Chapter 9

Effects on Summer Monsoon and Rainfall Change Over China Due to Eurasian Snow Cover and Ocean Thermal Conditions

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Additional information is available at the end of the chapter

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1. Introduction

Climate over China is mainly governed by the East Asian monsoon, which mainly arises from the seasonal variation of the land-sea thermal contrast. Climatologically, the East Asian continent is warmer than surrounding oceans in summer (June-August) and colder in winter (December-February), leading to the occurrence of the East Asian summer monsoon and winter monsoon, respectively. During summer monsoon period, warm and moist air over tropical oceans is transported to the East Asian region [1], forming plenty summer monsoon rainfall. Since the rainfall in China is mainly concentrated in summer, the change and variability of the summer monsoon rainfall have been the research focus because of their important effects on economy, society and human life.

In the past several decades, one predominant feature of global warming is that the increasing of air temperature over land is stronger than that over sea [2], implying strengthening of the East Asian summer monsoon. However, many researches have shown that the summer climate in eastern China is characterized by multi-timescale variations [3-5]. The long-term variation is featured by the distinguished inter-decadal variability. During the period of 1880-2002, over eastern China there were no long-term trends for both annual and seasonal mean rainfalls, exhibiting dry and wet cycles in a noticeable inter-decadal timescale [6]. There are significant differences in the rainfall change among the areas of northern China, the Yangtze-Huai River Valley and southern China. In eastern China the above-normal rainfall zone moves successively from northern to southern China in the past several decades. More summer rainfall appeared over northern China in 1960s and 1970s; the Yangtze-Huai River valleys experienced a wet period from the late 1970s to the late 1980s; from the late 1980s or early 1990s more rainfall zone shifted to southern China [7-9].

The above-normal rainfall over eastern China is associated with the warm and moist southerlies. In 1960s and 1970s the southerlies reached northern China and more rainfall appeared there [10]. The more rainfall over the Yangtze-Huai River valleys from the late 1970s to the early 1990s and that over southern China afterwards were also related with the southerlies appeared over these regions [8,9].

Under the background of the global climate warming, the observed summer rainfall change over the East Asian monsoon region in eastern China exhibits unique features. Many studies have revealed that the Eurasian snow cover and ocean thermal conditions have important impacts on the summer monsoon rainfall change over eastern China. In this Chapter, possible causes of the summer monsoon rainfall changes over eastern China will be discussed based on relevant studies. This chapter is organized as follows. The observed features of the summer monsoon rainfall change over eastern China in the past half century will be given in Section 2. Section 3 describes the impact of the spring (March-May) Eurasian snow cover on the summer monsoon rainfall change over China. The role played by the changes of the thermal condition in the Atlantic Ocean and the sea ice in the Arctic Ocean will be discussed in Section 4. Section 5 focuses on the influence of the changes of sea surface temperature (SST) in the tropical Pacific and Indian Oceans. A conclusion is given in Section 6.

2. Observed summer monsoon rainfall change over eastern China since 1950s

The geographic location of the East Asian monsoon region over East Asia is usually taken to be in the area to the east of 100°E [11-13]. In order to check the long-term variation of the summer monsoon rainfall over China, we average the observed summer (June-August) rainfall to the east of 100°E in China in each year from 1958 to 2011. The rainfall data is from the monthly rainfall dataset observed at 160 stations in China and the station locations can be found in [14].

2.1. Changes of summer rainfall in eastern China and East Asian summer monsoon

The variation of the summer monsoon rainfall averaged to the east of 100°E over eastern China with time is shown in Figure 1. From Figure 1 it can be seen that, besides the inter-annual variability, the inter-decadal variability is a predominant feature for the long-term variation. No clear trend can be found for the summer monsoon rainfall from 1958 to 2011. A weak decline was observed from 1950s to the late 1980s, when the rainfall began to increase until the middle 1990s. From then on the rainfall declined again to the early 2000s and kept stable afterwards.

Since the summer rainfall in eastern China is greatly influenced by the East Asian summer monsoon, to see the long-term variation of the East Asian summer monsoon, we calculated the western North Pacific-East Asian summer monsoon (WNP-EASM) index [15] in 1958-2011 by

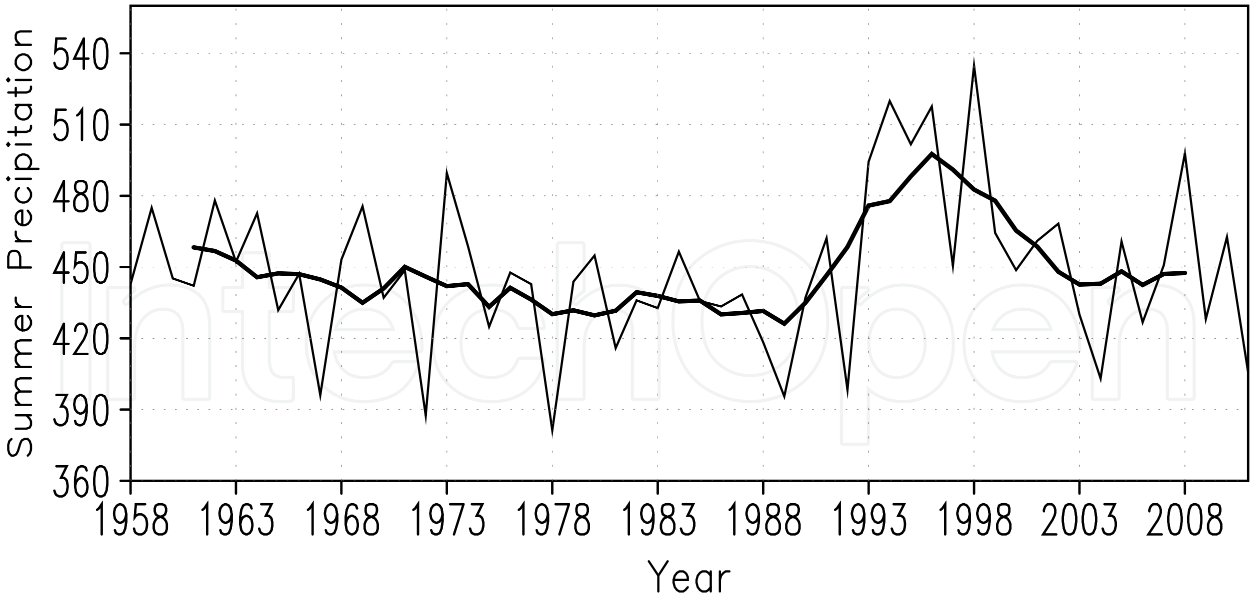


Figure 1. Observed summer (June-August) rainfall (thin line) and its 7-year running mean (Thick line) averaged in the area to the east of 100°E in China (Unit: mm).

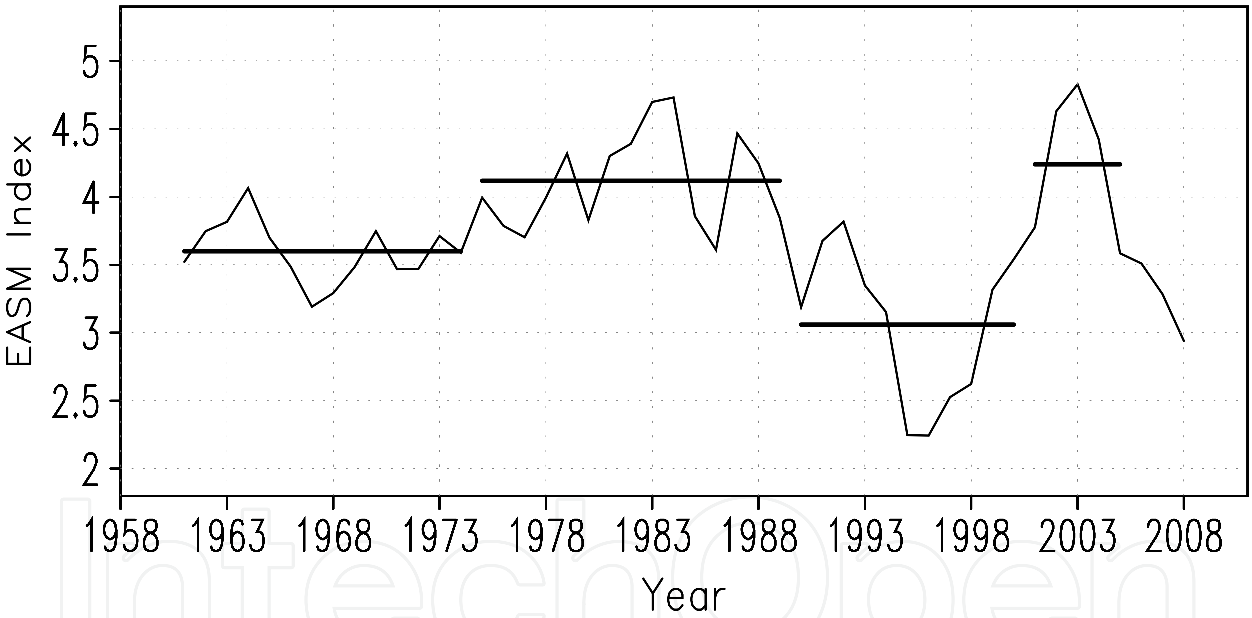


Figure 2. 7-year running mean of the western North Pacific-East Asian summer (June-August) monsoon (WNP-EASM) index (Unit: m/s). The thick lines indicate averaged WNP-EASM indexes in 1958-1974, 1975-1989, 1990-2000 and 2001-2005, respectively.

using the reanalysis data from National Center for Environmental Prediction (NCEP) and the National Center for Atmospheric Research (NCAR) [16]. Figure 2 shows the 7-year running mean of the WNP-EASM index. Same as the variation of the East Asian summer monsoon rainfall, there is also no clear trend for the WNP-EASM index, and strong inter-decadal variability appears. From Figure 2 we can see that climate shifts for the East Asian summer monsoon appeared at middle 1970s, late 1980s and early 2000s, respectively. In 1958-1974 the East Asian summer monsoon was weak. It turned to be strong in the period from 1975 to 1989, and became weak again in 1990-2000. In the period of 2001-2005 the East Asian summer monsoon strengthened and weakened afterwards. In addition to the inter-decadal variation, Figure 2 also shows that the larger amplitudes and shorter periods appeared since the late 1980s. Such feature can also be observed in the changes of the rainfall as shown in Figure 1. It implies a larger variability and more frequent variation of the East Asian summer monsoon have occurred since then.

2.2. Changes of abnormal rainfall pattern in eastern China

In order to check changes of abnormal pattern in summer monsoon rainfall over eastern China in association with the inter-decadal variability of the East Asian summer monsoon, in Figure 3 we show the differences of the averaged summer rainfall over eastern China between 1975-1989 and 1958-1974, between 1990-2000 and 1975-1989, and between 2001-2008 and 1990-2000, respectively. Corresponding to the climate shift in the middle 1970s (see Figure 2), the summer rainfall in 1975-1989 increased in the area over the middle and lower reaches of the Yangtze River valley and decreased over southern and northern China compared to the summer rainfall in 1958-1974 (Figure 3a). This shift of the summer rainfall anomalies caused droughts in the northern China and floods around Yangtze River valley [17,18]. After the weakening of the East Asian summer monsoon in the end of 1980s, more rainfall appeared in southern China in the period of 1990-2000 relative to 1975-1989 (Figure 3b), which was in accordance with the findings in [7]. Compared to the rainfall in 1990-2000, in 2001-2008 the East Asian monsoon strengthened and the rainfall decreased around Yangtze River valley, and increased to the south of about 25°N in the south of China and between about 31°N-36°N to the north of Yangtze River valley (Figure 3c).

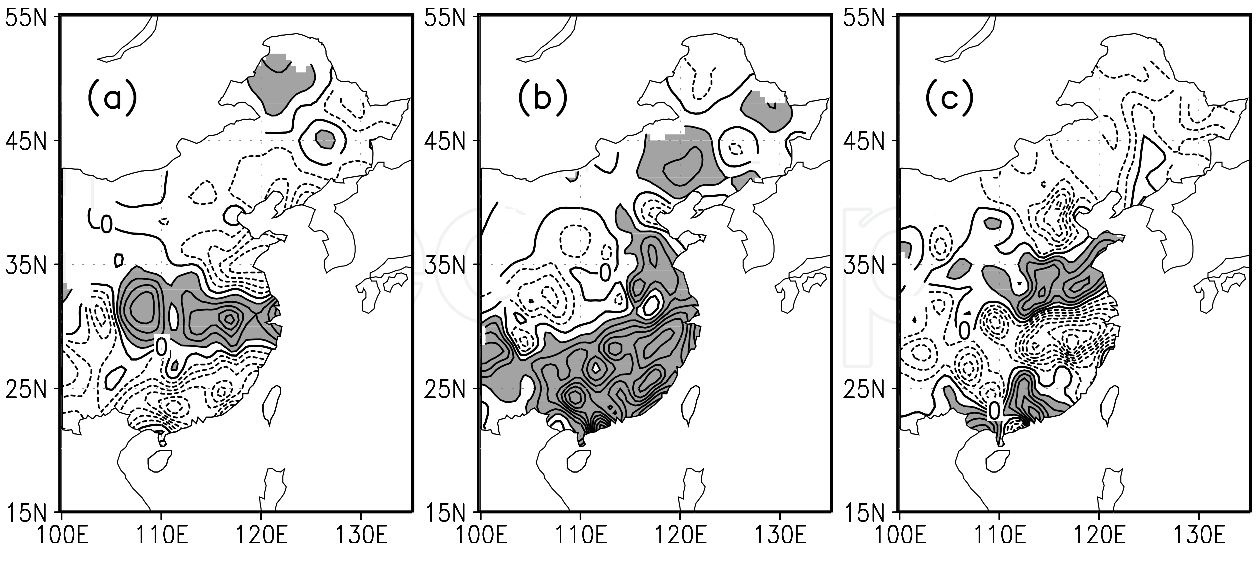


Figure 3. Averaged summer rainfall differences between (a) 1975-1989 and 1958-1974, (b) 1990-2000 and 1975-1989, and (c) 2001-2008 and 1990-2000. The dotted and real lines represent the negative and positive, respectively. The line interval is 30 mm. The thick line is the 0 isoline. The difference larger than 30 mm is shaded.

Here we can see that in association with the global climate warming in the recent half decade, although no clear trends for both the summer rainfall over eastern China and the East Asian summer monsoon, the abnormal rainfall pattern changes obviously. Compared to 1958-1974, more rainfall appeared around the middle and lower reaches of the Yangtze River valley in 1975-1989; the increased rainfall zone moved southward from 1975-1989 to 1990-2000 to the south of the Yangtze River valley in southern China; relative to 1990-2000, the rainfall increased in both the further southern area and around the area to the north of Yangtze River valley in 2001-2008. In following sections the possible physical reasons contributing to the changes of the abnormal rainfall pattern will be discussed.

3. Impacts of the spring (March-May) Eurasian snow cover

Snow cover is largely controlled by atmospheric circulation, while widespread snow cover affects local and large-scale atmospheric circulation and hydrological processes through changing water and energy flux. In addition to its high albedo and low thermal conductivity, snow cover acts as a heat sink through sublimation and melting processes. Snow cover cools the overlying atmosphere and warms the underlying ground [19].

3.1. Tibetan Plateau snow cover

The association between the Tibetan Plateau snow cover variability and Chinese rainfall has been extensively explored. The excessive snow cover in the Tibetan Plateau during winter/spring is associated with above-normal May-June rainfall in southern China [20,21]. Summer rainfall in central China along the mid- and low-reaches of the Yangtze River has a positive correlation, whereas that in northern and southern China has a negative correlation with the Tibetan Plateau snow depth in the preceding winter/spring [22-24].

Chinese summer rainfalls in the Yangtze-Huaihe River valleys and in the upper-lower reaches of the Yangtze River showed a remarkable transition from drought period to rainy period in the end of 1970’s, in good correspondence with the decadal transition of the winter snow cover on the Tibetan Plateau [22]. It is further demonstrated that there is a close relationship among the inter-decadal increase of snow depth over the Tibetan Plateau during March–April, a wetter summer over the Yangtze River valley, and a dryer one in the southeast coast of China and the Indochina peninsula; the excessive snowmelt and increased surface moisture supply over the Tibetan Plateau lead to a wetter summer in the vicinity of the Yangtze River valley [24]. In fact, a moderate positive correlation between the Tibetan Plateau snow cover and spring rainfall in southern China contains El Niño-South Oscillation (ENSO) effects, and ENSO has larger impacts than the Tibetan Plateau snow cover on spring rainfall in southern China [25].

3.2. Spring Eurasian snow cover

Utilizing monthly mean 513-station rainfall data in China from the National Meteorological Information Centre of China spanning the period from 1968 to 2005 and monthly mean

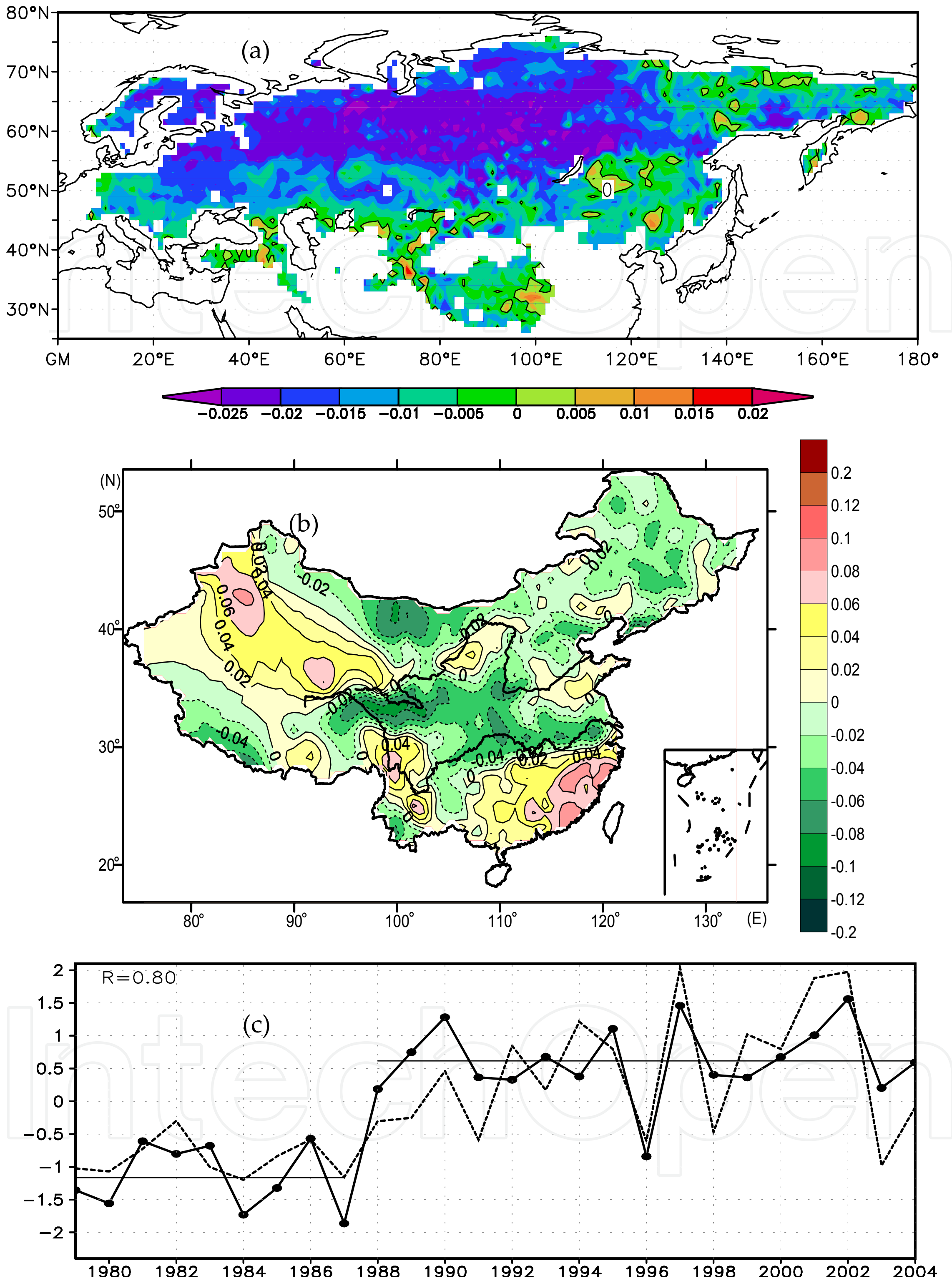


Figure 4. Spatial distributions of the SWE and rainfall fields of the leading SVD mode and their time series, derived from spring Eurasian SWE and the succeeding summer station rainfall in China during the period 1979-2004, (a) spring Eurasian SWE, only "0"-isoline is plotted, (b) summer station rainfall, and (c) normalized time series of spring SWE (solid line) and summer rainfall (dashed line), their correlation is 0.8. In (a) and (b), units are arbitrary. The straight lines indicate averaged time series in 1979-1987 and 1988-2004, respectively. (from [27])

snow water equivalent (SWE) data derived from National Snow and Ice Center during the period from 1979 to 2004 [26], the singular value decomposition (SVD) method is applied to calculate the coupled modes for the summer rainfall in China and spring (March-May) SWE during the period from 1979 to 2004 [27]. The leading SVD mode accounts for 36% of the co-variance, and its spatial distributions and corresponding time series are shown in Figure 4.

The leading SVD mode of the spring SWE variability shows a coherent negative anomaly in most of Eurasia (Figure 4a). It is seen that negative anomalies are dominant, particularly in the central Siberia, whereas positive anomalies are confined to some small areas including the most northeastern Russia, east and southeast to the Lake Baikal, the western and eastern Tibetan Plateau. Besides a strong inter-annual variability, the leading SVD mode of the spring SWE displays strong inter-decadal variation with persistent negative phases in 1979–1987 and frequent positive phases afterwards; an apparent inter-decadal shift occurred in the late 1980s with persistent negative phases during 1979-1987 (the mean value is -1.17) and predominant positive phases during 1988-2004 (the mean value is 0.62) (Fig. 4c). Figures 4a and 4c indicate that there was excessive spring SWE in Eurasia during 1979-1987, followed by abrupt decreases in the period from 1988 to 2004.

For the anomalous summer rainfall field corresponding to the leading SVD mode (Figure 4b), positive anomalies appear in southern China to the south of the Yellow River valley, and negative anomalies emerge in most parts of northern China to the east of 95°E. The relationship between spring rainfall variations in southern China and anomalous spring snow cover in western Siberia is in agreement to the previous results [25]. For the relation of the Eurasian winter snow cover with summer rainfall in China, the Eurasian winter snow cover is positively correlated with the following summer rainfall in northern and southern China and negatively in central, northeast, and western China [28]. Here we can see that the spring snow cover is mainly related with a reversed north-south summer rainfall anomalies in the eastern China, which differs from the rainfall anomalies associated with winter snow cover. The time evolution of the anomalous summer rainfall field shows that there is an upward trend, and coherent negative phases before 1989, followed by frequent positive phases afterwards, which are in good agreement with time series for spring Eurasian SWE (Figure 4c). The two fields are significantly correlated and their correlation coefficient is 0.8. The inter-decadal shift of the spring Eurasian SWE in the late 1980s corresponds well to that of the East Asian summer monsoon as shown by the WNP-EASM index in Figure 2, and the increasing summer rainfall in southern China shown in Figure 3b.

In order to check the summer rainfall difference between the periods before and after the late 1890s, Figure 5 shows the summer mean rainfall anomalies during 1979-1987 (Figure 5a) and 1988-2004 (Figure 5b), respectively, relative to the mean averaged for the period from 1979 to 2004. The pattern of rainfall anomaly in either Figure 5a or 5b resembles that in Figure 4b. In the inter-decadal time scale, accompanying spring excessive SWE in most of Eurasia during 1979-1987, summer rainfall decreased by about 40 mm in southern China, whereas it increased by about 20 mm along the Yangtze River valley (Figure 5a). In contrast, summer rainfall increased in southern China and decreased along the Yangtze River valley during 1988-2004 (Figure 5b). The results are same as those in that the East Asian summer monsoon experienced an inter-decadal shift in the late 1980s, and more rainfall moved from the Yangtze-Huai River valleys to southern China. The shift of the spring Eurasian SWE in the late 1980s may be one of reasons for the inter-decadal shift of the summer monsoon rainfall decrease around Yangtze River valley and increase in southern China.

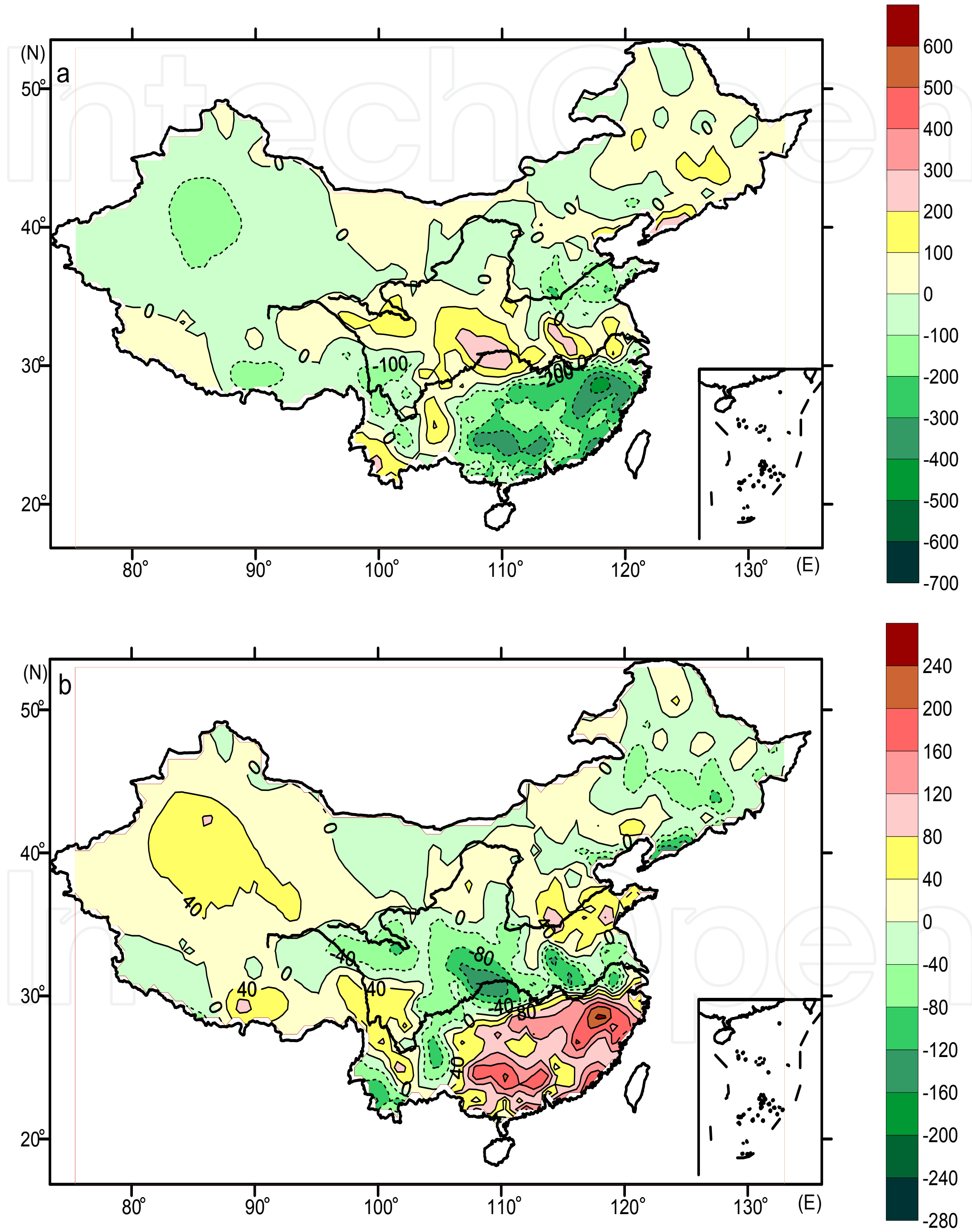
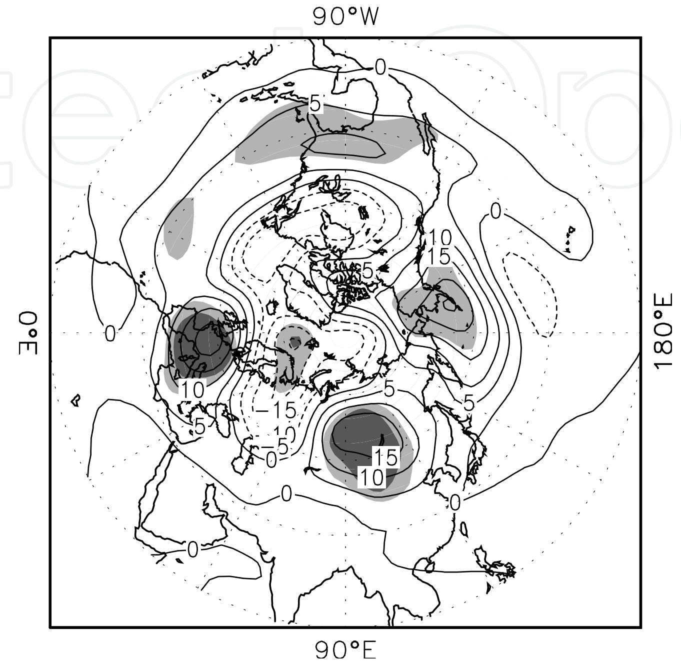


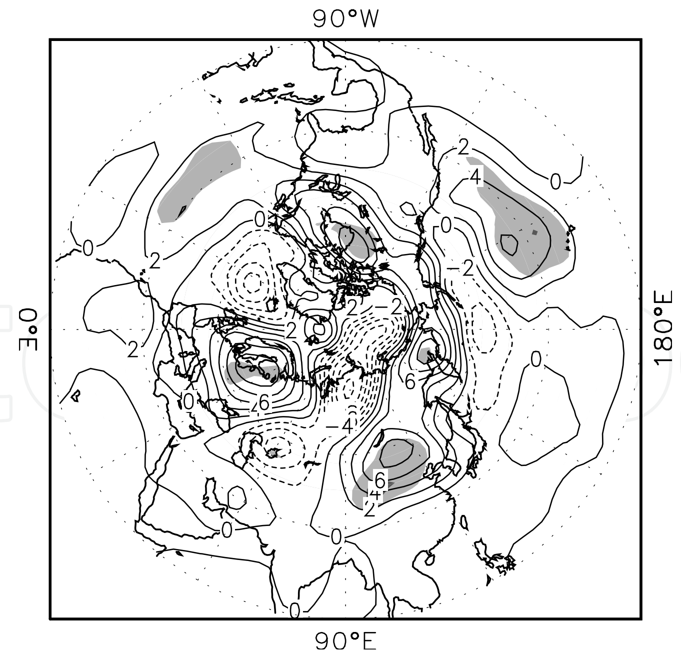
Figure 5. (a) Summer mean rainfall anomalies during 1979-1987 relative to the mean averaged for the period from 1979 to 2004, (b) same as (a) but for the period 1988-2004. In (a) and (b), the intervals are 100 and 40, respectively. Units are 0.1 mm.

3.3. Physical linkage between Spring Eurasian snow cover and summer rainfall

By using the NCEP/NCAR reanalysis data in 1960-2006, the physical process for the effects of spring Eurasian snow cover on the summer rainfall in China were diagnosed [27]. Figure 6 shows geopotential height anomalies at 500 hPa regressed by the leading SVD



(a)



(b)

Figure 6. Geopotential height anomalies at 500 hPa in (a) spring and (b) summer regressed by the leading SVD mode of spring SWE (unit: gpm). The contour intervals are 5 and 2 gpm in (a) and (b), respectively. (from [27])

mode of spring SWE in spring and summer, respectively. In both spring (Figure 6a) and summer (Figure 6b), a wave train structure can be observed over Eurasian continent. There are two positive anomalous centers located over western Europe and northeastern Asia, respectively. The positive height anomalies over northeastern Asia between 30°N and 60°N are unfavorable for the rainfall to the north of the Yangtze River valley, and more rainfall appears to its south in southern China. The similarity between the regressed height anomalies in spring and summer implies that the atmospheric circulation anomalies associated with Eurasian snow cover persist from spring to summer, which exerts significant effect on the summer rainfall in China.

As shown in Figure 6, in spring and summer similar features of the atmospheric circulation anomalies appear over East Asia. In fact, in 1979-2004 the spring rainfall in southern China also exhibited a climate shift in the late 1980s with more rainfall in southeastern China and less in southwestern China before the late 1980s, and less rainfall in southeastern China and more in southwestern China after then [29,30]. Figure 7 shows the leading SVD mode between Chinese spring rainfall (Figure 7a) and spring Eurasian SWE (Figure 7b). It explains 31.5% square covariance fraction and correlation coefficient between the expansion

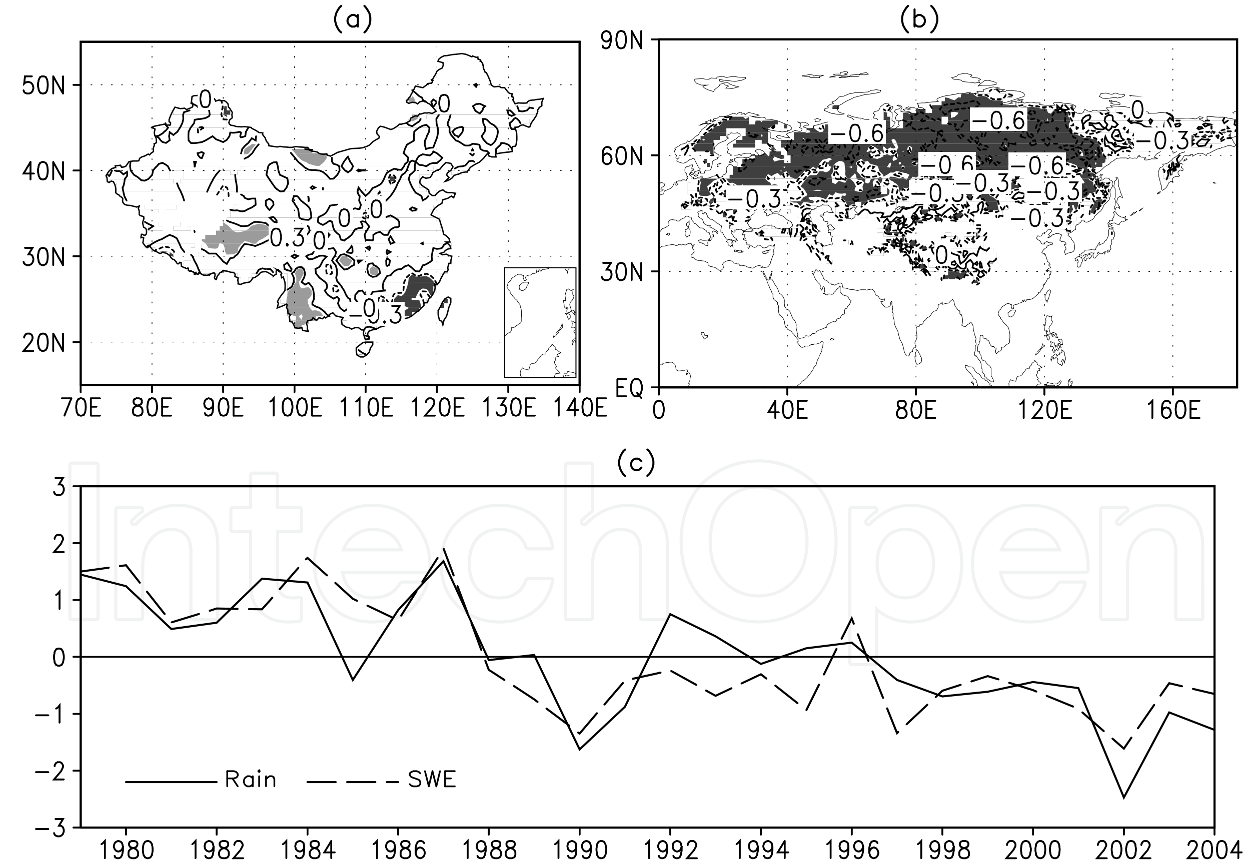


Figure 7. Spatial patterns of the leading coupled SVD mode for (a) spring rainfall and (b) Eurasian spring SWE and (c) corresponding time series (solid line for rainfall, dash line for SWE). Shaded areas in (a) and (b) denote the correlations exceeding the 0.05 significance level. (from [29])

coefficients of both variables is 0.81, exceeding the 0.01 significance level. The time series (Figure 7c) for both the Chinese spring rainfall and Eurasian SWE were mainly positive phase in 1979-1987 and frequently negative phase afterward. The decrease in Eurasian spring SWE is accompanied by the reduced rainfall over Southeast China and enhanced rainfall over Southwest China. The reduction in Eurasian SWE results in reduced upward and poleward wave flux activity, which alters the atmospheric circulation and thus affects the rainfall in southern China [30]. Actually, using the Monthly tabulated Scandinavia teleconnection index from NCEP (http://www.cpc.ncep.noaa.gov/data/teledoc/scand.shtml), the inter-decadal shift in the late 1980s can also be detected in the time series of the Scandinavian pattern [31] of atmospheric circulation variability. As shown in Figure 8, the positive phases of the Scandinavian pattern were dominant before 1988, followed by coherent negative phases. The abrupt shift of the atmospheric circulation possibly reflects the role of Eurasian spring SWE forcing.

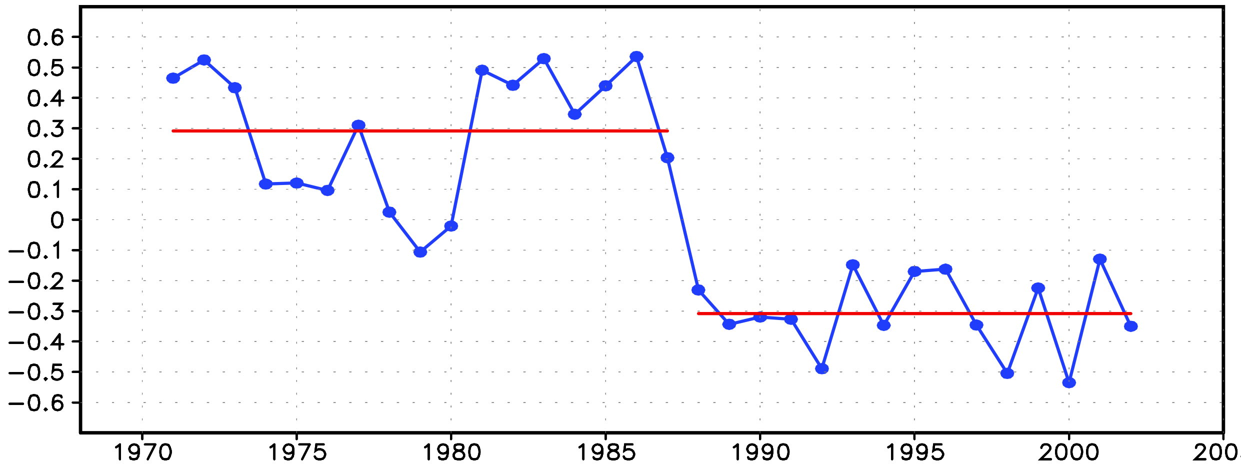


Figure 8. The 7-year running mean of the normalized time series of the Scandinavian pattern averaged over March to May (blue line). The red straight lines indicate averaged time series in 1971-1987 and 1988-2002, respectively.

4. Impacts of the Atlantic and Arctic Oceans

4.1. Atlantic Multidecadal Oscillation (AMO)

One of the predominant features in long-term SST variation in Atlantic is the Atlantic multidecadal oscillation (AMO). The AMO is the alternation of cool and warm SST anomalies (SSTAs) throughout the North Atlantic Ocean in the period of several decades [32,33]. A lot of studies have shown that the AMO has the influence in the global scale, and many regional multidecadal climate variability are related to the AMO [34].

To see the association of the summer precipitation in China with the AMO, the AMO positive phases of 1951-1965 and 1996-2000, and negative phase in 1966-1995 are selected [35]. The summer rainfall composites are made for the AMO positive phases and negative phase, respectively. The composite difference between positive and negative AMO phases is shown in Figure 9. It can be seen that except a small area in southwestern China, in the most areas in eastern China the summer rainfall is increased by 0.3-0.6 mm d^{-1}, implying that the warm AMO phase is favorable for more summer rainfall in eastern China [36].

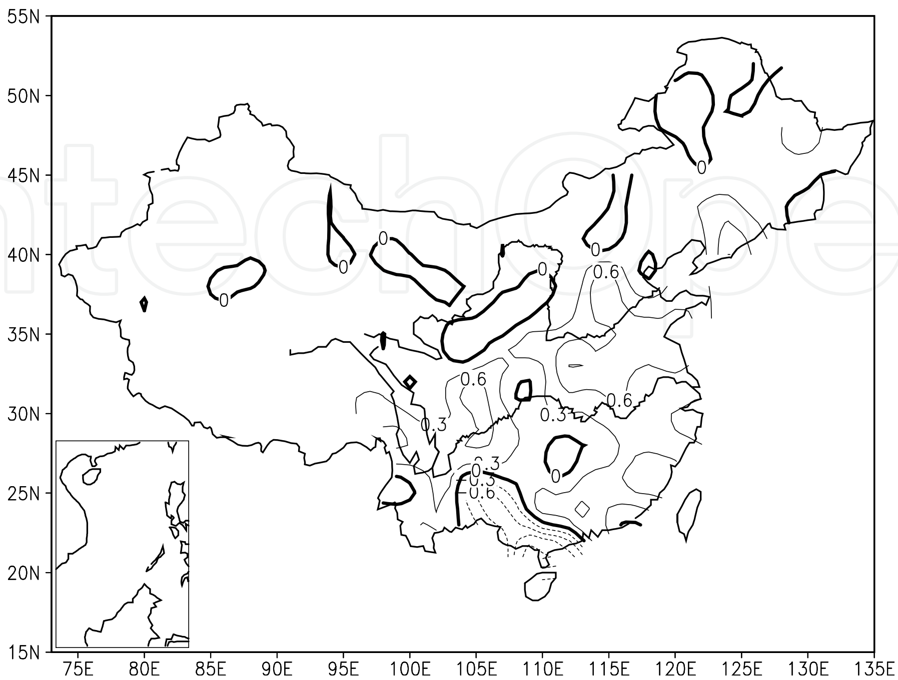


Figure 9. Difference of summer rainfall composites between positive and negative AMO phases (unit: mm d^{-1}). Thick lines are "0"-isolines. (from [36])

In order to verify the influence of the AMO on the summer rainfall in China, a coupled atmosphere-ocean general circulation model was utilized to perform numerical experiments [36], in which the SSTAs related with the positive or negative AMO phase in the Atlantic are specified, and the atmosphere-ocean interactions outside the Atlantic are allowed. The results of numerical experiments show in the positive AMO phase strong East Asian summer monsoon and more rainfall in eastern China, and reversed rainfall anomalies can also be found in the AMO negative phase, which resemble the observed summer monsoon rainfall in China in association with the AMO phases as shown in Figure 9. The model results suggested that, through coupled feedback, warm AMO phase result in the warmer SST in the eastern Indian Ocean and maritime continent. The warmer SST strengthens the convective heating over there and leads to an anticyclone anomaly over western North Pacific, which is responsible for the more rainfall over China.

4.2. Triple mode of North Atlantic SSTAs

In association with the seasonal march of the East Asian summer monsoon, climatologically the summer rainfall in eastern China moves northward in a stepwise way. The more rainfall zone exists mainly in southern China from mid-May to early June, which is referred to as the pre-Meiyu rainy season. In the period from mid-June to mid-July, the more rainfall zone shifts northward and stays around the middle and lower reaches of Yangtze River and Huai River valleys, which is named as the Meiyu. From late July to early August the more rainfall zone appears further northwards in North China.

By using long-term Meiyu dataset from 1885 to 2005 from the National Climate Center of China Meteorology Administration (CMA) and monthly SST data of HadISST from 1870 to date from the Hadley Center of the UK Meteorological Office [37], it is found that the decadal variability of the Meiyu precipitation amount and duration are closely correlated with a triple mode of North Atlantic SSTAs in the preceding winter [38]. The triple mode constitutes of three zonally elongated centers of SSTAs in the North Atlantic, with two positive SSTAs centered at about 50°N to the south of Greenland and at about 15°N in the subtropics, respectively, and one negative center to the east of the U. S. continent at about 30°N. An index I_{NA} of the triple mode is defined as the difference between normalized SSTAs averaged over the areas around two northern centers. Figure 10 shows the 9-16-yr band-pass filtered normalized Meiyu rainfall amount and duration as well as the I_{NA} index. In general, the Meiyu rainfall amount and duration coincide well with the I_{NA} index in the last century except short periods of 1930-1940 and 1985-1995. The triple mode of Atlantic SSTAs can persist from winter until late spring and excite a stationary wave-train propagating from west Eurasia to East Asia, exerting an impact on summer rainfall over eastern China.

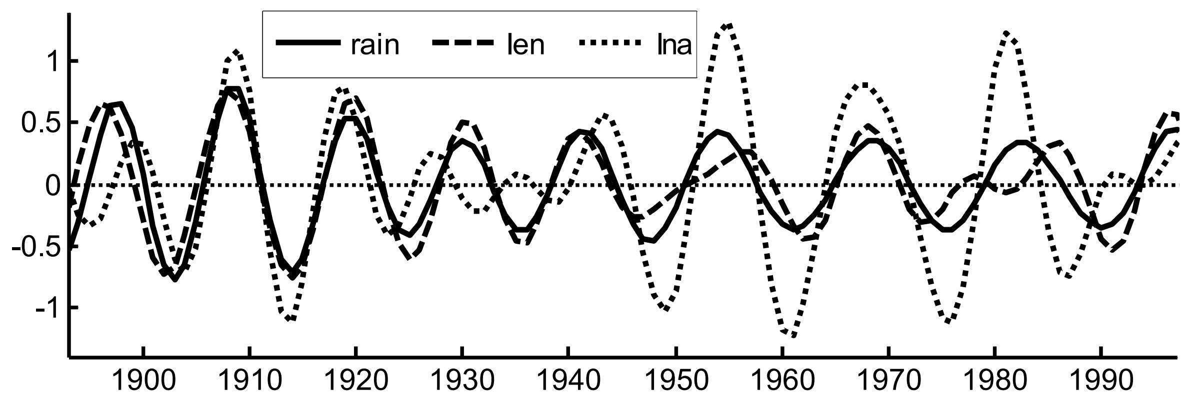


Figure 10. The 9-16-yr band-pass filtered normalized Meiyu rainfall amount (solid line) and duration (dashed line), and the I_{NA} index (dotted line). (from [38])

4.3. Spring Arctic sea ice

Utilizing the monthly mean Arctic sea ice concentration (SIC) dataset for the period of 1961–2007 obtained from the British Atmospheric Data Centre (BADC, http://badc.nerc.ac.uk/data/hadisst/), and a monthly 513-station rainfall dataset for China obtained from the National Meteorological Information Centre of China spanning the period from 1968 to 2005, a statistical relationship between spring Arctic SIC and Chinese summer rainfall is identified using singular value decomposition (SVD) [39]. The leading SVD mode accounts for 19% of their co-variance. As shown in Figure 11, both Arctic SIC and summer rainfall in the leading SVD mode display a coherent inter-annual variability (the correlation is 0.83, Figure 11a) and apparent inter-decadal variations with turning points occurring around 1978 and 1992, respectively. For spring SIC, negative phases were frequent during the periods 1968-1978 and 1993-2005, and separated by dominant positive phases. The

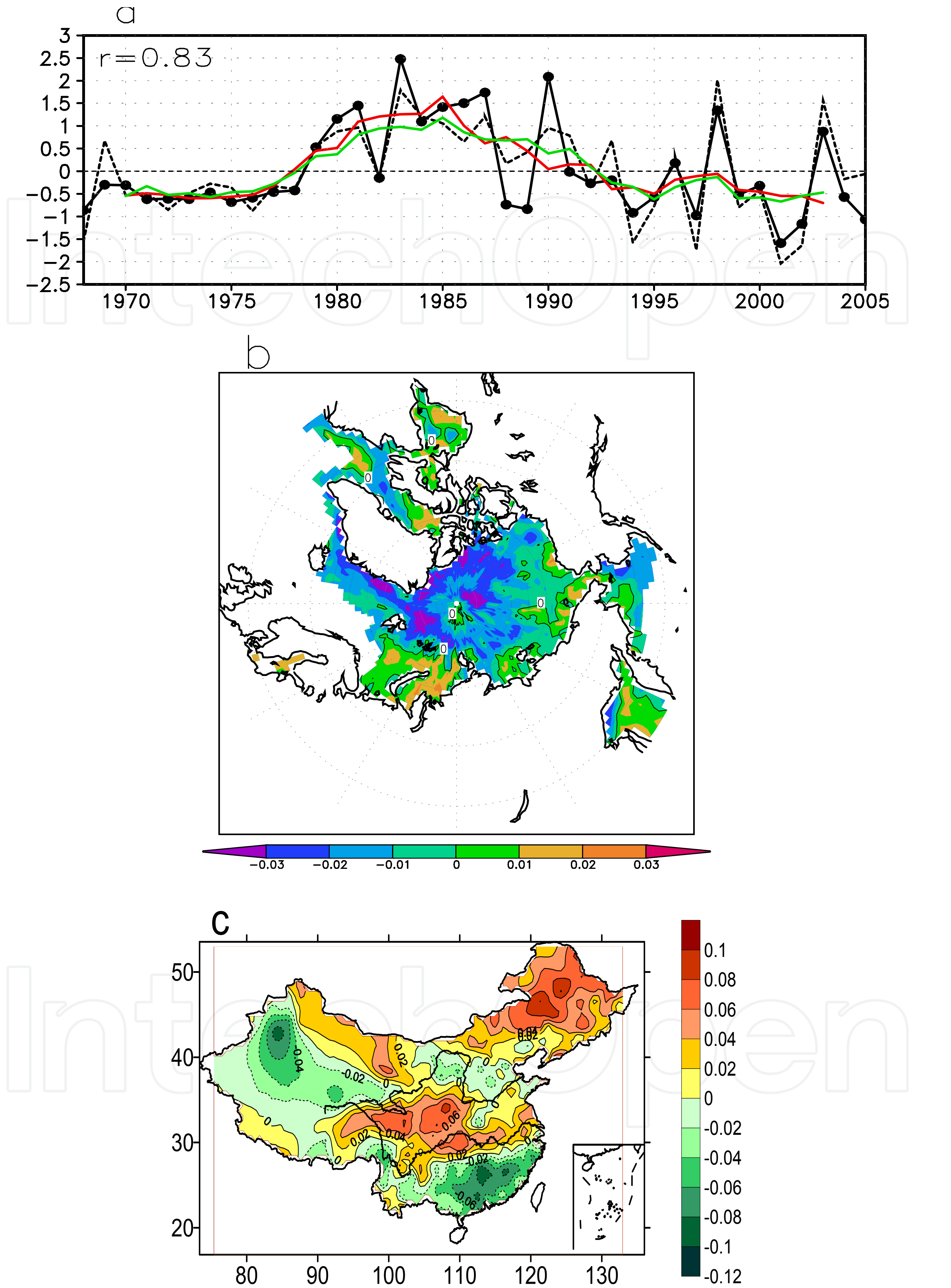


Figure 11. (a) Normalized time series of spring Arctic SIC (solid line) and summer rainfall (dashed line) variations in the leading SVD mode, the red and green lines denote their 5-year running means, respectively, (b) the spatial distribution of spring Arctic SIC in the leading SVD mode, (c) same as in (b) but for the summer rainfall. In (b) and (c), units are arbitrary. (from [39])

corresponding SIC anomalies were positive in most of Eurasian marginal seas, with negative SIC anomalies in the Arctic Basin, the Beaufort Sea, and the Greenland Sea during 1979-1992 (Figure 11b). Opposite SIC anomalies occurred during the periods 1968-1978 and 1993-2005. During the period 1979-1992, positive rainfall anomalies frequently appeared in northeast China and central China between the Yangtze River and the Yellow River (28°-36°N), with negative rainfall anomalies in south and southeast China (Figure 11c). Opposing spatial distribution of summer rainfall anomalies frequently occurred in the period 1968-1978 and 1993-2005.

The SVD analysis shows that in the decadal time scale the decreased (increased) spring SIC in the Arctic Ocean and the Greenland Sea corresponds to increased (decreased) summer rainfall in northeast China and central China between the Yangtze River and the Yellow River; and decreased (increased) rainfall in southern China. Figure 12 shows the regressed 500 hPa geopotential height in summer by the time series of the spring SIC leading SVD mode. The summer 500 hPa height anomalies associated with the spring SIC in the leading SVD mode show a Eurasian wave train structure, which originates in northern Europe and extends southeastwards to northeast China, and a south-north dipole structure over East Asia south to Lake Baikal. Compared to the regressed height anomalies in summer by the time series of the leading SVD mode for the spring Eurasian SWE (Figure 6b), the wave-train over the Eurasian continent is quite similar. In fact, the positive correlation exists between the time series of the leading SVD mode for the spring Eurasian SWE (Figure 4c) and that for the spring Arctic SIC (Figure 11a), and the correlation coefficient is 0.64 in the period of 1979-2004. Therefore, although a Eurasian wave train in association with the Arctic SIC can be found at 500 hPa lasting from spring to summer and extending southeastwards to northeast China, the spring Arctic SIC provides a complementary impact on the summer rainfall over China. The Eurasian wave train should result from combined effects of both spring Arctic SIC and Eurasian snow cover.

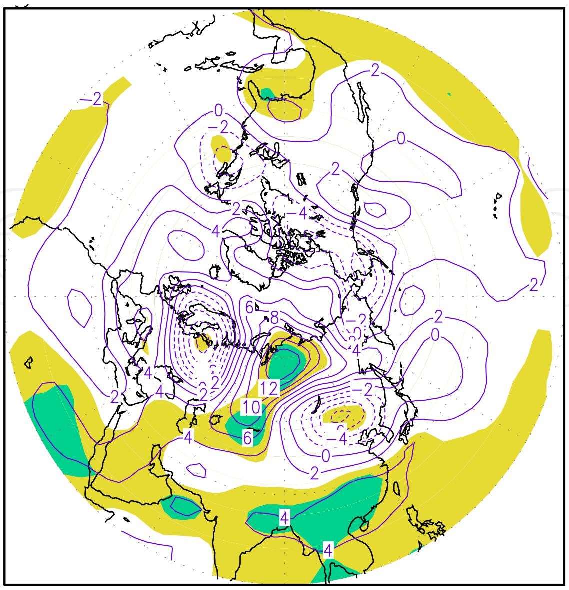


Figure 12. Regressed 500 hPa geopotential height in summer by the time series of the spring SIC leading SVD mode (unit: gpm). The yellow and blue shadings represent height anomalies exceed the 0.05 and 0.01 confidence levels, respectively. (from [39])

5. Impacts of the tropical Pacific and Indian Oceans

5.1. Central and eastern equatorial Pacific SSTAs

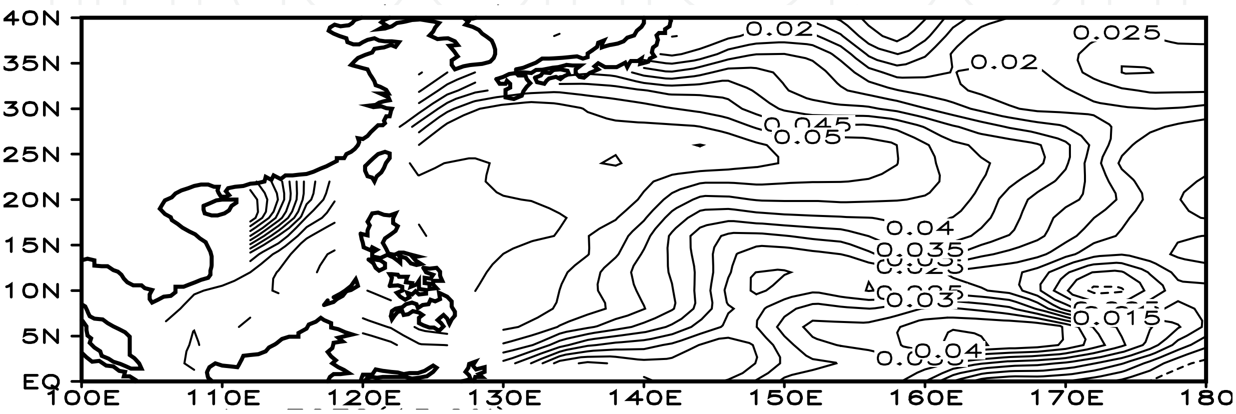
In the central and eastern equatorial Pacific, there exists large inter-annual variability of the SST. An El Niño is referred to as the period when the ocean temperature of upper layer in the central and eastern equatorial Pacific rises abnormally. In the atmosphere over the tropical Pacific, there is a widespread inter-annual oscillation in sea-level pressure between the area near northern Australia and that in the southeastern Pacific. This oscillation is called Southern Oscillation. Because of the intrinsic relationship between El Niño and Southern Oscillation, they are combined and referred to as ENSO (El Niño/Southern Oscillation). The El Niño is the most outstanding inter-annual variability in oceans. The occurrence of El Niño changes greatly the pattern of the thermal heating of the atmosphere, which leads to large circulation anomalies.

The El Niño has significant effects not only in the tropical region, but also in the extratropics. The influence of El Niño on the climate over East Asia has been intensively studied by many investigators [40]. It is found that the effect of El Niño on the summer rainfall in eastern China depends on the phase of the El Niño [41]. In the El Niño developing phase, more summer rainfall exists in the Yangtze and Huai River valleys and less over northern and southern China, whereas in the El Niño decaying phase opposite summer rainfall anomalies appear. The impact of El Niño on the East Asian climate is through an anomalous anticyclone appeared over the western North Pacific in the lower troposphere [12], which can significantly affect the rainfall in China by altering water vapor transport over East Asia [42]. This anomalous anticyclone acts as a bridge connecting the warm events in the eastern tropical Pacific and the East Asian monsoon [43].

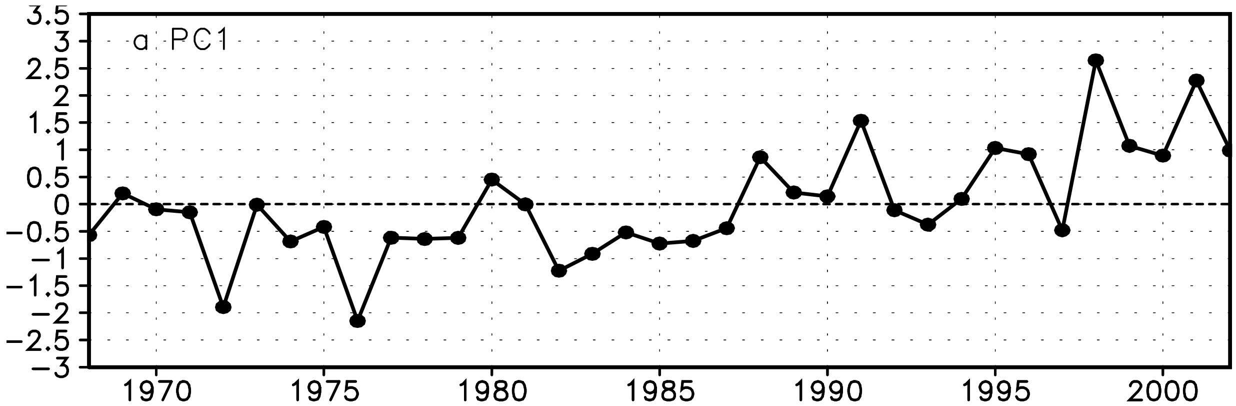
The above-mentioned researches focus on the inter-annual time-scale. The El Niño also appears in the inter-decadal time-scale [44]. By examining the variation of the 5-year running mean of SSTAs averaged over 0-10°N in the central and eastern equatorial Pacific, a concept of 'inter-decadal ENSO cycle' was proposed [18]. An 'inter-decadal El Niño event' appeared in the period from middle 1950s to late 1960s when SSTAs in central and eastern equatorial Pacific was above normal, with highest 0.6°C in middle 1960s; an 'inter-decadal La Niña event' from early to late 1970s with -0.6°C below normal; and from late 1970s to late 1980s an 'inter-decadal El Niño event' again with 0.4°C above normal. Same as the impact of El Niño on the summer rainfall in eastern China in the inter-annual time scale, the impact of the 'inter-decadal ENSO cycle' on the inter-decadal change of the summer rainfall in eastern China also depends on the phase of the 'inter-decadal El Niño event' [18]. In the decaying phase of the 'inter-decadal El Niño event' from middle 1960s to the middle 1970s, more summer rainfall appeared in northern and southern China and less in the Yangtze and Huai River valleys, while opposite distribution of the summer rainfall anomalies in the developing phase from middle 1970s to late 1980s. Such inter-decadal changes of the summer monsoon rainfall in eastern China are in good agreement with those shown in Figure 3a. Therefore, the inter-decadal changes of SSTAs in the central and eastern equatorial Pacific is possibly an important factor for the inter-decadal changes of summer monsoon rainfall in eastern China.

5.2. Western North Pacific and Indian Ocean SSTAs

The empirical orthogonal function (EOF) analysis is applied to the summer SST in the area of 0°-40°N and 100°-180°E in the western North Pacific during the period 1968-2002 [45]. The leading EOF mode, which has a variance contribution of 30.5%, and its time series are shown in Figure 13. The leading EOF mode shows a wholly consistent SST distribution (Figure 13a). The time series indicates a sustained inter-decadal variability in the first principle component (PC1) and an inter-decadal climate shift occurred in the late 1980s (Figure 13b).



(a)



(b)

Figure 13. (a) Leading EOF mode of the sea surface temperature in western North Pacific and (b) its time series. (from [45])

As seen from Figure 13b, the values of the PC1 were basically negative before 1988, after then became almost positive. Referring to Figure 13a, we know that in the late 1980s, the leading EOF mode of the uniformly consistent SST in western North Pacific had a significant inter-decadal climate shift, namely the SSTAs were negative before the late 1980s but positive afterwards. In North Pacific, the most striking inter-decadal variability is the Pacific decadal oscillation (PDO), and a climate shift for PDO occurred in late 1970s [46]. It can be seen from Figure 13b that the leading EOF mode of the western North Pacific SST did not show a climate shift in the late 1970s. With correlation coefficient between the summer mean PDO index and the leading mode being -0.06, the leading mode is not related with the PDO. Therefore, the leading EOF mode of the summer SST in the western North Pacific is independent from the PDO.

Here we can see that the inter-decadal shift of the western North Pacific SSTAs in the late 1980s coincided well with the climate shift of the East Asian summer monsoon. To illustrate summer rainfall anomalies in China related to the inter-decadal variation of the western North Pacific SSTAs, a 5-year running mean is performed on both the summer rainfall and PC1 time series in 1968-2002, and their correlation coefficients [7] are shown in Figure 14. Positive correlation coefficients exist in southern China to the south of the Yangtze River valley, indicating that lower (higher) western North Pacific SST is associated with less (more) rainfall in southern China. Compared Figure 14 to Figure 5, the distribution of the correlation coefficients resembles the summer rainfall anomalies, especially in southern China. The western North Pacific had lower SST before the late 1980s and higher afterwards, implying that the inter-decadal shift of the summer rainfall to the southern China in the late 1980s is closely associated with the inter-decadal change of the western North Pacific SST.

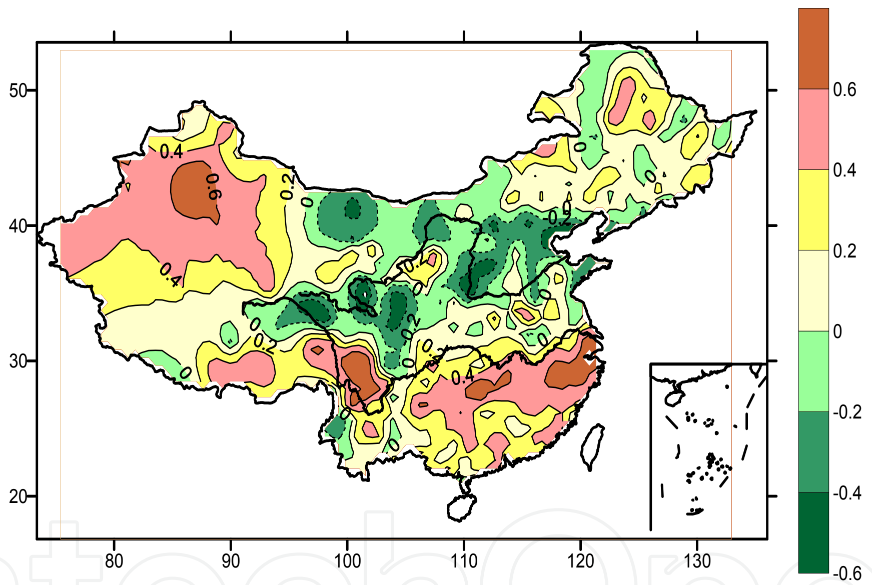


Figure 14. Correlation coefficients between 5-year running means of the summer rainfall in China and the PC1 time series of the western North Pacific SST. (from [7])

By checking the EOF modes of the summer rainfall in China in the period of 1958-2001, it is found the second EOF mode exhibits a north-south dipole distribution with opposite summer rainfall anomalies to the north and south of the Yangtze River valley [47]. An inter-decadal shift occurred in the late 1980s when less rainfall to the south of the Yangtze River valley and more to the north changed to be opposite distribution with more to the south of the Yangtze River valley and less to its north. The summer rainfall increasing after the late 1980s to the south of the Yangtze River valley is closely related with the warming of the summer SST in western North Pacific and northern Indian Ocean. The warmer summer SST in western North Pacific may affect the rainfall in eastern China through stimulating anomalous atmospheric circulations in the form of the EAP (East Asian-Pacific) teleconnection pattern over East Asia [48].

One of the most striking features of the inter-decadal shift of the East Asian summer monsoon is the westward extension of the western Pacific subtropical high (WPSH) [7,49,50], which affects the water vapor transport [1,51] and thus determines the location of the abnormal monsoon rainfall zone. Based on the ERA-40 Reanalysis data from the European Centre for Medium-Range Forecast (ECMWF) and taken the 5880 gpm isoline at 500 hPa in summer as the representative of the WPSH, Figure 15 shows the contour of the 5880 gpm at 500hPa in summer averaged in 1975-1989 and that averaged in 1990-2001, respectively, to illustrate the extent of WPSH [7]. It can be seen that, compared to that in the period of 1975-1989, the WPSH in the period of 1990-2001 became stronger, stretching farther westward with a larger south-north extent, which is favorable for the development of the southerlies over southern China. The strengthened southerlies are beneficial to the strengthening of the water vapor transport and therefore more precipitation in the south of the Yangtze River. Here we can see that, corresponding to the decadal climate shift of the East Asian summer monsoon in the late 1980s, both the summer circulation over East Asia and the summer rainfall in China changed significantly.

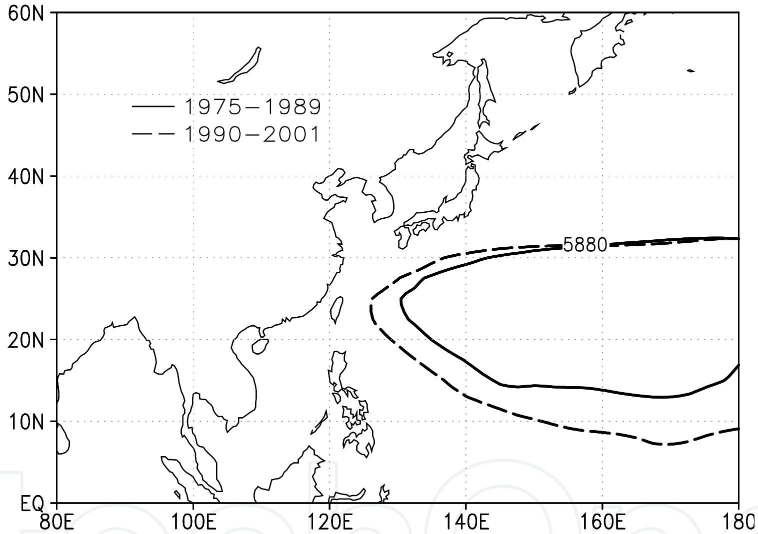


Figure 15. Summer (JJA) mean 5880 gpm contour at 500hPa averaged in 1990-2001 (dashed line) and in 1975-1989 (real line), respectively. (from [7])

Since the late 1970s, the SST in the tropical Indian Ocean-western Pacific (IWP) has increased over 0.4°C relative to that in the previous two decades [53]. By utilizing five atmospheric general circulation models (AGCMs), it is demonstrated that the forcing of the warming in IWP area is responsible for the westward extension of the WPSH [50]. All five AGCMs well reproduced the features that the WPSH extended further westward in warmer IWP period compared to that in the cooler period. Two ways for the IWP warming affecting the WPSH westward extension were proposed [50]. First, through Walker circulation the IWP heating leads to the weakening of convection in the central and eastern equatorial Pacific; the convective cooling anomalies force a response of the Gill-type anticyclone [54] in the North Pacific, which is favorable for the westward extension of WPSH. Second, the forced Kelvin wave response by the monsoon diabatic heating [55] reinforces the low-level equatorial flank of WPSH, while the poleward flow along the western flank of WPSH is intensified because of the Sverdrup vorticity balance.

6. Conclusion

Observations show that from 1958 to 2011 the predominant feature of the summer monsoon rainfall variation over eastern China is its inter-decadal variability besides the inter-annual variability, and no clear trend can be found for the summer monsoon rainfall. Same as the rainfall, strong inter-decadal variability also appears in the East Asian summer monsoon index (WNP-EASM index), and no trend can be found for the index. In addition to the inter-decadal variation, larger amplitudes and shorter periods for both East Asian summer monsoon index and monsoon rainfall appeared after the end of 1980s, implying a larger variability and more frequent variation of the East Asian summer monsoon have occurred since then.

In association with the inter-decadal variability of the East Asian summer monsoon, climate shifts for the East Asian summer monsoon appeared at middle 1970s, late 1980s and early 2000s, respectively. Corresponding to the climate shift in the middle 1970s, the summer rainfall increased in the area over the middle and lower reaches of the Yangtze River valley and decreased over southern and northern China. After the weakening of the East Asian summer monsoon in the late 1980s, more rainfall appeared in southern China. From the early 2000s, the East Asian summer monsoon strengthened and the rainfall decreased around Yangtze River valley, and increased to the south of about 25°N in South China and between about 31°N-36°N to the north of Yangtze River valley.

The inter-decadal variability of the Eurasian snow cover, Arctic sea ice, and SSTAs in Atlantic, tropical Pacific and Indian Ocean can be causes of the inter-decadal variability of the East Asian summer monsoon and associated monsoon rainfall over eastern China. By examine the leading SVD modes between springtime SWE over Eurasian Continent and summertime rainfall over China in the period 1979-2004, it shows an apparent inter-decadal shift occurred in the late 1980s. The leading SVD mode of the spring SWE displays strong inter-decadal variation with persistent negative phases in 1979–1987 and frequent positive phases afterwards. For the anomalous summer rainfall field corresponding to the leading SVD mode, positive anomalies appear in southern China to the south of the Yellow River valley, and negative anomalies emerge in most parts of northern China to the east of 95°E. In the inter-decadal time scale, the decreased (increased) spring SIC in the Arctic Ocean and the Greenland Sea corresponds to increased (decreased) summer rainfall in northeast China and central China between the Yangtze River and the Yellow River; and decreased (increased) rainfall in south China. The effects of spring Eurasian snow cover and Arctic sea ice on the summer rainfall in China are through a wave train of 500 hPa geopotential height anomalies, which persist from spring to summer and exert significant effect on the summer rainfall in China.

The composite difference of summer rainfall between positive and negative Atlantic AMO phases shows that in the most parts of eastern China the summer rainfall increases, implying that the warm AMO phase is favorable for more summer rainfall in eastern China. The results of numerical experiments by a coupled atmosphere-ocean general circulation mode suggest that, through coupled feedback, warm AMO phase result in the warmer SST in the eastern Indian Ocean and maritime continent. The warmer SST strengthens the convective heating over there and leads to an anticyclone anomaly over western North Pacific, which is responsible for the more rainfall over China.

The leading EOF mode of SSTAs in western North Pacific shows a significant inter-decadal climate shift in the late 1980s, namely the SSTAs were negative before the late 1980s but positive afterwards in the period 1968-2002. Lower (higher) western North Pacific SST is associated with less (more) rainfall in southern China. The inter-decadal shift of the summer rainfall over southern China in the late 1980s is closely related with the inter-decadal change of the western North Pacific SSTAs. The warmer summer SST in western North Pacific may affect the rainfall in eastern China through stimulating anomalous atmospheric circulations in the form of the EAP (East Asian-Pacific) teleconnection pattern over East Asia. The summer rainfall in eastern China also depends on the phase of the 'inter-decadal El Niño event'. In the decaying phase of the 'inter-decadal El Niño event' from middle 1960s to the middle 1970s, more summer rainfall appeared in northern and southern China and less in the Yangtze and Huai River valleys, while opposite distribution of the summer rainfall anomalies in the developing phase from middle 1970s to late 1980s. The location of the abnormal monsoon rainfall zone over eastern China is related with the westward extension of the western Pacific subtropical high, which is determined by the forcing of the warming in the tropical Indian Ocean-western Pacific since the late 1970s.

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