Review of Advance in Research on Asian Summer Monsoon

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Abstract

The Asian monsoon is the most significant component of the global climate system. During recent two decades, concerned efforts have been made to investigate the Asian monsoon. A substantial achievement has been made in basic physical processes, predictability and prediction since the MONEX of 1978-79. The major advance in our understanding about of the variability of the Asian summer monsoon has been highlighted in this paper. The present paper is structured with four parts. The first part present the introduction, indicating the new regional division of the Asian monsoon system and significant events of the history of the Asian monsoon research. The second part discusses the annual cycle and seasonal march of the Asian monsoon as the mean state, with a special emphasis on the onset, propagation, active-break cycle and withdrawal of the Asian summer monsoon which takes place in the near-equatorial East Indian ocean-central and southern Indochina Peninsula. The third part deals with the multiple scale variability of the Asian summer monsoon, including the intraseasonal, interannual and inter-decadal variability. Their dominant modes such as 10-20 day and 30-60 day oscillations for the intraseasonal variability, the Tropospheric Bennial Oscillation (TBO), the Indian Ocean Dippole Mode (IODM) and teleconnection patterns for interannual variability and the 60-year oscillation for the inter-decadal variability, as well as related SST-monsoon relationship and land-monsoon relationship have been discussed in more details. The fourth part is the conclusion, summarizing the major findings and proposing future work.

Introduction

The Asian monsoon is characterized with a distinct seasonal reversal of wind and clear partition between dry and wet season in the annual cycle, which is related with the seasonal reversal of the large-scale atmospheric heating and steady circulation features (Webster et al. 1998; Ding and Chan 2005; Trenberth et al. 2006). About two decades ago, the Asian monsoon was mainly viewed as the Indian or the South Asian monsoon (ISM) in the English literature. The summer monsoon over East Asia was just thought of as the northward extension of the Indian monsoon, on the one hand, and on the other hand, the summer monsoon over the Western North Pacific (WNPSM) was fully taken to be the eastward extension of the Indian monsoon (Ding 1994). However, now it has been increasingly realized that the Asian monsoon system should further extend to incorporate these two regions due to similar features of monsoon climate, as a large amount of literatures has been contributed to the study of the East Asian summer monsoon (EASM) (Chang 2004) and the WNPSM in recent two decades (Murakami and Matsumoto 1994; Ueda and Yasunari 1996; Wu and Wang 2001; Wu 2002; Wang and Lin 2002; Wang et al. 2005a; Wang et al. 2005c). Thus, the Asian-Pacific monsoon is demarcated into three sub-systems: the Indian summer monsoon, the western North Pacific summer monsoon and the East Asian summer monsoon The EASM domain defined by Wang and Lin (2002) includes the region of 20-45°N and 110-140°E, covering eastern China, Korea, Japan and the adjacent marginal seas. This definition does not fully agree with the conventional notion used by Chinese meteorologists (Tao and Chen 1987; Ding 1994; Ding and Chan 2005), who usually includes the South China Sea (SCS) in the EASM region. Wang and Lin (2002) believe that the ISM and WNPSM are tropical monsoons in which the low level winds reverse from winter easterlies to summer westerlies, whereas the EASM is a sub-tropical monsoon in which the low-level winds reverse primarily from winter northerlies to summer southerlies. However, if the SCS region is included in the EASM, the EASM should be a hybrid type of tropical and subtropical monsoon.

In the ISM region, the monsoon or seasonal changes of winds and rainfall could be interpreted as a result of northward seasonal migration of the east-west oriented precipitation belt accompanied by the inter-tropical convergence zone (ITCZ or TCZ), while in the EASM region the seasonal march of the monsoon is displayed in the northward excursion of the planetary frontal zone or Meiyu-Baiu frontal system. So, one of main differences between these two monsoon sub-regions is the different effect of mid-and high

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latitude events. In the WNPSM region, the ITCZ is the dominating weather system which is the major birthplace of typhoons and tropical convective systems. At the upper level, the Asian monsoon region is dominated by the Tropical Easterly Jet (TEJ) to the south of the huge South Asian high. The major monsoon rainfalls are located in the right sector of the jet entrance zone where the upper-level divergence and low-level convergence, i.e. upward motion is observed (Fig. 1b,c) (Hoskins and Wang 2006; Chen et al. 2006). In short, the Asian monsoon system is a huge monsoon system and constitutes an essential part of the Asian-Australian monsoon system. On the other hand, it is closely related to the African monsoon system through the TEJ.

Research on the Asian monsoon has a long history (Webster 2006), but the substantial progress has been made since 1960's through a number of international and regional monsoon projects and field experiments such as the International Indian Ocean Expedition (IIOE) in the mid-1960s and in 1975-1976, the FGGE Monsoon Experiments (MONEX) in 1978-1979 (Krishnamurti 1985), the Coupled Ocean-Atmosphere Response Experiment (TOGA-COARE) in 1992-1993 (Webster and Lukas 1992), the GEWEX Asian Monsoon Experiment (GAME) in 1996-2000 (GAME ISP, 1998), the South China Sea Monsoon Experiment (SCSMEX) in 1996-2000 (Lau et al. 2000; Ding and Liu 2001; Johnson et al. 2004), the Bay of Bengal Monsoon Experiment (BOMEX) in 1999 (Bhat et al. 2001) and the Joint Air-Sea Monsoon Interactive Experiment (JASMINE) in 1999 and the Arabian Sea Monsoon Experiment (ARMEX) in 2002 (Webster et al.2002). At the same time, a number of books, monographs and review papers written in English summerizing the major scientific achievements of the Asian monsoon in different periods have been published, including Monsoon Meteorology (Ramage 1971), Southwest Monsoon (Rao 1976), Monsoons (Fein and Stephens 1987), Monsoon Meteorology (Chang and Krishnamurti 1987), Monsoon over China (Ding 1994), The East Asian Monsoon (Chang 2004), The Global Monsoon System (Chang et al. 2005) and The Asian Monsoon (Wang 2006). They provide international scientific communities with continuously updated input of knowledge of observations, processes, dynamics, prediction and socio-economic impact of the Asian monsoon. Now, it has been realized that the Asian monsoon can not only have significant disastrous events, but also can exert very important influences on the global climate system and the global climate prediction through, e.g., the tropical-extratropical interaction, the monsoon-ENSO relationship and the global teleconnection (Webster et al. 2005b; Ding and Wang 2005; Wang et al. 2005a).

During recent two decades, a more and more attention has been paid to study the Asian monsoon. Recent studies on the Asian monsoon have been devoted to the following aspects: (1)

Annual cycle and seasonal march of the Asian monsoon

The onset and the seasonal march of the Asian summer monsoon

The Asian monsoon shows a strong annual cycle which distinguishes it from the other monsoon systems over the world that have much weaker annual cycles. Observations indicate that the largest amplitude of the annual cycle of precipitation, zonal and meridional winds at low-level occurs in the Asian monsoon regions where there is the strongest atmospheric heating or heat sources driving the monsoon. In the process of the annual cycle, the onset of Asian summer monsoon is a key indicator characterizing the abrupt transition from the dry season to the rainy season and subsequent seasonal change (Lin and Wang 2002; Qian et al. 2002; Wang and Ding 2006). Numerous investigators have studied this problem from the regional perspectives. It is to some extent difficult to obtain a unified and consistent picture of the climatological onset of the Asian summer monsoon in different regions due to differences in data, monsoon indices and definitions of monsoon onset used in these investigations. Ding (2004) has summarized the climatological dates of the onset of the Asian summer monsoon in different monsoon regions based on various sources, with dividing the whole onset process into four stages: Stage 1 (late in April or early in May): the earliest onset in the continental Asia is often observed in the central Indochina Peninsula late in April and early in May. Over the ocean, one may trace earlier onset in the near-equatorial East Indian Ocean in late April. Stage 2 (from mid- to late May): this stage is characterized by the areal extension of the summer monsoon, advancing northward up to the Bay of Bengal and eastward down to the South China Sea. Stage 3 (from the early to the middle June): this stage is well known for the onset of the Indian summer monsoon and the arrival of the East Asian rainy season such as the Meiyu over the Yangtze River Basin and the Baiu season in Japan. At the same time, the summer monsoon over the western North Pacific extends from the SCS to the southwestern Phillipine Sea, accompanied by the increase in convective cloudiness and the eastward shift of the ITCZ (Wu and Wang 2001; Wu 2002). Stage 4 (the early and middle July): the summer monsoon at this stage advances up to Northwest India, Pakistan North China, the Korean Peninsula (so-called Changma rainy season) and even Central Japan. After mid-July, the WNPSM and associated ITCZ further marches northeastward as monsoon rainfall and active convection abruptly penetrate to the region of 25°N, 150°-160°E in Pentad 42 (July 25-29) (Ueda et al. 1995). This is the maturing process of the WNPSM which may maintain till late September, the major period of the typhoon season of the western North Pacific.Summer mons (Ueda and Yasunari 1996; Wu and Wang 2001;/ rasul et at.2004; Wang et al. 2005c, 2006).

The earliest onset occurs in the end of April (the 24th pentad) in the near-equatorial East Indian Ocean and Sumatra. Then the onset process propagates northeastward and northwestward, respectively. In western coast of the Indian subcontinent, the onset begins first across the Kerala coast, normally by the end of May or early June (Ananthakrishnan and Soman 1988) when heavy rains lash the coastal state after the cross-equatorial low-level jet is established across the Somali jet into the near-equatorial Arabian Sea and a cyclonic vortex, so-called onset vortex (Krishnamurti et al. 1981) is often observed. Then the Indian summer monsoon gradually advances northward across the western Indian subcontinent and eventually merges with the summer monsoon simultaneously propagating northwestward from the Bay of Bengal. By the middle of July the whole of the Indian subcontinent comes under the grip of summer monsoon. Therefore, the onset of the Asian summer monsoon is one of most important events, as a singularity of annual cycle of the monsoon. Furthermore, the onset process of the summer monsoon in a certain location or region is very rapid or even abrupt, with taking a couple of days or one week to complete the dramatic change from dry season to rainy season. It clearly shows abrupt changes in rainfalls, OLR and 850hPa zonal wind around the onset dates for the South China Sea, the Indochina Peninsula and the Indian subcontinent, respectively. The rob of convective activity during the onset process is very significant The abrupt increase in precipitation after the summer onset was previously indicated by various investigators, e.g., Ananthakrishnan and Soman (1989) for south Kerala, Matsumoto (1997) for the central Indochina Peninsula and Tao and Chen (1987) for the SCS.

The onset of the Asian summer monsoon brings with it a dramatic change in large-scale circulation features and weather situation. Numerous investigators have examined this problem from climatological and synoptic perspectives based on the large-scale wind, geopotential height, precipitation and OLR patterns (Lau and Yang 1997; Matsumoto 1997; Fong and Wang 2001; Ding and Liu 2001; Ding and Sun 2001; Wang and Lin 2002; Liu et al. 2002; Goswami 2005c; Zhang et al. 2004; Ding and Liu 2006a). Their detailed description will not be given, and instead, based on these studies, the sequence chain of significant events during the onset of the Asian summer wind may be identified below:

A. The development of a cross-equatiorial current in the equatorial East Indian Ocean (80-90°E) and off the Somali coast, and the rapid seasonal enhancement of heat sources over the Indochina Peninsula, South China, Tibetan Plateau, and neighboring oceanic areas, thus leading to seasonal reversal of sign of the tropospheric meridional temperature gradient in the Asian monsoon region.
B. The acceleration of low-level westerly wind in the tropical East Indian ocean;
C. Formation of a mid-tropospheric shear zone across the central Bay of Bengal to the Southeast

Arabian Sea in which may be embedded a cyclonic vortex in most of cases, which may even intensify into a cyclonic storm either in the Bay of Bengal or the Southeast Arabian Sea (the onset vortex).

D. The eastward and northward expansion of tropical monsoon from the tropical East Indian Ocean, initiating the arrival of the summer monsoon and associated rainy season in the regions of the Bay of Bengal and Indochina Peninsula with involvement of impacts from mid-latitudes;

E. The significant weakening, breaking around the Bay of Bengal and eastward retreat of the main body of the subtropical high, and eventual onset of the SCS summer monsoon with convective clouds, rainfall, low-level southwesterly wind and upper-level northeasterly wind suddenly developing in this region.

F. The ITCZ or TCZ usually used in the South Asian region and accompanying zonally oriented precipitation belt rapidly moves from a mean position about 5°S in winter and early spring to about 20°N in northern summer with the progress of the monsoon onset in the Indo-pak subcontinent.

G. Significant development and northwestward movement of the South Asian high at 200hPa, thus allows the establishment of the tropical easterly jet on its southern flank and rapid northward jump of the upper-level westerly jet on its northern flank over the northern Tibetan Plateau. This signals the start of rainy season and noth eastern Pakistan as well as Kashmir nomally get first monsoon shower during first week of July (Rasu etat 2003).

Numerous Scientists have studied the problem on generation and maintenance of deep tropospheric heat source in the north which is a central issue of the onset of the Asian summer monsoon. They have found that the heating over the Tibetan Plateau plays an important role in the seasonal evolution of the meridional gradient of heating and in triggering the onset of the Asian summer monsoon (Murakami and Ding 1982; Ding 1992a; Yanai et al. 1992; Wu and Zhang 1998; Ueda and Yasunari 1998; Zhang et al. 2004; Yanai and Wu 2006). Goswami (2005c) used the evolution of the mean temperature of the tropospheric layer between 200hPa and 700hPa averaged between 30°E and 110°E to illustrate the seasonal evolution of the meridional gradient of tropospheric heating. The change in the sign of meridional gradient of the tropopheric temperature ushers in the setting up of an off-equatorial large-scale deep heat source. The monsoon onset itself is a problem of internal atmospheric dynamic involving the instability of the basic flow in tropics. However, the change in the large-scale heat source is a necessary prerequisite. The atmospheric response to such a heat source leads to cross-equatorial flow and strengthening of the low-level westerlies. This causes the zero absolute vorticity line in lower atmosphere to move north of the equator to about 5°N facilitating symmetric inertial instability (Tomas and Webster 1997; Krishna Kumar and Lau 1998). The change in the sign of meridional gradient of tropospheric temperature thus makes the circulation conductive for symmetric instability that forces frictional boundary layer convergence, overcomes the inhibition from the subsidence above the planetary boundary layer over the Indian continent and the Bay of Bengal caused by compensating subsiding air of the convection over the Maritime Continent during April and May and capping effect of southward flow of dry and colder air on pre-existing moist and warm air, and leads to explosive development of off-equatorial convection over India and the Bay of Bengal and the subsequent onset of the Indian summer monsoon.

The reversal of the meridional temperature gradient has a significant regional difference, with the earliest reversal taking place in the region from the eastern part of the Bay of Bengal to the Indochina Peninsula in the early May (Wu and Zhang 1998; Zhang et al. 2004; Ding and Liu 2006a). And this reversal process was first found in the upper troposphere and then it rapidly propagated downward, mainly through vigorous convective vertical transport of strong sensible and latent heat through

(Yanai et al. 1973; He et al. 1987; Yanai et al. 1992; Li and Yanai 1996; He et al. 2006). The sensible heat driven air-pumping effect also plays a considerable role in vertical heat transport (Yanai and Wu 2006). Then, the reversal process extends eastward to the SCS (the 4th pentad of May) and westward to the Indian Peninsula (the 1st-2nd pentad of June). Timing of the large-scale reversal of the meridional temperature gradient is fully consistent with onset dates of the summer monsoon in these regions. The cause why the earliest reversal of the meridional gradient occurs in the Indochina Peninsula rather than the Indian Peninsula is attributed to earlier occurrence of sensible and latent heating and vigorous convective activity over the land of Indochina Peninsula under favorable effect of upward motion ahead of the Indian-Burma trough. In contrast, at this time the Indian Peninsula is located behind the trough where downward motion suppresses the development of convection (Wu and Zhang 1998; Zhang and Qian 2002; He et al. 2006; Ding and Liu 2006a).

The summer monsoon progress over the Western North Pacific (WNP) cannot be explained by the change in the tropospheric heat source that reflects the land-sea thermal contrast. The results obtained by most of investigators suggest that the summer monsoon advance over the WNP may result from air-sea interaction. The key is to understand the role of SST change and SST gradient as well as related feedback processes. The zonal asymmetry in SST between western and eastern Pacific along 10°-20°N is thought to be a leading factor for the development of the WNP summer monsoon (e.g., Murakami and Matsumoto 1994). Ueda and Yasunari (1996) indicated the importance of the development of a warm SST tongue around 20°N and 150°-160°E in early July. The northeastward monsoon advance over the WNP seems to follow the northeastward extension of the warm SST tongue (Wu and Wang 2001) which is created by the cloud-radiation and wind-evaporation feedback processes. Monsoon-induced changes in cloudiness and surface wind produce contrasting changes in surface short-wave radiation, and latent heat flux between the convection and pre-convection region. The resulting SST tendency difference turns around the SST gradient east of the convection region in about one month, and induces the northeastward shift of the highest SST center. Under this condition, the convective instability and low-level moisture convergence increase, thus triggering convection and atmospheric destabilization and subsequent monsoon onset.

After the onset of the Asian summer monsoon, the monsoonal rainfall, the TCZ or ITCZ and other properties or variables (e.g., Psudo equivalent potential temperature θ_{se}) assume significant northward advection, but with quite marked regional differences. In India, the seasonal advance of the monsoon gradually proceeds northwards. However, the northward progress is not a fully smooth affair as well and takes place often in surge or in stages, sandwiiched by periods of weakening or stagnation of monsoon activity (Gadgil and Kumar 2006). For each surge or stage, the advance process is accompanied by a synoptic-scale disturbance pushing the leading boundary of monsoon either northward or westward. Along the Indo-Ganetic Plains the advance of the monsoon occurs from east to west and is associated with the formation of 2 or 3 monsoon depressions/lows (Ding and Sikka 2006). The monsoon advances takes nearly 3 to 4 weeks to cover the entire Gangetic Plains. Thus, the advance of the monsoon over the entire subcontinent is a rather slow process taking on average nearly 40-45 days from its start off Kerala around 31 May to its culmination by mid-July over central Pakistan. Pakistan receives summer monsoon associated with southeasterly current in the north which set up in late June or early July and southwesterly current in south established in middle of July (Rasul et at 2004). Onset over Kashmir and northeast Pakistan takes simultaneously on 1st July with a deviation error of 15 days.. Similarly on southern half of the country, rains start around 15 July with 16 days standard deviation(Rasul et at 2005). The ITCZ or monsoon trough and accompanying major rain belt in this region move farther northward compared to the ITCZ in the WNPSM region and reach the northernmost latitude (~25°N) in mid-July. The climatological pentad-mean maximum rainfall rate (10-12 mm day-1) occurs in early June. Note also that another rain belt is located at 5°S, and the Indian monsoon rain has a close linkage with the equatorial convection throughout the entire summer. On the other hand, in the East Asian-West Pacific regions, the ITCZ or the tropical rainfall belt is located in the latitudinal range of 5~25°N, after the onset of the summer monsoon. There is no such near-equatorial convective band as in the Indian region. The maximum intensity and the northernmost location (~25°N) are observed in August when the subtropical high over the western North Pacific considerably moves northward. North of the ITCZ, there is another significant seasonal rain belt which is separate from the ITCZ and propagates northward in a stepwise way, not in continuous way. Numerous studies have demonstrated that it generally undergoes three standing stages and two stages of abrupt northward shifts (Ding 1992b). These stepwise northward jumps are closely related to seasonal changes in the general circulation in East Asia, mainly evolution of the planetary frontal zone or the Meiyu-Baiu frontal zone, the upper-level westerly jet stream and the subtropical high over the western North Pacific. The peak rainy seasons tend to occur primarily in three standing stages i.e., the first standing period from May to mid-June for the pre-summer rainy season in South China and Taiwan, the second standing period from mid-June to mid-July for the Meiyu-Baiu-Changma rainy season, and the third standing period from mid-July to mid-August for the rainy season in North and Northeast China and tropical western North Pacific. From the end of August to early September the monsoon rain belt rapidly moves back to South China, with most of the eastern part of China dominated by a dry spell, which symbolizes the termination of the East Asian summer monsoon. Therefore, in the EASM region, the monsoon rainy season has a shorter duration and weaker precipitation intensity than the ISM and the WNPSM which generally end in the late September and the late October, respectively (Li et al. 2005; Wang et al. 2005a,c). In addition, in the EASM region, three separate low-level westerly wind systems originating in the mid-latitude, subtropical and tropical regions, respectively, actually exist, and their interaction sets a large-scale stage for occurrence of many significant weather and climate events in this region that is very unique in the world monsoon climate zone.

Active-break cycle and retreat of the Asian summer monsoon

The active and break periods of the monsoon are characterized by precipitation maxima and minima over South Asia, depending on the season (Webster et al. 1998). These periods are thought to be associated with shifts in the location of the monsoon trough in India (Krishnamurti and Ardanuy 1980; Ding and Sikka 2006). During the active period of the monsoon, the trough is generally located in the central and northern Indian Peninsula. The break monsoon phenomenon is a reverse of the active monsoon spell over central and northern India. During the monsoon break the monsoon trough hugs the Himalayan rim, the low-level easterlies disappear entirely along the Indo-Gangetic plains and are replaced by west-northwest flow along the periphery of the Himalayas, resulting in decrease of rainfall over much of India, but enhanced rainfall in the far north and south. These anomalies are large-scale and extend across the entire South Asia. Active and break cycles vary in duration and may last for a few days and weeks. They are significant variations on the intra-seasonal scale to cause alterations between wet spells and dry spells. The phenomenon of break monsoon has been of interest because prolonged droughts often occur during intense break. For example, a prolonged break situation in the peak monsoon of July in 2002 not only resulted in rainfall deficit of the month of July, but also caused a seasonal-scale drought over the whole country in India (Gadgil and Joseph 2003; Gadgil and Kumar 2006). The break and active periods clearly demonstrate the two modes of the South Asian monsoon system at the intra-seasonal oscillation. The meridional shear of the low-level zonal wind and cyclonic vorticity at 850hPa are significantly enhanced (weakened) during an active (break) phase of the ISO. Hence, conditions for cyclone genesis are more favorable during an active phase compared to a break phase. Based on the analysis of genesis and tracks of low pressure systems (LPS) for 40 years (1954-1993), Goswami et al. (2003) show that genesis of the LPS is nearly 3.5 times more favorable during an active condition compared to a break condition of the monsoon (Fig. 5b). They also show that the LPS are spatially strongly clustered to be along the monsoon trough region during an active phase.

The above result show that there is an association of active and break periods of the monsoon with the ISO. Numerous investigators have studied this problem, with a special emphasis on the role of 30-50 day oscillation (MJO mode) and 10-20 day oscillation in modulating the active-break cycle of the summer monsoon (Yasunari 1979; Sikka and Gadgil 1980; Krishnamurti and Surgi 1987; Goswami 2005c). Their observations have revealed that the active and break period of the Indian summer monsoon or the wet and dry spells over the Indian continent are manifestation of repeated northward propagation of the TCZ or the monsoon trough from the equatorial position to the continental position and results from superposition of 10-20-day and 30-60-day oscillations. Krishnamurti and Ardanuy (1980) examined the 10 to 20-day waves at 20°N, utilizing 30 years of sea level pressure. They noted that alternate passage of troughs and ridges of these west-propagating oscillations modulated the pressure of the monsoon trough over northern India, and there is a strong relationship between the arrival of the ridges of these waves over the central India and occurrence of breaks in the monsoon. On the other hand, the arrival of ridges of northward and eastward propagating 30-50-day mode over the region of a climatologiacal monsoon trough has been known to coincide with the periods of occurrence of breaks in the summer monsoon (Yasunari 1980; krishnamurti 1985). The nearly simultaneous arrival (phase-locking) of ridges of these two ISO modes appeared to modulate the pressure of the monsoon trough considerably. Recently, a number of investigators have found similarities in the ISO and the inter-annual oscillation of monsoon system (Krishnamurthy and Shukla 2000; Lawrence and Webster 2001). This is not surprising as it is due to the dominance of two modes of active and break of monsoon activity in the monsoon season (Ding and Sikka 2006).

The breaks in the monsoon also occurs under mid-latitude interaction in which case a westerly trough in the upper-middle troposphere and the associated western disturbance in the lower troposphere greatly influence the activity of the monsoon trough over the Indo-Pakistan region. The western disturbances from mid-latitude can shift the monsoon trough to the foothills of the Himalayas, thus leading to the break in the monsoon when the monsoon has already remained active over central and northern India for 10-15 days and it is time for intra-seasonal reversal of the monsoon activity in terms of rainfall as the ridge phase of the ISO is approached over the northern Bay of Bengal (Ding and Sikka 2006).

In the East Asian monsoon region, the similar active-break cycle of the summer monsoon is climatologically observed. After the pre-summer rainy season in May and early June in South China and the Meiyu-Baiu rainy season from the early June to mid-July in the Yangtze River basin and Japan are terminated, breaks of the monsoon rainy period occur, respectively. Breaks of different spans are observed in South China, Taiwan, central East China, Northeast China and Korea (Chen et al. 2004; Wang and Ding 2006). Therefore, the monsoon rainfall variation during the warm season in East Asia is generally characterized by two active rainfall periods separated by a break spell. The northward passage of the Meiyu-Baiu rain band is followed by a break spell (monsoon break) which also propagates northward. Then the monsoon rainfall revival after the break is clearly observed. Chen et al. (2004) has shown that the monsoon revival in East Asia is caused by different mechanism associated with the development of other monsoon circulation components including the ITCZ, typhoons, and weather systems in mid-latitudes. Therefore, the seasonal variations with the dual peak in South China and Southwest China and the triple peak in the Yangtze River basin can be observed. In the Yangtze River basin, in addition to two peaks associated with northward advance and southward retreat of the summer monsoon, respectively, the first peak occurring before the onset of the summer monsoon in East Asia (the 25th pentad) is produced by frontal systems (the spring rainy season). It seems that the active-break cycle of the monsoon is mainly observed in the region to the south of the Yangtze River where the rainfall events are greatly affected by tropical and subtropical summer monsoon. Farther northward, only single peak of rainfalls is observed in North China, implying that the active-break cycle is not very marked. For Changma rainy season in Korea, the peak rainfall occurs in the period of late June-mid-July. Afterwards, a short break is generally observed in the late July. Starting from mid-and late August, the revival of monsoon rainy period is also observed

(Chen et al.2004). This second rain spell is not long and it maintains until early September, forming the autumn rainy season in Korea (Qian et al. 2002). In Japan, the autumn rainy season or the Shurin season (Akisama) generally takes place during the period of September 6-26 after the Baiu season (e.g., Matsumoto 1988). Previously, the Shurin frontal zone is regarded as the polar frontal zone with strong temperature gradient in both Japan and China, but according to the later study made by Matsumoto (1988), the Shurin frontal zone is characterized by a weak thermal gradient (~4℃/1000㎞) and a strong moisture gradient (~5g/㎏/1000㎞) to the west of 130°E along the southern coast of Japan like the Baiu frontal zone, with little sunshine observed (Inoue and Matsumoto 2003). However, the stagnating nature of the Shurin frontal zone was less pronounced than the Baiu frontal zone. During the Shurin season, the mid-latitude westerlies show a blocking pattern over Eastern Eurasia. In China, the autumn rainy season begins in early September and ends in mid-October, with pronounced occurrence over western China and the Yangtze River delta (Kao and Kuo 1958). This rainy season causes the second and third peaks of precipitation in southwest China and the Yangtze River basin, respectively. The end of the autumn season, accompanied by the southward shift of the subtropical jet stream over the Tibetan Plateau, is nearly simultaneous with the end of the Indian summer monsoon (Yeh et al. 1958).

The retreat of the Asian summer monsoon has been earliest observed in East Asia, which generally starts from the 44th pentad (6-10 August). The process of the retreat is very rapid, taking only one month or even less to retreat from northern to southern China. Two pentads later, the low-level southwesterlies disappear in the region to the south of the Yangtze River basin. Early in September, the leading zone of the summer monsoon quickly withdraws southward to the northern part of the South China Sea and then is stationary there, making the end of the summer monsoon in East Asia. Matsumoto (1997) also indicated that the monsoon westerlies are already replaced by easterlies in the northern part of the South China Sea in early September, while monsoonal westerlies are still dominant over the Indochina Peninsula, until late October when the summer in East Asia fully ends over Southeast Asia. So, the life cycle of the activity of the East Asian summer monsoon is about four months from May to early September. Pakistan western border makes the western boundary of summer monsoon and withdrawal starts in the first weeks of September. In almost a decade monsoon rolls back to the northwestern India which is generally associated whith a quick shift of westrlies to the south of mid-latitudes (Rasul et at 2005) The retreat of the Indian summer monsoon begins in the western parts of the Northern Indian state of Rajasthan in early September (Gadgil and Kumar 2006). Thus, the effective duration of the southwest monsoon rains over this area is only about one and a half month. The southward retreat of the summer monsoon rains continues rapidly until about the middle of October, by which time it withdraws completely for the northern half of the Indian Peninsula. During the period of October and December, major parts of South Asia is generally dry, only with the south and southwest peninsular of Indian and Sri Lamka receiving significant amounts of rainfall. By the end of December when the northeast monsoon blows with its full strength over the northern Indian Ocean, the rainy season has practically ended (Matsumoto 1990; Ding and Sikka 2006). In the WNPSM region, the summer monsoon retreats latest, generally in the end of October. Therefore, the summer monsoon season has a life cycle of about five months from June to October in this region (Li et al. 2005; Wang et al. 2005c).

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