4. SUMMER MONSOONS IN EAST ASIA, INDOCHINA AND THE WESTERN NORTH PACIFIC

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As essential parts of the Asian-Australian monsoon system, monsoon variabilities in East Asia, Indochina, and the western North Pacific are interconnected on a range of time scales. This paper provides a review for some aspects of multi-scale variability of three regional monsoon systems. The diurnal, sub-seasonal, and annual cycle of monsoon precipitation across three regions based on new observational findings are described. Progresses in observational studies of the interannual, interdecadal, and long-term variabilities of monsoon are summarized. The relations of tropical ocean–atmosphere interaction, water vapor transport, and high latitude teleconnections to the interannual variability of monsoon are addressed. How to understand the interdecadal/long-term variability of regional monsoon from three-dimensional structure and global monsoon are discussed. The hypothesized mechanisms are commented, including some results from climate model simulations. A short review on the intra-seasonal variations in the tropical regions is also included. Finally a summary is provided on possible areas of future work. The need for both observational analysis and climate modeling from the perspective of interconnected monsoon systems is emphasized.

1. Introduction

The combination of thermal contrasts between the Eurasian Continent and the Indo-Pacific Ocean produces a powerful Asian-Australian monsoon (A-AM) system (Wang 2006). The monsoons in East Asia and the western North Pacific regions are essential parts of the integral A-AM system (Wang et al. 2005a). The East Asian monsoon lies downstream of the Tibetan Plateau. The distinctive topography and orography produce features unique to the East Asian summer monsoon. The generally recognized East Asian summer monsoon (EASM) system consists of the East Asian subtropical front, the western North Pacific subtropical high and the western North Pacific monsoon trough, also known as the western North Pacific intertropical convergence zone (ITCZ) (Ding et al. 2005). While the EASM is...
mainly a continental monsoon, the western North Pacific summer monsoon (WNPSM) is an oceanic monsoon system (Li and Wang 2005; Li et al. 2005b). The prominent circulation features of the WNPSM that affect the summer climates in marine East Asia (specifically, the region comprised by the South China Sea (SCS), Philippine Sea and surrounding domains) and the western North Pacific are the monsoon trough and the subtropical anticyclonic ridge.

The EASM and WNPSM exhibit variability on a variety of time scales, including synoptic, intra-seasonal, interannual and interdecadal time scales. Although monsoon variability in the immense A-AM domain is highly seasonally dependent and varies remarkably by region, there is evidence that monsoon variability in East Asia, Indochina, and the western North Pacific is interconnected on several time scales. For example, the precipitation and circulation anomalies in these regions exhibit a major 2-year to 3-year spectral peak (Yang and Lau 2006). Recent work has revealed that since the late 1970s, the overall coupling between the A-AM system and the El Niño-Southern Oscillation (ENSO) has strengthened significantly. The relationships between the ENSO, EASM and the WNPSM have also strengthened during the ENSO’s developing, mature and decaying phases (Wang et al. 2008a). Given these linkages, it is reasonable to consider studies of the East Asian, Indochinese, and western North Pacific summer monsoons together. There have been many studies on the numerous atmospheric and oceanic features associated with these monsoon components. The principal goal of this review is to provide a synopsis of the major developments in studies of the EASM, Indochinese monsoon, and WNPSM in the last 5 years, focusing on the climatic aspects of the monsoon components.

2. Diurnal and Annual Cycles

2.1. East Asian Monsoon

Diurnal variations in precipitation are useful observational metrics for evaluating model physics. Diurnal variations in summer (June-August, hereafter JJA) precipitation over eastern China were analyzed by Yu et al. (2007a) based on rain gauge records. The spatial distributions of the phase and amplitude (normalized by the daily mean) of the diurnal cycle of JJA precipitation are shown in Fig. 1. The prevailing nocturnal maximum precipitation over the eastern part of the Tibetan Plateau and Sichuan Basin (east to 100°E along 30°N) is evident. Inland southern China (23°N-26°N, 110°E-117°E) and northeastern China (40°N-50°N, 110°E-130°E) have late afternoon rainfall maxima. The diurnal cycle of summer precipitation over central eastern China between the Yangtze River and the Yellow River valley (30°N-40°N, 110°E-120°E) is characterized by two comparable peaks. Yu et al. (2007b) discussed the relation between rainfall duration and diurnal variation during the warm season (May to September) over central eastern China (26°N-36°N, 105°E-120°E). It was found that long-duration rainfall events (events lasting longer than 6 hours) tended to reach their maximum hourly rainfall in the early morning, while short-duration rainfall events (events lasting 1-3 hours) reached their peak rainfall in the late afternoon. While seasonal variations in the diurnal phases of precipitation in southwestern China were weak, the rainfall
peak in southeastern contiguous China shifted sharply from late afternoon in the warm seasons to early morning in cold seasons (Li et al. 2008a).

High spatial and temporal resolution data are required for studying the precipitation characteristics over East Asia in order to resolve the complex terrain. Zhou et al. (2008a) compared the summer precipitation frequency, intensity, and diurnal cycles derived from Precipitation Estimation from Remote Sensing Information using Artificial Neural Network (PERSIANN) and Tropical Rainfall Measuring Mission (TRMM) 3B42 satellite data and rain gauge records, respectively. The satellite data overestimated rainfall frequency but underestimated its intensity. Both frequency and intensity contributed to diurnal rainfall variations over most of eastern China. The contribution of frequency to the diurnal cycle of rainfall quantity was generally overestimated using satellite data. Both satellite data sets effectively capture the nocturnal peak over the eastern periphery of the Tibetan Plateau, as well as the late afternoon peaks in southern China and northeastern China, but fail to capture
the observed secondary early morning peak over the region between the Yangtze and Yellow Rivers, partly due to the dominance of warm front clouds or shallow convection over this region.

The seasonal reversal of wind direction is generally regarded as a common and essential characteristic of monsoons. Zhang and Li (2008) introduced a new concept, the directed angle, to study the seasonal variation of monsoons and found six categories of wind direction rotation with seasonal cycles in the global rotation regimes of the wind direction. The directed angle describes the evolution of wind vectors on a daily scale, which in turn provides spatio-temporal information about wind vector variation and facilitates model evaluation on the basis of monsoon rotation dynamics. For example, the South Asian monsoon follows a “counter-clockwise to clockwise” rotation, while the East Asian monsoon follows a “full counter-clockwise” rotation. A “clockwise to counter-clockwise” rotation is seen in the SCS. The South Indochinese Peninsula is covered by “full clockwise” rotation. To describe the abrupt seasonal variations in wind direction and circulation of the Asian monsoon, Li and Zhang (2009) proposed monsoon parameter “onsets”, including circulation onset (wind onset), reversal of zonal wind, reversal of meridional wind, outgoing longwave radiation (OLR) onset, rainy season onset, etc. They show evidence that wind onset is a good predictor of monsoon and rainy season onset in most monsoon regions. The angle amplitudes of wind vectors during wind onset and withdrawal have distinct regional differences in Asian monsoon regions.

2.2. Monsoon in Indochina

Since monsoonal wind reversal is pronounced, Chang et al. (2005) showed that there is a sharp contrast of main rainy season in Southeast Asia between east and west coastal regions. Xie et al. (2006) revealed that narrow mountain ranges are important for anchoring monsoon convection centers on the windward side. For example, the heaviest precipitation in the Bay of Bengal is located on its eastern coast. This is in contrast to the widely held view that this convection is centered over the open ocean, as is often shown in coarse-resolution datasets. Hirose et al. (2008) demonstrated the orography-related fine structure of rainfall frequency in the Asian and Australian sectors. As for the effect of elevation in mountains in northern Thailand, Kuraji et al. (2001, 2009) demonstrated that rainfall quantity increases linearly with elevation by utilizing their dense rain gauge network installed in the mountainous region. Dairaku et al. (2004) attributed the larger amount of high-altitude rainfall to increased duration and frequency, rather than increased rainfall intensity.

Oki and Musiake (1994) described seasonal changes in the diurnal cycle over Malaysia. Nitta and Sekine (1994) revealed the diurnal cycle of convective activity in the western Pacific and found conspicuous differences in the peak time of convective activity between the land and ocean. As for the summer monsoon season rainfall, Ohsawa et al. (2001) analyzed both satellite-derived cloud data and rainfall data from weather stations in Bangladesh, Thailand, Malaysia and Vietnam. They presented minute regional aspects of the diurnal variations in convective clouds and rainfall in the Indochinese Peninsula (Fig. 2). Although overall a late afternoon peak is predominant as in South China, a morning peak is observed in
some specific inland regions implying strong effect of orography on diurnal cycle. Interestingly, the climatological rainfall amount in these regions is more abundant. However, the physical mechanism of morning rainfall in these regions is still not very clear. It is noted that the peak of maximum convective activity seems to be delayed from west to east in some portion of the peninsula. Satomura (2000) succeeded in simulating eastward migrating diurnally varying rainfall systems in the Indochinese Peninsula using a regional model, and Okumura et al. (2003) confirmed the occurrence of an eastward migrating rainfall system in northern Thailand using radar data. Hirose and Nakamura (2005) revealed the detailed distribution of diurnal cycles using TRMM data.

Recent studies have suggested that the earliest onset of heavy rainfall occurs over the Indochinese Peninsula (Matsumoto 1997; Wu and Wang 2000). Heating over the northern part of Indochina is an important factor in the onset of the Southeast Asian summer monsoon, and following the commencement of the Indochinese monsoon, this zone of heating acts as a circulation center, thus inhibiting the onset of the Indian monsoon (Hsu et al. 1999).
Kawamura et al. (2002) stressed the importance of land–sea interactions in promoting monsoon onset. Continental heating during the pre-monsoon season produces low-level circulation between the continent and the adjacent ocean, which is a necessary precondition for sudden monsoon onset in both Australia and India. In the case of the Indochinese monsoon, this process is weak, and the role of the ocean is more important (Minoera et al. 2003). Numerical model simulations suggest that warming over the western North Pacific is crucial for the onset of the Indochinese monsoon, while continental warming is important for the onset of the Indian monsoon (Kanae et al. 2002). Both the warming over the Tibetan Plateau (Li and Yanai 1996; Ueda and Yasunari 1998) and the latent heat released over the Bay of Bengal (Xu and Chan 2001) may trigger the onset of the Indochinese monsoon. Zhang et al. (2004) demonstrated the occurrence of early reversals of the meridional temperature gradient throughout the entire troposphere and the corresponding establishment of an easterly vertical wind shear. The El Niño event has also been shown to influence the inter-annual variations in monsoon onset (Zhang et al. 2002).

Pre-monsoon rains occur in April and early May in Indochina, in stark contrast to the pre-monsoon conditions in the Indian subcontinent (Matsumoto 1997). The majority of moisture in pre-monsoon Indochina comes from the east, as compared with the westerly flows characteristic of the summer monsoon (Kiguchi and Matsumoto 2005). Ishizaki and Ueda (2006) also revealed the seasonal evolution of heat sources and moisture sinks before and during the monsoon onset over and around the Indochinese Peninsula.

Following the onset of the summer monsoon, a so-called “break” in monsoon rainfall occurs over a wide region in Indochina (Matsumoto 1997). Takahashi and Yasunari (2006) also showed a distinct climatological monsoon break occurring over Thailand in late June. The occurrence of this break coincides with drastic changes in large-scale monsoon circulation in March. In the latter half of the monsoon, during August and September, westward-propagating tropical cyclones frequently cross over the Indochinese Peninsula (Fudeyasu et al. 2006), causing a secondary rainfall peak. In general, this season is the annual rainfall peak season in Indochina (Wang and Lin 2002). Another peculiar feature of the seasonal changes in Indochina is the existence of a strong inversion layer during the winter monsoon season (Nodzu et al. 2006). Even after the peak summer monsoon season, heavy rainfalls sometimes occur in the east coastal region of Indochina. The coexistence of cold surge and tropical depression-type disturbance is important for the heavy rainfall in central Vietnam (Yokoi and Matsumoto 2008).

2.3. Western North Pacific Summer Monsoon

Murakami and Matsumoto (1994) noted several differences between the WNPSM and the South Asian summer monsoon (SASM). The latter is characterized by strong land–sea contrasts and high-rising terrain in the north. Both factors serve as strong forcing elements which initiate and maintain the monsoon. By contrast, the WNPSM occurs over the open ocean, where meridional sea surface temperature (SST) and pressure gradients are relatively small, and yet the convection between 5°N and 20°N is as active as that in the SASM. Despite occurring over warm ocean regions, the strong underlying forcing occurring in the
South Asian summer monsoon is absent in the western North Pacific. Lack of strong forcing may be one of the reasons why the onset of the WNPSM occurs in mid-summer, much later than monsoon onsets in Southeast Asia, East Asia, and South Asia (e.g., Wu and Wang 2000; Wang and Lin 2002; Lin and Wang 2002).

The large-scale circulations and convection patterns in marine East Asia and the western North Pacific exhibit significant sub-seasonal variations. In June, the monsoon trough is weak and convection in the SCS and Philippine Sea is located primarily south of 15°N. A convection belt associated with the Meiyu frontal activity is located over the Yangtze River valley and Japan. Between the north and south convection belts is the subtropical anticyclonic ridge at approximately 20°N. In August, the subtropical anticyclonic ridge shifts to approximately 35°N and the monsoon trough and western North Pacific convection belts have reached their mature stages. This shift between June and August patterns tends to occur over a relatively short period in July when the WNPSM onset occurs. This onset is characterized by the sudden northward shift of large-scale convective activity in the western North Pacific in late July (Ueda et al. 1995). This phenomenon is accompanied by the abrupt monsoon trough strengthening and the northward shift of the subtropical anticyclonic ridge from 20°N to 35°N. Ueda et al. (1995) have suggested that the preceding SST warming in the Philippine Sea, beginning in early July, is an important ingredient for the abrupt northward shift of these convection belts. The involvement of local atmosphere-ocean interactions in the monsoon development process was further explored by Ueda and Yasunari (1996) and Wu (2002).

3. Variability of the East Asian Summer Monsoon

3.1. Interannual Variability

While the monsoon-ENSO relationship is still an active topic of research (see Wang et al. 2005a for a review), recent efforts have also been devoted to the analysis of water vapor budgets and teleconnections with mid-to-high latitude processes and westerly jets.

The influence of the ENSO on the EASM is different at different stages of the ENSO cycle. The ENSO plays a major role in the summers following the mature phases of El Niño events (Zhang et al. 1999; Gong and Wang 1999; Lu 2005). The Yangtze River valley is wet in the pre-Meiyu/Meiyu season after the ENSO’s peak phase (Chang et al. 2000). Wang et al. (2000) have proposed a mechanism to explain this lagging influence. During the boreal winter of each El Niño event’s mature phase, an anomalous anticyclone is established over the western North Pacific, which is maintained until the subsequent early summer through local positive wind-evaporation-SST feedback. In the succeeding year after the onset of the ENSO, the EASM tends to be weaker, and the western Pacific subtropical high (WPSH) tends to extend further southward which leads to more rainfall along the Yangtze-Huaihe River valley (Zhang et al. 1996, 2002). The anomalous anticyclone in the off-equatorial western Pacific has also been identified to play an important role for ENSO’s oscillation (e.g., Weisberg and Wang 1997; Wang et al. 1999). The impact of the ENSO on monsoon rainfall is
intensity-dependent. In moderate ENSO years, there is excessive rainfall along the Yellow-Huaihe River valleys but deficient rainfall along the Yangtze River valley (Xue and Liu 2008). This pattern is far different from the classical ENSO-related rainfall anomalies (Zhang et al. 2002). The response of East Asian climate to El Niño and La Niña is highly asymmetric (Chen et al. 2008) and nonlinear (Liu et al. 2008). The EASM exhibits strong quasi-biennial oscillations. The biennial oscillations may result from both the atmosphere–ocean interaction in the warm pool (Li et al. 2006b) and the El Niño turnabout, the latter of which leads to a significant change in summer western Pacific circulation anomalies (Wang et al. 2003; Li et al. 2005a). Remote eastern Pacific forcing affects the interannual variability of EASM in conjunction with local SST feedbacks (Sui et al. 2007; Wu and Zhou 2008). Although the off-equatorial SST anomalies in the western Pacific are smaller than those in the equatorial eastern Pacific, they can produce atmospheric response of comparable magnitude because the atmospheric mean state is convergent in the western Pacific and divergent in the equatorial eastern Pacific (Wang 2000). The atmospheric response induced by the off-equatorial western Pacific pattern has a connection with climate variability in the Indian Ocean and East Asia (Chen et al. 1985; Xie et al. 2009; Wu, B. et al. 2009a).

The significant correlation between East Asian monsoon and SST over the tropical Indian Ocean has long been documented (Chen et al. 1985; Hu 1997; Gong and Ho 2002). Numerical experiments have also demonstrated the dominant influence of the tropical Indian Ocean SST on the heavy rainfall along the Yangtze River valley (Guo et al. 2004). However, the means by which the Indian Ocean exerts its influence on the EASM is unclear. This influence may be attributed to the involvement of the ENSO signals in the formation of Indian Ocean SST anomalies. There exist two types of leading modes of SST variability in the Indian Ocean: the basin-scale warm/cold mode (IOBM) and the longitudinal dipolar mode (IOD). The IOBM may play a role in the excessive rainfall over the middle and lower reaches of the Yangtze River following El Niño events such as those occurring in 1983 and 1998 (Yang et al. 2007; Li, S. et al. 2008; Xie et al. 2009; Wu, B. et al. 2009a). The decadal variability of EASM is also related to IOBM (Zhou et al. 2009c). The Asian monsoon interacts with IOD (e.g., Li and Mu 2001; Li et al. 2003).

Air–sea coupling also plays a dominant role in Asian monsoon variability, especially for the prediction of summer monsoon rainfall (Wang et al. 2005b). The observational analysis shows that the rainfall–SST relation over the western North Pacific monsoon region experiences a significant interannual variation (Wu, B. et al. 2009b). Analysis of the performance of 11 Atmospheric General Circulation Models (AGCMs) involved in the 2nd phase of Atmospheric Model Inter-comparison Project (AMIP) demonstrated the dominance of remote El Niño forcing in producing the predictable portion of monsoon rainfall variability; however, the AMIP run has deficiencies in simulating the seasonal phase of two anticyclones (specifically, the western North Pacific anticyclone and the South Indian Ocean anticyclone) associated with the first mode of monsoon rainfall, which are not in phase with ENSO forcing in observations, but strictly match that of Nino 3.4 SST in the AMIP run (Zhou et al. 2009a). In contrast, air–sea coupled models generally show reasonable performance with respect to reproducing the seasonal phase of two anticyclones (Wang et al. 2007), reinforcing the
importance of local air–sea coupling effects. Recent results from the Climate Variability and Predictability (CLIVAR) C20C project have shown that among the Asian–Australian monsoon subsystems, the interannual variability of the EASM has the lowest reproducibility in AGCM runs forced by historical SST (Zhou et al. 2008b).

Water vapor transport feeds the monsoon rainfall. Traditionally the Somali jet, the cross-equator flow associated with the Australian cold high and the easterly flow from the south of the western Pacific subtropical high are regarded as the key moisture sources for heavy rainfall over eastern China (Tao and Chen 1987). Ding and Sun (2001) emphasize the moisture transport from the Bay of Bengal, the SCS and the West Pacific. The water vapor transport from the Indian summer monsoon is in inverse proportion to that over East Asia (Zhang 2001). The origins of water vapor supply related to anomalous rainfall patterns are different from those related to the normal monsoon rainfall. A heavier rainbelt along the middle and lower reaches of the Yangtze River valley follows a convergence of tropical southwesterly water vapor transport with midlatitude northeasterly transport. This tropical water vapor transport originates over the Philippine Sea, then passes through the Bay of Bengal and the SCS before reaching the Yangtze basin (Zhou and Yu 2005).

The interannual EASM variability can be described in terms of several different modes (Wu et al. 2008a), aspects of which are linked to mid- and high latitude disturbances of circulation. There is a significant correlation of the EASM with the late spring Arctic oscillation (AO) on an interannual time scale. Following a strong spring AO, the East Asian jet tends to move northward in the summer, as does the rainbelt, and a downwelling phenomenon appears along the Yangtze River valley (Gong and Ho 2003). The association between spring Arctic sea ice concentration and summer rainfall in China has been reviewed by Wu, B. Y. et al. (2009a, b), and a Eurasian teleconnection pattern has been revealed. Wu et al. (2008b) found a dipole anomaly in the Arctic atmosphere, which may play a more important role than the AO or North Atlantic Oscillation (NAO) in influencing summer rainfall over China. The extreme cold anomalies over East Asia during January-February 2008 were attributed to the combined effect of Madden-Julian Oscillation (MJO), La Niña and high-latitude blocking high (Hong and Li 2009). Recent analysis also suggests a teleconnection of the EASM with the southern hemisphere circulation (Gao et al. 2003; Nan and Li 2003). An intensified Antarctic oscillation (AAO) is followed by an enhanced Mascarene high and Australian high, as well as by stronger cross-equatorial flow (Xue et al. 2004; Xue and He 2005). Nan et al. (2009) have suggested that the Indian Ocean SST plays a role in bridging the boreal spring SAM (Southern Annular Mode) and the EASM. Given the significant connection between AAO variability and ENSO occurrence (Zhou and Yu 2004), further study on the extent to which the reported EASM-AAO correlation is ENSO-dependent is warranted.

The East Asian subtropical westerly jet (EASWJ) plays an important role in East Asian climate change. On the seasonal timescale, the westerly jet axis experiences meridional migration from winter to summer, an event which is closely tied to the monsoon climate in East Asia (Liang and Wang 1998; Zhang and Guo 2005; Zhang et al. 2006). On interannual time scales, the meridional shift of the EASWJ is also associated with shifts in the rain belt
Tianjun Zhou et al. (Zhou and Yu 2005; Lin and Lu 2005). The center of the westerly jet over mid-latitudes is located at 200 hPa throughout the year. The seasonal evolution of the EASWJ experiences both meridional and longitudinal changes, and the latter is related to the land–sea thermal contrast in the longitudinal direction (Zhang et al. 2006).

3.2. Interdecadal Variability

The East Asian climate experienced an interdecadal transition in the late 1970s. A surface cooling occurred in some regions of East Asia, contrasting with the warming trends commonly observed elsewhere on earth (Li et al. 1995; Hu et al. 2003; Xu et al. 2006). At the same time, precipitation increased over the middle and lower reaches of the Yangtze River valley, whereas it decreased over the middle and lower reaches of the Yellow River valley (Hu 1997). This rainfall pattern is often referred to as a “southern-flood-and-northern-drought” (hereafter SFND) pattern. The mechanism responsible for this interdecadal transition has been an active topic of research. Theories range from natural variability, to global warming and the introduction of anthropogenic forcing agents (see Ding et al. 2007; Zhou et al. 2009b for comprehensive reviews). Analysis of reconstructed Chinese summer rainfall records spanning the last 500 years reveal strong cyclical variability over an 80-year time period (Wang and Li 2007). The effect of this 60-80-year oscillation on the summer precipitation in China is described by Ding et al. (2008). The increased emission of anthropogenic aerosols might play a role in recent trends (Qian et al. 2001; Menon et al. 2002), although the response of climate models to such aerosol forcing is highly variable and model-dependent (Li et al. 2007; Li et al. 2010). While still there is no consensus on the relative contributions of natural and anthropogenic forcing to changes in monsoon events, recent studies have found that the weakening of summer monsoons over East Asia is evident in three dimensional (3-D) circulation system (Yu and Zhou 2007). This bigger picture exhibits how the 3-D structure of East Asian atmospheric cooling contributes to surface monsoon circulation change.

Surface summer cooling in south-central China is associated with a large-scale lower tropospheric pressure rise in East Asia, which directly prevents the northward advance of the EASM (Yu et al. 2004a). As shown in Fig. 3, East Asia was dominated by a distinctive strong tropospheric cooling phenomenon during July and August. This cooling trend is most prominent in the upper troposphere around 300 hPa. Accompanying this summer cooling, the upper-level westerly jet stream over East Asia shifts southward and the EASM weakens, resulting in a tendency toward increased droughts in northern China and flooding in the Yangtze River valley.

Similar coherent changes between the surface and upper troposphere also exist in spring. South China (26°N-31°N, 110°E-122°E) has undergone a significant decrease in late spring (April-May) precipitation since the late 1970s. The reduction of precipitation coincides with a cooling in the upper troposphere over central China (30°N-40°N, 110°E-125°E). The upper-level cooling is associated with an anomalous meridional cell characterized by descending motion in the latitudes of 26°N-35°N and low-level northerly winds over southeastern China (22°N-30°N, 110°E-125°E), causing deficient rainfall over southern China (Xin et al. 2006).
Yu and Zhou (2007) developed an integral picture of the 3-D structure of East Asian climate change. The upper tropospheric cooling shows strong seasonal variation. In the troposphere, the strongest cooling is located at approximately 300 hPa and persists throughout the year except in June. There are two peaks in the upper tropospheric cooling occurring in April and August. In June, the upper tropospheric cooling is replaced by a
moderate warming trend, and the strongest cooling occurs in the lower stratosphere. Cooling-induced mass changes result in positive (negative) anomalies of the geo-potential height (GPH) beneath (above) the cooling. The surface climate changes are closely related to atmospheric cooling and the associated circulation changes. Tropospheric temperature changes dominate the changes in westerly currents, and the zonal jet change largely determines the shifts in the surface rainbelt. A schematic diagram is used to show the relationship between upper tropospheric cooling and rainfall changes (Fig. 4).

Figure 4. Schematic diagram showing the effect of upper tropospheric cooling on atmospheric circulation and precipitation. The arrows denote anomalous winds. The dashed (solid) circle in the lower layer represents the region of dry (wet) conditions. The letter “C” in the upper layer denotes a cyclone; the letter “A” in the lower layer denotes an anticyclone (Xin et al. 2006).

As an extension of Yu et al. (2004b), Zhou and Zhang (2009) developed an integral view of the inter-decadal variability of tropospheric temperatures across the entire subtropical Northern Hemisphere during July and August. A major mode was identified, with one significant cooling center over East Asia and two warming centers over the North Atlantic and North Pacific Oceans, respectively. This dominant mode exhibits a decreasing trend over the entire period examined, particularly before 1980. After the mid-1980s, the trend begins to level off. Thus, the weakening trend in the EASM may be a local manifestation of Northern Hemisphere interdecadal-scale climate transitions and may not be explained simply in terms of local anthropogenic aerosol changes.

3.3. Mechanisms of Long-term Variability

Explanations for the interdecadal tropospheric cooling trend over East Asia remain elusive. For the springtime cooling, there is evidence suggesting a teleconnection with the North Atlantic Oscillation (NAO). NAO-related signals barotropically extend eastward over most of subtropical Eurasia and reach eastern China in March (Yu and Zhou 2004). During winters with a positive NAO phase, upper tropospheric cooling occurs first in the northern Tibetan Plateau in early-middle spring, then propagates southeastward to central China in late spring. Hence the interdecadal changes in the winter NAO is regarded as one factor responsible for
late spring droughts over southern China (Xin et al. 2006). Li, J. et al. (2005) found a strong cooling shift occurred in early spring (March and April) and late summer (July, August and September), in contrast to the warming shift observed in other seasons over the eastern flank of the Tibetan Plateau. This springtime cooling is related to the NAO. The mid-tropospheric westerlies also tend to intensify during positive NAO phases. The enhanced westerlies result in strengthened ascending motion against the leeside of the plateau, which favors the formation of mid-level stratiform clouds. The increased prevalence of stratus clouds induces negative net cloud radiative forcing, thereby cooling the surface air and triggering a positive cloud-temperature feedback loop. Continental stratiform cloud-climate feedback plays a significant role in amplifying the cooling shifts observed downstream of the Tibetan Plateau (Yu et al. 2004b). In addition, Li et al. (2008b) have revealed another teleconnection pattern, NAULEA (North Atlantic–Ural–East Asia), which links climate changes over the North Atlantic and Eurasia.

In addition to upper tropospheric temperature change, the interdecadal changes in tropical SST and tropical Pacific convection observed since the late 1970s have been regarded as another driving mechanism. Diagnostic studies revealed significant negative correlations of the EASM index with SST anomalies in the Indo-Pacific warm pool and the northwestern Pacific between 30°N-50°N (Li and Zeng 2002, 2005). Associated with the SFND rainfall change, the WPSH has intensified and extended westward (Nitta and Hu 1996). The change in SST over the tropical Indian Ocean and northwestern Pacific from 1976-1979 partly accounts for this interdecadal change (Hu 1997; Gong and Ho 2002). Zhou et al. (2009c) examined the responses of five AGCMs to specified identical Indian Ocean-western Pacific (IWP) warming. The specified IWP warming has led to a westward extension of the WPSH in all five AGCMs (Fig. 5). Analysis of model results showed that the negative heating in the central and eastern tropical Pacific and increased convective heating in the equatorial Indian Ocean/Maritime Continent associated with IWP warming are conducive to the westward extension of the WPSH in the decades since the late 1970s.

The SFND pattern is a local manifestation of global land monsoon change. Wang and Ding (2006) found an overall weakening of global land monsoon precipitation over the last 56 years, primarily due to weakening of summer monsoon rainfall in the Northern Hemisphere, in particular the African and Asian monsoon (Zhou et al. 2008c). Based on a set of AGCM simulations forced by historical SSTs, Zhou et al. (2008d) suggest that the decreasing tendency of global land monsoon rainfall was mainly caused by the warming trend over the central-eastern Pacific and the western tropical Indian Ocean. However, due either to model bias in monsoon rainfall simulations (Zhou and Li 2002) or to the neglect of air–sea coupling in stand-alone AGCM runs (Wang et al. 2005b), the skill of monsoon precipitation change over East Asia is low (Zhou et al. 2008d). Li et al. (2010) analyzed the ensemble simulations of two AGCMs forced by observed SSTs. The results demonstrated that the recent warming of tropical oceans has played a significant role in the weakening of EASM circulation during recent decades, although the two AGCMs still failed to reproduce the regional rainfall changes over eastern China.

The impact of global warming on the East Asian summer monsoon has been
controversial. The tremendous uncertainties among the models used in precipitation simulations make it difficult to link precipitation variations to global warming (Hu et al. 2003). Kripalani et al. (2007) analyzed the coupled model simulations and projections under Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4). They only found a significant increase over Korea and Japan and the adjoining north China region. Since precipitation is the most difficult variable for climate simulations to capture, current state-of-the-art climate models may not be able to correctly simulate rainfall changes; a study of atmospheric circulation variations is thus an essential pre-requisite for understanding rainfall variations. Zhou and Yu (2006) examined variations in the surface air temperature (SAT) simulated by nineteen coupled climate models driven by historical natural and anthropogenic forcings under the phase 3 of the Coupled Model Intercomparison Project (CMIP3) for IPCC AR4. Most models performed well in simulations of both the global and the Northern Hemispheric mean SAT evolutions. However, there were discrepancies between the simulated and observed regional features of SAT trends over China. Few models could produce the summertime cooling over the middle part of eastern China (27°N-36°N). These deficiencies in SAT simulation led to a poor reproduction of the land-sea thermal contrast, and hence of the monsoon circulation change.

Figure 5. The positions of a characteristic western Pacific subtropical high isoline at 500 hPa in Indo-western Pacific warming (red), cooling (blue) experiments and control runs (long-dashed black line). The name of the AGCM is marked in the left-hand corner of each panel. The model results are for 30-year means. The condition of NCEP/NCAR reanalysis is shown in panel (a) (red line for 1980-1999 mean, blue line for 1958-1979, and long-dashed black line for 1958-1999). (Zhou et al. 2009c)
The surface warming trends observed for the Tibetan Plateau (TP) have been proposed as a mechanism for EASM change. The linear rate of annual mean temperature increase over the TP during the period 1955-1996 is approximately $0.16^\circ$C/decade (Liu and Chen, 2000). The warming trend can be reproduced by coupled climate models with specified 20th century radiative forcing agents, suggesting that recent warming over the TP primarily results from increasing anthropogenic greenhouse gas emissions (Duan et al. 2006; Duan and Wu 2006). Numerical experiments using climate models show that atmospheric heating induced by rising TP temperatures can enhance East Asian subtropical frontal rainfall (Wang et al. 2008b). Partly due to mean state bias of the model, the specified TP warming led to a strengthened low-level southwest monsoon flow, not a weakened southwesterly flow as found in observations. In addition, the anomalous snowfall over the TP is also suggested to have impacts on monsoon variability (Wu and Qian 2003; Zhang et al. 2006; Zhao et al. 2007), but the mechanisms fall into a subject of debate.

4. Variability of Summer Monsoons in Indochina

4.1. Intraseasonal Variations

In South and Southeast Asia, intraseasonal variations (ISVs) of precipitation and tropospheric circulation are important features during the rainy season. The ISVs can be divided into two types according to their time scales: one has a 30-60-day variation (hereafter 30-60 DV) and the other a 10-20-day variation (hereafter 10-20 DV). The 10-20 DV of convection and accompanying cyclonic circulation in the lower troposphere originates over the SCS or the western Pacific Ocean. It propagates westward and produces rainfall over the Indochinese Peninsula (Yokoi and Satomura 2005). Yokoi et al. (2007) found that during the rainy season, variance of the 30-60 DV is generally larger in coastal regions than over inland regions, and it has two local maxima: one found in the Bay of Bengal and coastal Myanmar and the other in southern Laos and central Vietnam. On the other hand, the largest variance of the 10-20 DV is found in the coastal regions of northern and central Vietnam, while the variance in other coastal regions is generally smaller than that in inland regions. Yokoi and Satomura (2006) revealed the remarkable difference in the geographical distribution of variance between two types of intraseasonal variations in daily-mean radar reflectivity data in the western part of the Indochinese Peninsula. The 30-60 DV of reflectivity factor dominates most of the coastal region, while its variance in the inland region is small. Horizontal gradients in the variance are the largest over the mountain range, implying that the mountain range plays a significant role in this geographical contrast.

4.2. Interannual Variability and Long-term Trends

The Indochinese summer monsoon rainfall shows significant variations on both the interannual and decadal time scales. Chen and Yoon (2000) found a strong correlation between Nino-3 SST and the interannual variation of Indochinese monsoon rainfall. Lim and Kim (2007) revealed the impact of the ENSO on the space-time evolution of the regional
Asian monsoon. The rainfall in Thailand showed a remarkable decreasing trend from 1951 to 1994, which is only significant in September (Kanae et al. 2001). This decrease may be attributable to changes in land surface conditions (Kanae et al. 2001), while the decreasing tropical cyclone activity may also contribute (Takahashi and Yasunari 2008). Regional model results indicate that the deforestation of Indochina may affect the summer rainfall in East China (Sen et al. 2004).

5. Variability of WESTERN NORTH PACIFIC Summer Monsoons

5.1. Sub-seasonal Variability and Multi-scale Nature of Western North Pacific Monsoons

The western North Pacific (WNP) in the boreal summer is characterized by the multi-spatio-temporal scale circulation and convection (Holland 1995; Hsu 2005; Li and Wang 2005), e.g., intraseasonal oscillation (ISO) and tropical cyclones (TC). The well-explored 30-60 day intraseasonal oscillation is known to propagate northward and northwestward in the South China and Philippine Seas (Lau and Chan 1986; Nitta 1987; Chen et al. 1988; Wang and Rui 1990; Hsu and Weng 2001; Tsou et al. 2005). Hsu and Weng (2001) documented the evolution of the northwestward propagation of 30-60 day disturbances (Fig. 6) and proposed a convection-circulation mechanism to explain this northwestward propagation tendency. The frictional convergence near the center of the cyclonic circulation located to the northwest of the deep convection provides a favorable condition for the system to move northwestward. Another branch of interest for the ISO resides in the higher frequency (e.g., 10-25 day or 7-30 day) band of the ISO. The higher-frequency ISO, which often propagates westward in the tropical WNP, is often described as westward-propagating tropical waves (Chen and Chen 1995; Chen and Weng 1999; Fukutomi and Yasunari 1999; Straub and Kiladis 2003; Hsu 2005). The ISO has a modulating effect on higher-frequency perturbations. For example, tropical cyclones, the most energetic atmospheric phenomena in this region, preferentially occur during the convective phase of the MJO, and cluster around the cyclonic vorticity and convergence anomalies, which appear poleward and westward of the other large-scale convective anomaly along the equator (Nakazawa 1986; Liebmann et al. 1994). Maloney and Dickinson (2003) found larger synoptic (2.5-12.5 day) variance and more favorable conditions for the growth of tropical disturbances in the westerly phase of the MJO than in the easterly phase.

Other westward- or northwestward-propagating wave-like perturbations were also observed to be closely associated with TCs. Ko and Hsu (2006) identified a 7-30 day wave pattern propagating north-northwestward from the northeast of Papua New Guinea to the oceanic area between Taiwan and Japan. When the cyclonic circulation of the wave pattern moved into this area, more than 70% of the wave patterns were associated with at least one recurring TC. Ko and Hsu (2009) further demonstrated the ISO’s modulating effect on both the sub-monthly wave and the TC. The ISO in the westerly phase provided a favorable background (e.g., enhanced monsoon trough and moisture confluent zone) for the wave-TC
pattern development, while the ISO in the easterly phase provided a less favorable environment. Another similar wave-like pattern moving toward the SCS over a shorter time scale (3-8 days) was also identified by Lau and Lau (1990) and Chang et al. (1996). Straub and Kiladis (2003) suggested that TCs often develop from the high-frequency, westward-propagating mixed Rossby gravity-tropical depression type disturbances. While these wave-like patterns may modulate TCs, it has also been proposed that TCs may excite synoptic-scale waves to encourage development of another TC through Rossby wave energy dispersion (Holland 1995; Sobel and Bretherton 1999; Li and Fu 2006; Li et al. 2006a; Krouse et al. 2008).

Figure 6. Evolution of the 30-60-day OLR and low-level circulation patterns in the western North Pacific. Lag-adjusted regression coefficients between the OLR anomaly averaged at 120°E-160°E and 0°-20°N and the OLR (shaded) and 850-hPa vorticity (contoured) at (a) day -15, (b) day -10, (c) day -5, (d) day 0, (e) day 5, and (f) day 10. Contour intervals are 2 W m⁻² and 10⁻⁷ s⁻¹ for OLR and vorticity, respectively. Dark shading and solid lines indicate positive values, while light shading and dashed (and dotted) lines indicate negative values. The regression coefficients have been multiplied by one standard deviation of the OLR index, and only those that are significant at the 0.05 level are plotted. (from Hsu and Weng 2001)

In addition, TC activity is often modulated by low-frequency fluctuations such as ISO, ENSO, and even decadal fluctuation. These findings raise the question of whether clustered TCs have an ensemble effect on climate variability. Hsu et al. (2008) found that TCs
contribute a great deal (more than 50 percent in certain regions) to the intraseasonal and interannual variance of 850-hPa vorticity along TC tracks.

5.2. Interannual Variability

The WNPSM onset date exhibits more interannual variability than monsoons in other regions (Wu and Wang 2000). The standard deviation of the dates is about 5 pentads in the WNP, comparing to 2 pentads in South and Southeast Asia and 3 pentads in the SCS. The ENSO is one of the key factors leading to this large variability (Wu and Wang 2000). Following the mature phase of El Niño, the descending motion induced by anomalous deep convection in the equatorial central-eastern Pacific and enhanced by negative SST anomalies in the Philippine Sea delayed the onset of the WNPSM. Conversely, following the mature phase of La Niña, the local warm SSTs in the Philippine Sea helped to induce the monsoon trough and drive convection, leading to early monsoon onset. Recently, Akasaka (2010) found that the onset timing in the summer rainy season in the Philippines has been delayed after the latter half of the 1970s.

In terms of seasonal-mean characteristics, strong (weak) WNPSMs were characterized by positive (negative) rainfall anomalies, negative (positive) SST anomalies, warmer (colder) upper tropospheric temperatures, and low-level cyclonic (anticyclonic) circulation anomalies over the subtropical WNP (Chou et al. 2003). While these characteristics hold for all years (ENSO and non-ENSO; Fig. 7), the ENSO contributes significantly to interannual variation (Lau and Wu 2001; Wang et al. 2001; Li and Wang 2005). During the maturing El Niño (La Niña) phenomena in the boreal winter, a low-level anticyclonic (cyclonic) anomaly is often observed in the Philippine Sea as a Rossby wave response to the negative (positive) SST anomaly in the southeastern corner of the same sea (Wang et al. 2000; Hsu et al. 2001). Positive ocean-atmosphere feedback through wind-evaporation coupling may locally maintain this anomalous circulation through the spring and into the summer (Wang et al. 2000; Kawamura et al. 2001). As a result, the WNPSM tends to be weaker (stronger) during the decaying phase of El Niño (La Niña), especially following strong events (e.g., the 1997/1998 El Niño; Su et al. 2001). Chou et al. (2003) revealed an inverse relationship between the developing ENSO and WNPSM: a stronger (weaker) WNPSM was observed during developing El Niño (La Niña) events.

One of the most recurrent teleconnection patterns in the WNP is the Pacific–Japan (PJ) pattern (Nitta 1987; Huang and Sun 1992; Kosaka and Nakamura 2006), which is a wave-like circulation pattern emanating from the Philippine Sea and extending to the extratropical North Pacific, which has a notable effect on the monsoon trough and subtropical anticyclonic ridge. The pattern is often interpreted as a Rossby wave-like perturbation induced by anomalous heating over the unusually warm SST in the Philippine Sea. However, statistically significant correlations are also found between the vertically integrated heating above the Tibetan Plateau and the PJ pattern (Liu et al. 2002; Hsu and Liu 2003). While the heating anomaly over the Philippine Sea was considered the major source of forcing for the PJ pattern, Enomoto et al. (2003) proposed that the Rossby wave energy dispersing along the waveguide near the jet stream across the Eurasian Continent to East Asia also has a notable effect on the
The northward wave activity flux over the WNP, which is often associated with the SST anomaly in the Philippine Sea, is mainly confined in the lower troposphere (Kosaka and Nakamura 2006; Hsu and Lin 2007). The PJ pattern is a combination of the southward wave activity in the upper troposphere and the northward wave activity in the lower troposphere. Hsu and Lin (2007) demonstrated the prevalence of different mechanisms in the different phases of the PJ pattern: more extratropical Eurasian influence in the positive phase, but more tropical WNP influence in the negative phase. Considering that the PJ-like pattern is induced by various mechanisms (e.g., ENSO, WNP SST anomaly, Tibetan Plateau and Eurasian

![Composite difference of the summer surface pressure anomalies and 850-hPa wind anomalies](attachment:image)
heating, etc.), it has been recently proposed that zonally elongated meridional circulation patterns, such as the PJ pattern, in the WNP are intrinsically dynamical modes that can be easily excited in the East Asian and WNP summer monsoon mean flow (Kosaka and Nakamura 2006; Lu et al. 2006; Hsu and Lin 2007).

6. Final Remarks

The East Asian, Indochinese and western North Pacific monsoons are essential parts of the integral A-AM system. This chapter summarizes the main findings from studies of the three components of the A-AM system. Understanding East Asian, Indochinese, and western North Pacific monsoon variability has been a challenging task. To stimulate further research in this field and to improve our understanding of the dynamics of the East Asian, Indochinese, and western North Pacific summer monsoon systems, we list some issues that call for further investigations.

a) Diurnal Cycle and Sub-seasonal Variability:
- What are the physical processes governing the diurnal cycle of monsoon rainfall over continental East Asia?
- How can weather and climate models accurately reproduce the diurnal cycles of monsoon rainfall?
- How can these models capture the multi-scale interactions between the three monsoon regions?

b) Interannual Variability:
- How does the interannual variation in each region’s monsoons affect variability in other regions?
- How can the contributions of the ENSO, land surface processes, Tibetan Plateau forcing, and high latitude teleconnections to the interannual variability of the A-AM system be separated and understood?
- How can the major problems of climate models in simulating monsoon rainfall be identified, and what possible solutions can be found to overcome these problems?

c) Long-term Variability:
- How can variability of the East Asian, Indochinese and western North Pacific monsoons be understood in terms of global monsoon variability?
- How can the impacts of natural and anthropogenic factors on long-term changes in the A-AM monsoon be accounted for?
- What future directions should be explored for climate modeling and for predicting monsoon variability, as well as changes in monsoons during the 21st century as a result of global warming?

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