsoon, Pacific, and the Arctic ice area. The climatic factors influencing the interannual oscillations must be investigated to comprehend the East Asian Monsoon fully.

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東アジアモンスーンの年々変動の特徴

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中国大陸本土および周辺海域での 1951~90 年の期間における気象観測データを使用して、東アジアモンスーンの気候トレンドと年々変動の特徴を調べた。

その結果、過去 40 年間、中でも 1980 年代には、中国では全般的な乾燥化傾向と北緯 35 度以南の地域での寒冷化傾向がみられることが明らかになった。この結果は、中国では冬・夏のモンスーンがともに弱まっているためと解釈できる。楊子江中下流域での梅雨や他地域での降水量は、この気候トレンドの影響を受けている。

従来インドモンスーンは、中国での夏の降水量と関係があると信じられてきた。しかし、我々の研究の結果、インドでの夏の降水量は、華北西部地域での夏の降水量との間でのみ有意な相関が認められ、他地域との間での相関関係は弱いことがわかった。

中国の夏のモンスーンの年々変動は、東アジアの夏のモンスーン循環を構成する要素の変動によって起こっている。大気-海洋ならびに大気-陸面相互作用に関しても考察を行なった。東部赤道太平洋の海水温 (SST) や、北極域の海水面積と中国の冬・夏のモンスーンの強さとの関係には、有意な相関は得られなかったものの、それらは同様な年々変動を示し、同じ周期での年々変動成分間の相関は、大変強いことが明らかになった。