ADVANCES IN ATMOSPHERIC SCIENCES, VOL. 32, JANUARY 2015, 10–31

From MONEX to the Global Monsoon: A Review of Monsoon System Research

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(Received 21 July 2014; revised 28 September 2014; accepted 29 September 2014)

ABSTRACT

Substantial progress has been made over the past three decades since the Monsoon Experiments (MONEX) of 1978–79. Here, we review these achievements by highlighting four breakthroughs in monsoon research: (1) The identification of the coupled ocean–land–atmosphere nature of the monsoon in the process of the annual cycle of solar heating; (2) new understanding of the changes in the driving forces of monsoon systems, with anthropogenic factors (climate effects of increased greenhouse gas and aerosol emissions) playing an important role in the regulation of monsoons; (3) detection of the interdecadal- and centennial-scale variability of monsoon systems, and its attribution to the combined impact of global warming and natural (especially oceanic) effects; and (4) the emerging concept of the global monsoon and its long-term variation under the impact of global climate change. All the observational and model-derived evidence demonstrates that the monsoon system, as an important component of the global climate system, has already changed and will continue to change in the future. This picture of an evolving monsoon system poses great challenges for near-term prediction and long-term projection.

Key words: monsoon, coupled system, anthropogenic activity, global monsoon

Citation: Ding, Y. H., Y. J. Liu, Y. F. Song, and J. Zhang, 2015: From MONEX to the global monsoon: A review of monsoon system research. *Adv. Atmos. Sci.*, **32**(1), 10–31, doi: 10.1007/s00376-014-0008-7.

1. Introduction

Monsoon research has a long history (Webster, 2006), but the most substantial progress has been made since the 1960s (Ding, 2007; Ha et al., 2012; Hsu et al., 2014). In particular, the First GARP (Global Atmosphere Research Plan) Global Experiment (FGGE) Monsoon Experiments in 1978–79 (MONEX) was the most important milestone in the development of Asian monsoon research. Many new findings were made and numerous challenging issues raised and explored. The vital importance of monsoon prediction to the economic and social development of the Asian monsoon region, as well as the lives of the people living there and the resources upon which they depend, was widely realized.

During the three decades since MONEX, an increasing number of studies have been devoted to the various issues related to the monsoon. Among them, most have focused on the typical tropical monsoon systems of the Asian or Asian–Australian monsoon. In a recent review, Ding (2007) listed nine areas in which major achievements have been made since MONEX: (1) global perspective of the Asian monsoon; (2) the seasonal march and annual cycle of the Asian monsoon, including the onset, active-break cycle and withdrawal; (3) multiple-scale (i.e. intraseasonal, interannual, interdecadal) variability of the Asian monsoon; (4) the energy and water cycle of the Asian monsoon; (5) synoptic systems, mesoscale process and diurnal variation associated with the Asian monsoon; (6) large-scale physical processes and dynamics of the Asian monsoon, including the dynamics of tropical waves, the coupled monsoon system and SST–monsoon relationship, ecosystem–monsoon relationships and land–monsoon interaction, snow–monsoon relationships, Rossby wave teleconnection theory and the dynamical and thermal effects of the Tibetan Plateau (TP); (7) predictability and prediction of the Asian monsoon; (8) the evolution of the monsoon in the paleoclimate, including centennial and millennial variability; and (9) impacts of the Asian monsoon on the socioeconomic sectors, especially agriculture and water resources. Compared to the pre-MONEX era, perhaps the most important achievements have come from the (1) identification of the coupled ocean–land–atmosphere nature of the monsoon; (2) emergence of the concept of the global monsoon and its relationships with regional monsoon systems; (3) inclusion of new anthropogenic driving forces (e.g. global warming due to increases in greenhouse gases (GHGs) emissions and cooling due to increases in aerosols) and changes in the main driving forces of regional monsoon systems (ocean and land under the impact of climate change); and (4) detection of the long-term variation in certain characteristics of the global and regional monsoons (e.g. monsoon onset, circulation, precipitation) and multiple-scale variability, especially interdecadal variability under the effects of climate change.

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The present review focuses on the above four themes, and the remainder of the paper is structured as follows: Section 2 deals with the long-term variation in the driving forces of the monsoon system; section 3 discusses interdecadal variation in the onset and intensity of the Asian summer monsoon; section 4 is devoted to discussion on the global monsoon; and section 5 is a summary. In addition, it is important to point out that considerable advancements have been made in monsoon weather and climate prediction and its application to society. A vast amount of effort has been devoted to improving model initialization, resolution, physical parameterization, and ensemble techniques. In particular, the multimodel ensemble is a very promising approach for dynamical prediction of the monsoon. However, owing to space limitation, these aspects are not covered in this review.

2. Long-term variation of the driving forces of the monsoon system

2.1. *Coupled land–ocean–atmosphere monsoon system*

It has long been known that for a monsoon to be established, a thermal contrast between the land and ocean must exist. Due to a much more rapid response to the seasonal cycle in solar heating, large land masses heat up more significantly during the spring and summer than the surrounding ocean, thus establishing a large temperature gradient that can drive the generation of a steady monsoon circulation or wind regime. At the same time, the sustained low-level wind blowing toward these hot land masses draw humid air in from surrounding oceans, thus bringing the arrival of the rainy (or wet) season. Therefore, the land–sea thermal contrast is a primary driving force for regional monsoon. Strong monsoon occurs where there is a pronounced north–south distribution in the land and ocean, taking advantage of the north–south progression of the solar seasonal cycle (Slingo, 2002). In this process, a cross-equatorial pressure gradient develops. However, this simple monsoon picture, like a mass breeze, is further modified by the following physical properties (Webster and Fasullo, 2002): (1) the coupled land–ocean–atmospheric system, or the coupled land–ocean–atmosphere–cryosphere–biosphere system (Wang et al., 2006); (2) differential heating of the land and ocean produced by the different heat capacity of land and water; (3) the different manner in which heat is transferred vertically and stored in the ocean and land; (4) modification of differential heating by moist process, including the condensational heating caused by monsoon precipitation (e.g. Jin et al., 2013); (5) the generation of meridional-pressure-gradient forces resulting from the differential heating; (6) the meridional transport of heat in the ocean by dynamical processes; (7) the rotation of Earth, which has to be considered for each of the above processes and properties; and (8) the influence of local effects such as the geography of the ocean and land masses, and regional topography. Therefore, the genesis and maintenance of a monsoon system (e.g. the Asian monsoon system) is attributed to the complex interaction of multiple influencing factors under the effect of annual variation of solar forcing.

Furthermore, a fundamental problem is what important roles the ocean and land can play in regulating the coupled ocean–land–atmosphere monsoon system as natural driving forces. Or, if the atmospheric and oceanic components of monsoon are coupled, what are the processes responsible for the coupling? The fundamental clue to understanding the coupling comes from the fact that the majority of mass and heat transport in the ocean is wind-driven (Webster, 2006). The transport is a combination of geostrophic and Ekman transport. In the Northern Hemisphere (Coriolis parameter, $f > 0$) the transport will be southward if the winds are westerly and northward for easterly winds. As $f < 0$ in the Southern Hemisphere, the reverse is true. On the other hand, vertically averaged oceanic Ekman transport is to the right of the surface wind in the Northern Hemisphere and to the left of the wind in the Southern Hemisphere. Consequently, the overall oceanic transport has the opposite sign to the atmospheric heat transport in the direction of the lower tropospheric divergent wind. The overall effect of the wind-driven oceanic heat transport is to cool the SST in the summer hemisphere and warm the SST in winter, thus reducing the cross-equatorial SST gradient. The reduction of the SST gradient will on the one hand reduce the cross-equatorial pressure gradient, which will have the effect of reducing the intensity of the monsoon winds, while on the other hand it will reduce the SST in the summer hemisphere, which will in turn reduce the saturated vapor pressure of an air parcel approaching the continental region. The reduced monsoon winds will lower the convergence of moisture into the continental regions, thus reducing the release of latent heat in the precipitation regions. The overall impact of the above two processes is to reduce the amplitude of the monsoon annual cycle. As pointed out by Webster (2006), without ocean heat transport, the SST extreme would produce a monsoon that would be very different from that which is observed today, and would quite likely be much wetter.

The coupled nature of the monsoon extends from the intraseasonal to the interannual and interdecadal timescales. A number of studies have been devoted to interseasonal and interannual oceanic variability and their relationship to monsoon, especially the impact of ENSO and Indian Ocean Dipole (IOD) variability on monsoon rainfall, which has been comprehensively reviewed previously by, for example, Waliser (2006), Yang and Lau (2006), and Ding et al. (2013). The present review deals only (in section 3) with the relationship between interdecadal oceanic variability [Pacific Decadal Oscillation (PDO) and Atlantic Multidecadal Oscillation (AMO)] and the Asian monsoon. The fully coupled nature of the monsoon also involves interaction with the land surface in the continental regions, especially the Asian land mass where a huge elevated area in the form of the TP exists. Among the land surface processes, soil moisture, snow cover and vegetation conditions play significant roles in driving the Asian monsoon through changing the surface sensible and latent heating as well as surface albedo. Actually, the most important impacts come from variation in soil moisture and snow cover. Snow cover may have an important effect on the interannual variability of the monsoon because of its ability to alter the surface albedo and regulate soil moisture (Yang and Lau, 2006; Yasunari, 2006), which in turn can affect the subsequent monsoon circulation as a persistent or long memory signal.

2.2. *Driving forces of the monsoon system and their changes*

2.2.1. *Ocean and land*

Studies on the relationship between snow and monsoon have a long history, in which the main feature is a negative relationship: the Indian summer monsoon, for example, is generally weaker (stronger) than normal when more (less) preceding winter and spring snow cover occurs in Eurasia (Liu and Yanai, 2002). Such a relationship is consistent with Bamazai and Shukla (2003) for western Eurasia, but is different for central Eurasia (Shinoda, 2001), suggesting that such a carry-over on the summer circulation may not be produced directly through land-surface–atmosphere interaction. In contrast, the preceding winter and spring snow cover days/depths over the TP have a generally positive relationship with the Asian summer monsoon (Wu and Qian, 2003; Zhang et al., 2004) and the East Asian summer monsoon precipitation in the region of the Yangtze–Huaihe River basin (Ding,

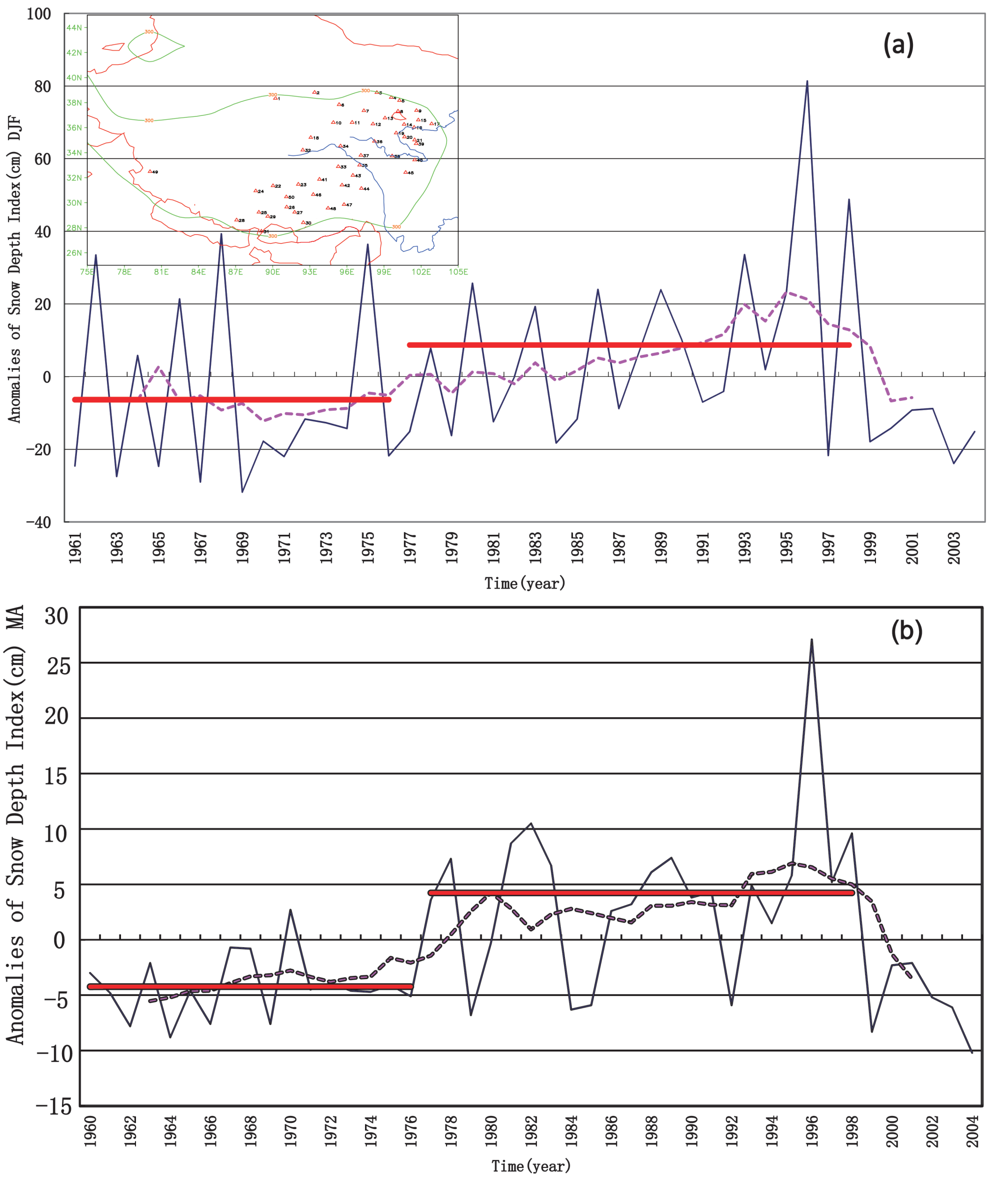

Fig. 1. Time series of snow depth (SDI) over the TP averaged for 50 surface stations (see the inserted small diagram in the upper-left corner) for (a) winter (DJF), (b) spring (March, April), and (c) the first EOF mode time coefficient series for spring. Horizontal bars represent averages for different periods, and dashed lines the 9-yr running average (units: cm) [Reprinted from Ding et al. (2009)]. (d) Time series of winter snow depth (cm d^{-1}) over the TP averaged for the 72 stations for 1979–2011. Horizontal solid lines indicate averaged values for the two decadal periods of 1979–99 and 2000–11. [Reprinted from Si and Ding, (2013)]

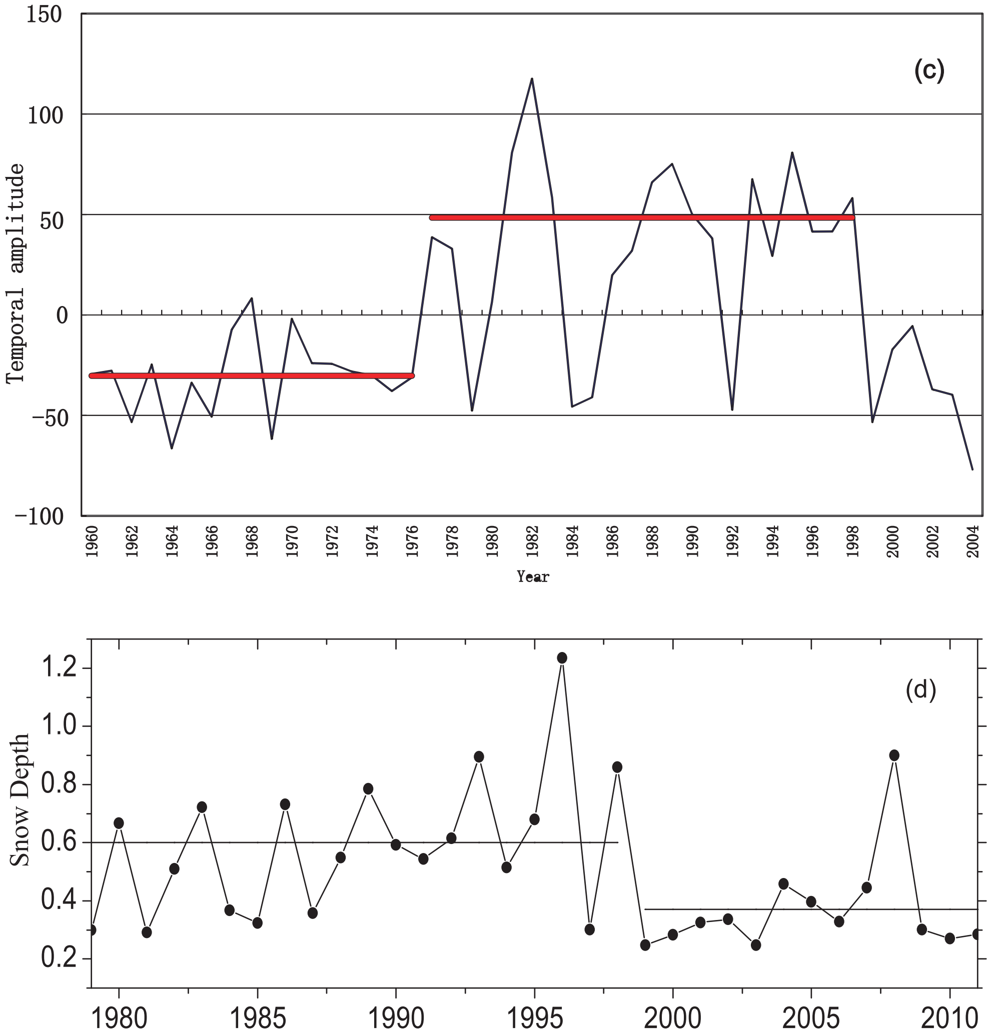

Fig. 1. Continued.

2007; Ding et al., 2008). Figure 1 shows the time series of snow depth index (SDI) over the TP for winter and spring 1960–2011. It can be seen that SDI over the TP underwent an increasing trend during this period, with a significant rise around 1977. This high-level SDI then remained for nearly 20 years until around 1999, after which it dropped down to a low level. Therefore, the winter and spring snow over the TP experienced a low–high–low pattern of variation from 1960 to 2011. It has long been recognized that the thermal conditions or heat source/sinks in spring and summer over the TP are a fundamental driving force for the development of the Asian summer monsoon (Yeh and Gao, 1979; Murakami and Ding, 1982; Luo and Yanai, 1984). It is very important to estimate the long-term variation in the heating conditions of spring and summer over the TP caused by increasing or decreasing trends in winter and spring snow. Figure 2 presents the time series of anomalous vertically integrated apparent heat sources (Q1) averaged for the whole TP for summer and spring. It is clear that Q1 assumes a similar pattern of long-term variation to winter and spring snow from 1960 to 2011, but with opposite sign (i.e. positive–negative–positive). Similarly, land–sea thermal contrast (QLS) has also experienced a similar three-stage long-term pattern of variation (Fig. 3). As the main driving force, the Asian summer monsoon clearly manifests interdecadal variation of strong–weak–strong intensity (see section 3.2). Figure 4 is a schematic diagram outlining the possible cause of the weakening and intensification of the Asian summer monsoon. It illustrates the importance of snow cover over the TP as a coupled forcing of the long-term variation of the Asian summer monsoon.

2.2.2. *Anthropogenic forcing*

There is growing evidence that the monsoon and its variation may be affected by anthropogenic forcing, mainly in the form of increases in GHGs emissions and aerosols. The effect of global warming caused by increased emissions of GHGs can lead to increases in moisture convergence due to the increased water vapor in the air column, and increased surface evaporation due to warmer surface temperature, thus changing the water cycle. Global and regional energy budgets will also be modified due to the change in radiative forcing at the top of the atmosphere, thus modulating the land–sea

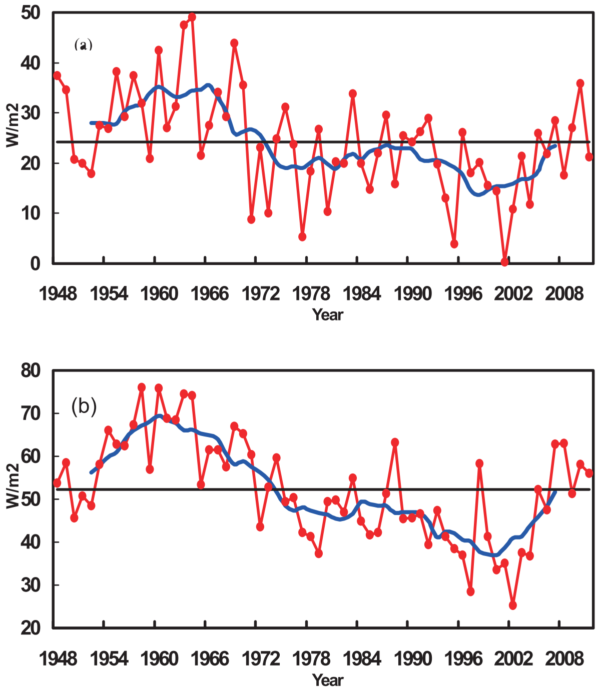

Fig. 2. Time series of the vertically integrated (surface to 200 hPa) atmospheric heat source averaged for the TP (28°–43°N, 70°–105°E) for (a) spring and (b) summer. The blue lines denote 9-yr mean curves. [Reprinted from Ding et al. (2014b)]

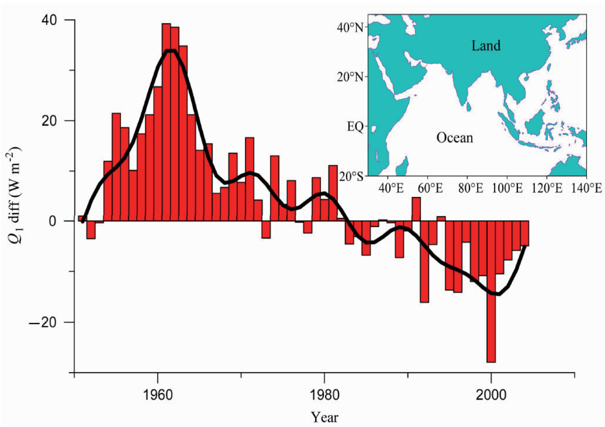

Fig. 3. Time series of the land–sea thermal index (LS) and the apparent heat source ($Q1$) difference between land areas (547 grid points) and oceanic areas (668 grid points) in the Asian monsoon and neighboring oceanic regions (20°S–45°N, 30°–140°E). The inserted diagram shows the domain of land and oceanic areas for estimating QLS. The solid line is the 9-yr running average (units: W m^{-2}). [Reprinted from Ding et al. (2009)]

thermal contrast, due to the change of energy gain of ocean and land. The configuration and strength of atmospheric circulation will be changed. Therefore, the global and regional monsoons will possibly be changed with this new additional driving force.

A key problem is separating the effect of anthropogenic global warming from natural forcing in terms of their effects on the decreasing interdecadal trend of the global and regional monsoons (e.g. the Asian monsoon) over the last half century, as well as the increasing trend during 1970–2008 when the oceanic monsoon is combined with the land monsoon to characterize the global monsoon. Recent studies have clearly indicated that the monsoon and its variation (monsoon precipitation and circulation) may be affected by anthropogenic forcing [e.g. Turner and Annamalai, 2012; Intergovernmental Panel on Climate Change (IPCC), 2013]. Annamalai et al. (2013) revealed that anthropogenic forcing through SST warming over the tropical West Pacific likely caused the drying trend over South Asia during the last 5 to 6 decades. Their model simulation experiments with the inclusion of the recent increase in GHGs concentrations suggested that the rising SST trend over the warm pool, perhaps anchored by an increase in GHGs concentrations, is instrumental in the east–west shift in monsoon rainfall (enhanced rainfall over the tropical West Pacific and decreased rainfall over South Asia) through changing the atmospheric circulation. Douville (2008) studied the Indian monsoon response to increasing concentrations of GHGs through assessing the sensitivity of the monsoon response to the magnitude and patterns of SST anomalies in the circumpolar Southern Ocean and tropical Pacific Ocean in a coupled climate model. The study indicated that the Indian monsoon response to increasing amounts of GHGs is sensitive to regional uncertainties in the prescribed SST warming. Within the tropical Pacific, the results demonstrated an even stronger impact on the monsoon response, through a modification of the Walker circulation. In the future, it seems likely that the anthropogenic effect will continue to increase. Coupled Model Intercomparison Project Phase 5 (CMIP5) models project that the global monsoon area and global monsoon total precipitation is likely, or very likely, to increase by the end of the 21st century (IPCC, 2013). For example, as a major monsoon system in the tropics, the total precipitation of the Asian–Australian monsoon system is projected to increase by about 7%, mainly due to an increase in Indian summer monsoon and East Asian summer monsoon rainfall (by 9%–10%). On the other hand, the monsoon circulation has weakened and is expected to continue to weaken in the future. The above results demonstrate that global warming is likely to affect the global and regional monsoons through its impact on the thermal conditions of surrounding oceanic regions. Changes in direct forcing, such as a reduction in the land–ocean thermal gradient under the effect of global warming can lead to a reduction in regional monsoons, e.g. the Indian monsoon, even though the large-scale monsoon may not change. The above results clearly indicate that the anthropogenic effect on monsoon is realized through the response of coupled land–sea–atmosphere processes.

Furthermore, in the expected warmer climate of the 21st century, despite weakened cross-equatorial flow, the time-mean precipitation over the Indian Peninsula is predicted to increase by about 10%–15% (Stowasser et al., 2009). This paradox is interpreted by suppressed equatorial precipita-

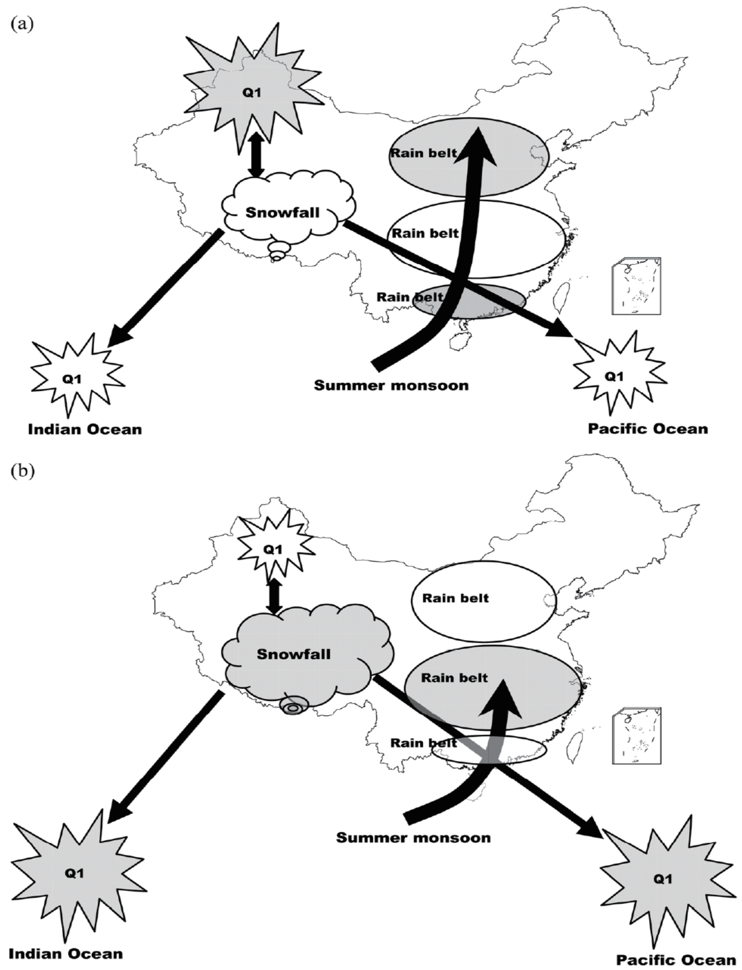

Fig. 4. Schematic diagram of the possible cause of the weakening of the Asian summer monsoon. An anomalously strong and weak Asian summer monsoon condition is illustrated by (a) and (b), respectively. Shaded areas indicate large snow cover, high SST anomaly, intense atmospheric heating, and large precipitation amounts. Bold arrows represent the summer monsoon airflow. [Reprinted from Ding et al. (2009)]

tion that can weaken cross-equatorial monsoon flow in Somalia and alter air–sea interaction over the southern Arabia Sea, thus leading to an increase in SST and rainfall over the Indian Peninsula. Another feature of many model projections is an increase in time-mean rainfall over India, and these results are generally reproducible in CMIP5 (IPCC, 2013). While intensity changes are model-scenario-dependent, agreements in the sign of the change are consistent with the expected changes due to thermodynamic effects under a warmer climate. Another interpretation of how global warming may have contributed to the weakening of the large-scale Asian summer monsoon (i.e. monsoon circulation) in recent decades is the relatively smaller warming magnitude in Asia compared to surrounding regions (Yu et al., 2004; Ding et al., 2008, Zuo et al., 2012). Although the surface temperature over Asia has increased considerably, the middle and high tropospheric temperature anomalies in recent decades (since the end of 1970s) have been relatively colder than those in surrounding areas. This could have led to the significant weakening of the tropical easterly jet and decreasing trend of the Webster-Yang monsoon index(U_{850}–U_{150}). However, it is not clear why the warming in the Asian landmass has been smaller than that over other regions under the global warming background. This relatively weaker warming, especially over the TP, will also likely lead to a future continued weakening of the Indian summer monsoon (Sun and Ding, 2009, 2011). For the East Asian summer monsoon, few studies have been conducted to detect the anthropogenic signal in the cause of its weakening (Wang, 2001; Yu et al., 2004; Ding et al., 2008). Generally, the weakening is attributed to natural factors, such as the 60–80-yr oscillation, mainly caused by oceanic interdecadal variability (Ding et al., 2009). However, Li et al. (2010) indicated that the primary response of the East Asian summer monsoon to a warming climate may be a position change of the major monsoon rainbelt, rather than an intensity change. Such a position change may have led to the observed increase in floods in the middle and lower reaches of the Yangtze River and southern Japan, and suppressed rainfall over South China and the Philippine Sea, from 1950 to 2008.

For future change in the East Asian summer monsoon, the precipitation in CMIP Phase 3 (CMIP3) models under the A1B scenario of the IPCC's Special Report on Emissions Scenarios (SRES) increases gradually ($\sim$ 9%) by the 2040s due to an increase in both circulation and water vapor (Sun and Ding, 2009). CMIP5 projections similarly indicate a likely increase of East Asian summer monsoon circulation and precipitation throughout the 21st century under both the 4.5 and 8.5 Representative Concentration Pathway (RCP4.5 and RCP8.5) scenarios (IPCC, 2013). This is different to other parts of the Indian monsoon, which may weaken in terms of monsoon circulation, but increase in terms of rainfall (Sun and Ding, 2011; Bollasina et al., 2011).

Besides warming by GHGs, climate change in monsoon regions may be strongly affected by increased aerosol forcing and induced feedback (e.g. Ramanathan et al., 2005). The aerosol direct forcing (heating) in the atmosphere and negative forcing (cooling) at the surface alters atmospheric stability and establishes horizontal pressure gradients that modulate the large-scale circulation and hence monsoon rainfall. In the IPCC Assessment Reports on Climate Change, the impact of aerosol on the Asian monsoon has been continuously dealt with and assessed. Based on the earliest climate model projections under the forcing of GHGs and with the inclusion of sulfate aerosols, the monsoon in India and some regions of China would weaken and related precipitation would decrease by 10%. The main reason involved a decreasing land–sea thermal contrast due to the cooling effect of hazy clouds and a decrease in water vapor content over land areas. In the IPCC's Third Assessment Report and the Second Assessment Report (IPCC, 2001; The Editorial Committee for Second National Assessment Report on Climate, 2011), these conclusions were further verified. In particular, the impact of aerosols on the Indian summer monsoon was extensively studied. A major finding was the "elevated heat pump" effect caused by black carbon, which favors the intensification and an earlier onset of the South Asian summer monsoon (Ramanathan et al., 2005; Lau et al., 2005; Lau and Kim, 2006; Bollasina et al., 2011). The recent weakening tendency in seasonal monsoon rainfall in the Indian monsoon region has been attributed to factors including black carbon and/or sulphate aerosols (Bollasina et al., 2011). Model simulations that include sulfate and black carbon have indicated a wide variety of aerosol effects on monsoon climate, including a weakened intensity of the East Asian summer monsoon and decreased monsoonal precipitation (Sun and Liu, 2008). Therefore, nearly all model studies have reached the same conclusion that the holistic climate effect of aerosols is to reduce summer monsoon circulation and associated precipitation in East Asia and precipitation in South Asia, but black carbon and dust aerosols possibly enhance convective activity and monsoonal precipitation in some local areas (e.g. the Indo-Gangetic Plain in May–June).

At the same time as emissions and the atmospheric loading of aerosols over the Asian monsoon regions have increased in recent decades (IPCC, 2013), observations show that South Asia experienced a widespread summertime drying during the second half of the 20th century. However, it is unclear whether this trend was due to natural variation or human activity. A series of climate model experiments designed to investigate the South Asian monsoon response to natural and anthropogenic forcing have been conducted, and the results suggested that the observed precipitation decrease could be mainly attributed to human-influenced aerosol emissions (Bollasina et al., 2011). The drying may have been the outcome of a slowdown in the tropical meridional overturning circulation, which led to a weakening of monsoon rainfall and thus compensated for the aerosol-induced energy imbalance between the Northern and Southern hemispheres. These results provide compelling evidence of the prominent role of aerosols in shaping regional climate change over South Asia.

2.2.3. *Summary*

Based on the work reviewed in this section, the variation in monsoonal precipitation, intensity and timing is closely related to changes in multiple influencing factors or driving forces, as well as the effect of climate change. These include land–sea thermal contrast, land cover and use, atmospheric moisture content, and atmospheric aerosol loadings. The variation in these factors, to different extents, have been subject to climate change, thus further causing the following changes to the global and regional monsoon systems, as a response of the coupled atmosphere–land–ocean system:

(1) Enhanced land–sea thermal contrast due to a more rapid warming over land than over the ocean surface. However, the tropical atmospheric overturning circulation, on average, slows down as the climate warms, due to energy balance constraints in the tropical atmosphere. These changes in the atmospheric circulation lead to regional changes in monsoon intensity, area and timing (IPCC, 2013).

(2) Surface heating change due to different intensities of solar radiation absorption during the annual cycle of solar heating, which itself is affected by any land-use change that alters the albedo of the land surface.

(3) Atmospheric aerosol loading, such as air pollution, affects how much solar radiation reaches the ground, which can change the monsoon circulation by altering the summer solar heating of the land surface. Absorption of solar radiation by aerosols, on the other hand, warms the atmosphere, changing the atmosphere's heating distribution and stability.

(4) The strongest effect of climate change on the monsoon is the increase in atmospheric moisture associated with warming of the atmosphere as a feedback effect, resulting in an increase in total monsoon rainfall even if the strength of the monsoon circulation weakens or does not change (Fig. 5). This figure indicates that the effect of various factors or driving forces under the influence of climate change leads to increased monsoonal precipitation and a weakening of monsoon circulation.

(5) The oceanic interannual to interdecadal variabil-

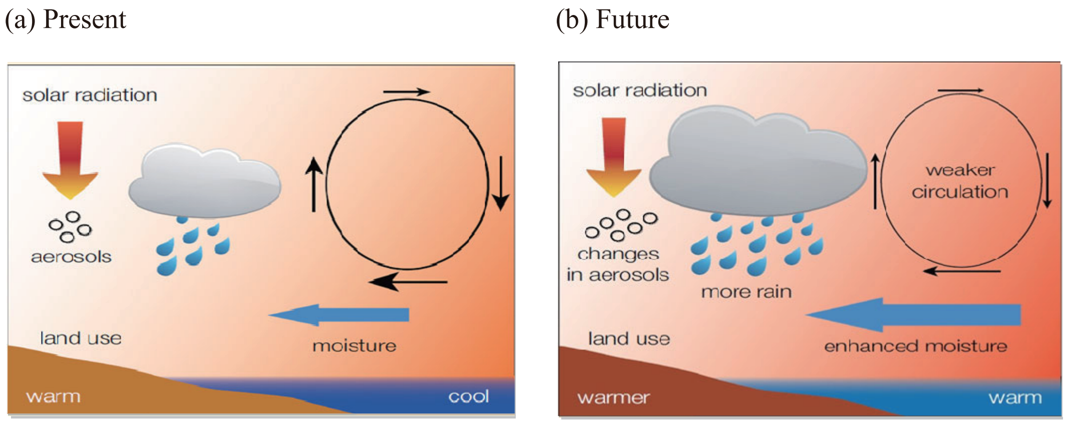

Fig. 5. Schematic diagram illustrating the main ways that human activity influences monsoon rainfall as the climate warms, increasing water vapor transport from the ocean to over land because warmer air contains more water vapor. This also increases the potential for heavy rainfall. Warming-related changes in large-scale circulation influence the strength and extent of the overall monsoon circulation. Land-use change and atmospheric aerosol loading can also alter the amount of solar radiation that is absorbed in the atmosphere and by land, potentially moderating the land–sea temperature difference. [Reprinted from Christensen et al. (2013a)]

ity may play a more active role in driving large-scale or hemispheric-scale monsoon, and not only as a component of land–sea thermal contrast. Recently, Wang et al. (2013) revealed that the intensification of Northern Hemisphere summer monsoon (NHSM) can be primarily attributed to a "mega" ENSO. They showed that during the recent warming of about 0.4°C since the late 1970s, a decadal change in precipitation and circulation is clear across the entire NHSM system, with the NHSM as well as the Hadley and Walker circulations all showing substantial intensification. At the same time, a striking increase of NHSM rainfall by 9.5% per 1°C of global warming is apparent. These factors, driving the present changes of NHSM system, are instrumental for understanding and predicting future decadal change, and determining the properties of climate change that are attributable to anthropogenic effects and long-term variability in the complex system.

3. Interdecadal variation of the onset and intensity of the Asian summer monsoon

3.1. *Onset of the Asian summer monsoon*

The onset of the Asian summer monsoon signifies the commencement of the rainy season in this region. As an important component of the Asian summer monsoon, it is one of the most spectacular features, with its sudden increase in precipitation and convective activity as well as a rapid reversal of the low-level pre-monsoon easterly to the monsoon westerly. Many studies have been devoted to the definition and indices of the onset, large-scale circulation conditions, multiple-scale variability, and the fingerprints of global warming on the onset (Webster and Yang, 1992; Wang and Fan, 1999; Goswami et al., 1999; Lau et al., 2000; Janowaiak and Xie, 2003; Fasullo and Webster, 2003; Wang et al., 2009a; Huang et al., 2007; Kajikawa and Wang, 2012; Kajikawa et al., 2012; Zhou and Murtugudde, 2014; Hsu et al., 2014). The most important finding derived from these studies is the advanced onset of the South Asian summer monsoon and South China Sea summer monsoon. Kajikawa et al. (2012) revealed an advanced onset of the monsoon over the Bay of Bengal and the western Pacific during the 30 years from 1979 to 2008, due to intensification of thermal contrast between the Asian land mass and the tropical Indian Ocean in May. This intensification of land–sea thermal contrast is primarily caused by a warming trend over land. Xiang and Wang (2013) suggested that shift to a La Niña-like mean state has promoted the advanced monsoon onset in recent decades. It should be pointed out that this oceanic mode may be closely related to the change of the PDO. Krishnamurthy and Krishnamurthy (2013) showed that the warm (cold) phase of the PDO is associated with a deficit (excess) of rainfall over India in June–September. Recently, Watanabe and Yamazaki (2014) further found that the early onset of summer monsoon in Kerala, considered as the beginning of India's rainy season by many studies (e.g. Wang et al., 2009a), is closely associated with the negative PDO through the teleconnection of the stationary wave train, thus intensifying the land–sea thermal contrast. Therefore, the interdecadal variability of heating or warming over the Atlantic Ocean and Pacific Ocean is a major factor in early monsoon onset.

Kajikawa and Wang (2012) and Huang et al. (2011) also detected a significant sudden change in the South China Sea summer monsoon onset around 1993/94: the mean onset date is 30 May for 1979–93 and 14 May for 1994–2008, representing a shift (advancement) of more than two weeks. These studies stressed the important effect of the enhanced activity of northwestward-moving tropical disturbances from the equatorial western Pacific for the advanced onset during the second period (1994–2008), attributed to a significant increase in SST over the western Pacific from the 1980s to 2000s. They believed, therefore, that the advanced South

China Sea summer monsoon onset is rooted in the decadal change of the SST over the equatorial western Pacific—a conclusion that appears to be consistent with previous results. However, an interesting question is whether or not this interdecadal variation of the land–sea thermal contrast is ultimately caused by climate change or global warming.

Another important question related to the early onset of the Asian summer monsoon is its triggering mechanism. It is well known that the onset date of the Indian summer monsoon has a pronounced interannual variability, with outbreaks in early May in some years and delayed onsets until mid-June in other years. Based on the relationship between ENSO and monsoon onset, it is difficult to predict the exact timing of monsoon onset due to this relationship being far from linear. Other factors, such as the impact of land–sea thermal contrast, large-scale topography, instabilities of large-scale circulation, and positive ocean–atmosphere feedback, are also important influencing factors (Zhou and Murtugudde, 2014). However, the above factors are slow and continuous in nature. The triggering mechanisms should be a fast process to cause such sudden or rapid onset of the monsoon, if other preconditions are satisfied. Wang et al. (2009a) showed that a prominent onset precursor of the Indian summer monsoon on the biweekly time scale is the westward extension of convective centers from the equatorial eastern Indian Ocean toward the southeast Arabian Sea. On the intraseasonal time scale [i.e. intraseasonal variability (ISV)] the onset tends to be led by a northeastward propagation of an intraseaonal convective anomaly from the western equatorial Indian Ocean. In boreal summer, these northward-propagating ISVs carry a large amount of moisture and momentum from the tropics to the subtropics (Saha et al., 2012), which delivers a strong pulse of perturbation to the background state over the Bay of Bengal and the Indian subcontinent. Therefore, northward-propagating ISVs are a key precursor of monsoon onset, as pointed out by Wang et al. (2009a). Strong northward-propagating ISVs tend to favor an early onset of the Indian summer monsoon. However, the study by Zhou and Murtugudde (2014) shows that not all early onsets are attributable to such northward-propagating ISVs, and they do not all lead to an early onset of the Indian summer monsoon. Based on a dataset of the onset of the Indian summer monsoon from 1982 to 2011, the study indicates that, for years of early onset that are closely related to northward-propagating ISVs, the convective features are prominent, as a key index. The ocean–atmosphere interaction is found to be important for northward-propagating ISVs before the Indian summer monsoon onset. Evidence shows that warm SST anomalies drive the atmosphere and lead to the atmospheric instability and convection.

The above analyses stress the importance of northward-propagating ISVs with strong convective activity over warm SST oceans in the initiation of an early onset of the Indian summer monsoon. This precondition has also been found in the South China Sea by Kajikawa and Wang (2012), who detected a significant advance in the onset dates around 1993/94, as pointed out above (Fig. 6). The advanced onset during the second period (14 May for 1994–2008) is affected by enhanced activity of northward-propagating tropical disturbances from the equatorial western Pacific, which includes enhanced ISVs during the period from mid-April to mid-May and high tropical cyclone activity (roughly doubled compared to the relatively late onsets during 1979–93) passing through the South China Sea and Philippine Sea during the same period. The study also attributed the enhanced ISV and tropical cyclone activity over the South China Sea and Philippine Sea to a significant increase in SST over the equatorial western Pacific from the 1980s to 2000s.

Therefore, overall, the following main conclusions can be drawn: (1) The early onset of the Asian summer monsoon is a decadal (since 1993/94 in the South China Sea) or interdecadal (the past three decades in the Indian summer monsoon) phenomenon; (2) The primary triggering system is northward-propagating tropical disturbances (ISVs and tropical cyclones), with accompanying strong convective activity; (3) Warm SST over the tropical Indian ocean and/or equatorial western Pacific ocean may be a critical condition for northward-propagating ISVs and active convection in the atmosphere. This critical oceanic influence reinforces the more recent view that the ocean does not just play a passive role in

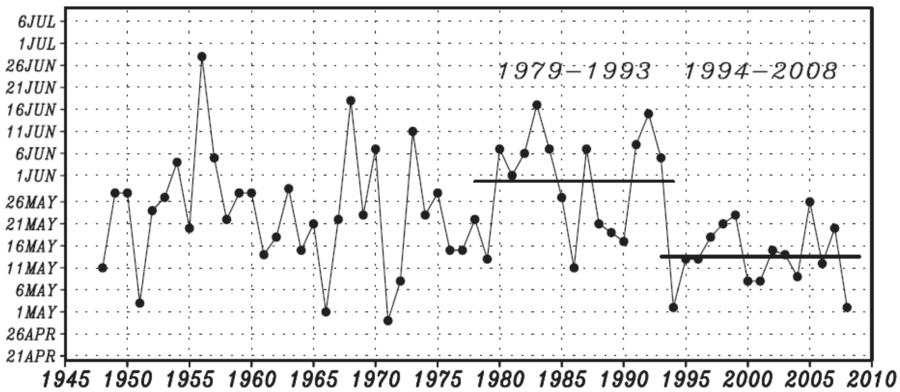

Fig. 6. Time series of the South China Sea summer monsoon onset date. The definition of the onset date is explained in detail in the main text. [Reprinted from Kajikawa and Wang (2012)]

the northward-propagating ISVs; the ocean–atmosphere interaction can play an active role for many ISVs (Goswami, 2012; Zhou and Murtugudde, 2014).

3.2. *Interdecadal variability of the intensity of the Asian monsoon*

3.2.1. *Summer monsoon*

Many studies have shown that the East Asian summer monsoon and South Asian summer monsoon possess significant interdecadal variation, and both of these two monsoon systems experienced an abrupt change in the late 1960s and end of the 1970s with two obvious processes, respectively (Wang, 2001; Xue, 2001; Lü et al., 2004; Guo et al., 2004; Wu, 2005; He et al., 2005; Ding et al., 2013) (Fig. 7). When the East Asian summer monsoon was in its strong period, summer rainfall in North China increased, while rainfall in the valley of the Yangtze River decreased (Guo et al., 2004, Liu et al., 2012). For the Indian summer monsoon, He et al. (2005) noted that the two weakening processes of the Indian summer monsoon were very close in time with the above climate jumps in northern China.

Consistent with the weakening trend, the rainfall pattern in summer in China has taken on a remarkable change, with less rainfall occurring in North China and more over the Yangtze River basin (Fig. 8). Since the early 1990s, the East Asian summer monsoon has enhanced. Although its amplification has not yet reached the level at the end of the 1970s, along with the enhancement of the East Asian summer monsoon, the major rain belt in eastern China has begun to show a decadal northward shift. With the Mei-yu advancing northward, the summer rainfall over the Huaihe River basin has begun to increase (Si et al., 2009; Tang and Wu, 2012; Liu et al., 2012). From 2008, the East Asian summer monsoon has demonstrated an obvious turning signal, with an intensifying trend (Fig.7a). The South Asian monsoon in the past 50 years has mainly shown a weaker feature. Especially during the period from 2000 to 2013, the South Asian summer monsoon was persistently abnormally weaker, corresponding to the less summer rainfall in India in the past 20 years (Fig. 7b).

The interdecadal anomalies of the global ocean–land–atmosphere interaction may have a significant impact on the interdecadal variation of the Asian monsoon, in which ENSO, the PDO, AMO, Arctic sea ice, the Arctic oscillation (AO), and the TP are all important factors affecting the East Asian summer monsoon (Zhu and Yang, 2003;

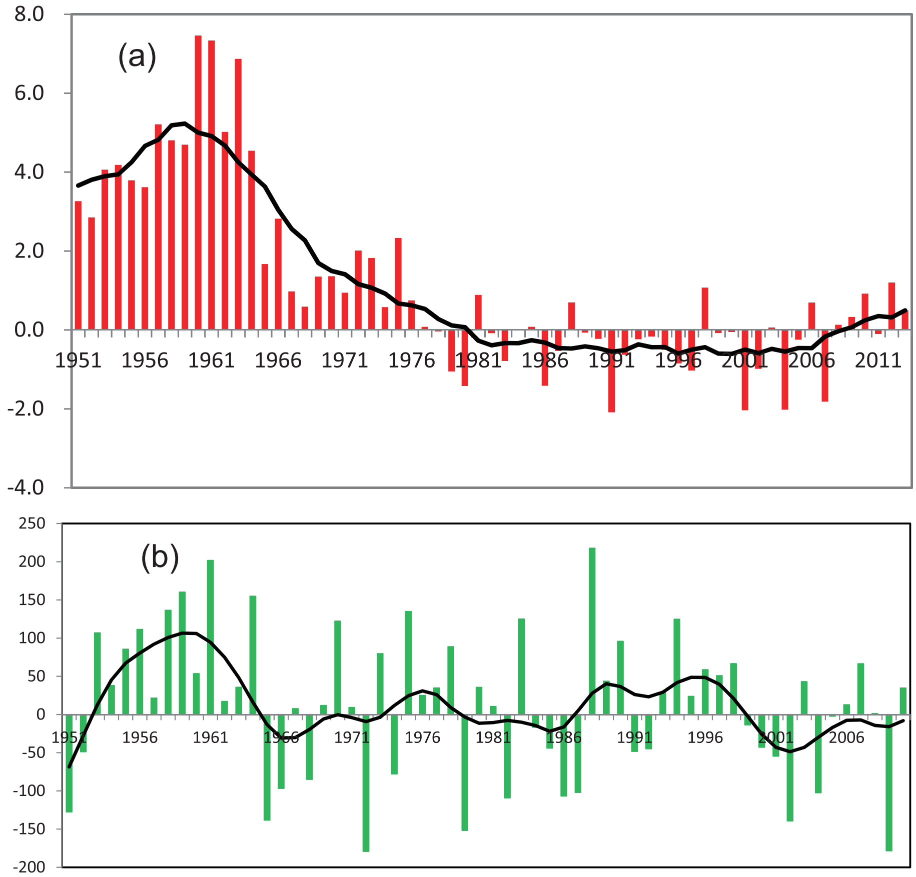

Fig. 7. (a) Temporal variation in East Asian summer monsoon index during 1950/51–2012/13. The black solid line is the Gauss low-pass filtering curve. [reprinted from National Climate Center (2013)] (b) Temporal variation of June–September rainfall anomalies in India during 1961–2010 (unit: mm). The black solid line is the Gauss low-pass filtering curve.

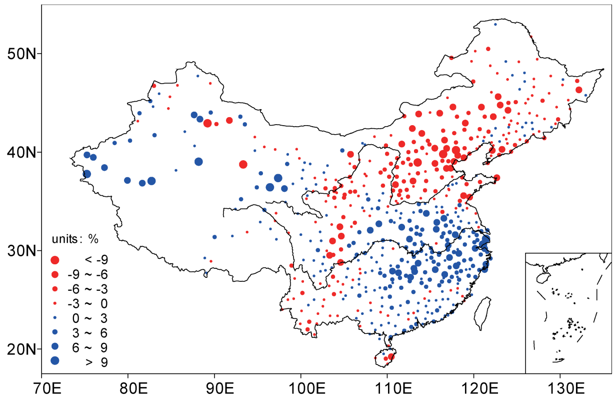

Fig. 8. JJA rainfall trend in China from 1961 to 2010 [units: % (10 yr)$^{-1}$].

Yang et al., 2005; Yang and Zhu, 2008; Sun et al., 2009; Wang et al., 2009b; Zhou et al., 2009; Wang and Zhang, 2010; Yoon and Yeh, 2010; Gong et al., 2011; Wu et al., 2012; Zhao et al., 2012). To summarize, there are three main views on the causes of the interdecadal variation of the Asian summer monsoon: (1) the weakening trend of the Asian summer monsoon from the late 1970s bears certain relationships with the abnormal changes in SST over the warm pool, South China Sea and the tropical Indian Ocean (Xue, 2001), reducing the thermal contrast between the Asian land mass and surrounding oceans; (2) the weakening of the Asian summer monsoon is connected with a reduction in Eurasian snow and an increase in TP snow, mainly responsible for changes in land surface process, especially heat sources; and (3) the interdecadal weakening is related to increased anthropogenic aerosols, such as sulfate aerosols.

As noted in section 2, there is a significant negative correlation between South Asian aerosol concentration in summer and summer monsoon rainfall. As indicated previously, in recent climate simulation experiments, if aerosols are included in models, the monsoon is weakened and drought occurs; otherwise, the monsoon is enhanced (Bollasina et al., 2011; Shen et al., 2013). Ding et al. (2009) argued that the weakening of the Asian summer monsoon in the late 1970s was the result of the combined action of enhanced tropical Pacific SST and increased preceding winter and spring snow over the TP. To begin with, significant SST warming in the tropical eastern Pacific since late 1978 led to more frequent El Niño events; while at the same time, the winter and spring snow increased significantly in the late 1970s and weakened heating (heating sources) over the TP and its neighboring land areas. Both of these two factors reduced the summer temperature difference between the land and sea, thus weakening the driving force of the Asian summer monsoon, which significantly weakened the monsoon (Fig. 4). In addition, other studies show that the temperature change of Eurasian land can also affect the land–sea temperature difference in the East Asian monsoon region and the monsoon intensity (Zhou et al., 2009; Zhao et al., 2011; Zhu et al., 2012; Qian et al., 2012). The warming over Lake Baikal in the past 50 years (1954–2010) is also regarded as an important reason for the weakening of East Asian summer monsoon (Zhu et al., 2012).

3.2.2. *Winter monsoon*

The East Asian winter monsoon has exhibited significant interdecadal variation, with remarkable cycles of 9–10 years, 20–30 years, 40 years, and longer. Although different interdecadal cycles can be obtained using different indices, most winter monsoon indices show and obvious interdecadal weakening for the East Asian winter monsoon since the mid-1980s (Xu et al., 1999; Wu and Wang, 2002; Yan et al., 2003; Chan and Li, 2004; Huang and Wang, 2006; Pei and Li, 2007; Shi et al., 2007; Zhu, 2008; Wang et al., 2010; He and Wang, 2012; Shao and Li, 2012; He, 2013; Wang and Fan, 2013). It is also suggested that the East Asian winter monsoon in this phase is reflective of the most significant weakening over the past 100 years. Recent studies have shown that the weakening period of the East Asian winter monsoon that occurred from the mid-1980s ended by the beginning of the 2000s; and since then, it has recovered and entered into another strong period. Correspondingly, the winter temperature in East Asia also has begun to enter a distinct interdecadal cooling or flattened period. In general, for nearly 60 years, the interdecadal variability of the East Asian winter monsoon can be divided into three periods: the stronger period before the mid-1980s; the weaker period from 1987 to 2004; and the re-enhancement period since 2005 (Liang et al., 2014; Wang and Chen, 2014; National Climate Center, 2013) (Figs. 9 and 10).

The indication is that the interdecadal variation of the East Asian winter monsoon is closely related to atmospheric general circulation and the regional mode of Pacific SST. Some studies suggest that, in the mid and late 1970s, the North Pacific SST warmed remarkably (i.e. the North Pacific SST was in a positive phase of PDO), which was consistent with the rapid rise in average global surface temperature. From the winter of 2005 until now, the North Pacific SST has been in a negative phase of PDO (Tollefson, 2014). This kind of interdecadal variation of the North Pacific SST or PDO has exerted obvious effects on the East Asian winter monsoon, as well as climate, over this region. Numerous recent studies have shown that if the PDO is in a negative (positive) phase, it corresponds to a stronger (weaker) East Asian winter monsoon and colder (warmer) winter temperature (Li and Xian, 2003; Wang et al., 2007; Wang et al., 2008; Ding et al., 2014a).

In addition, a significant negative correlation exists between East Asian winter monsoon index and the AO. The AO, mainly through affecting Eurasian large-scale circulation patterns, e.g. the jet stream and Siberian high, affects the winter monsoon (Gong et al., 2002; Wu et al., 2004; Liang et al., 2014). Since the mid-1980s, the East Asian winter monsoon's weakening was probably directly due to the significantly enhanced AO (He and Wang, 2012). There are also some other studies that have shown that the Northern Hemisphere annular mode (NAM) could cause East Asian winter monsoon anomalies by affecting the variation in the intensity of the Siberian high, or influence the interannual and interdecadal variability of the East Asian climate through af-

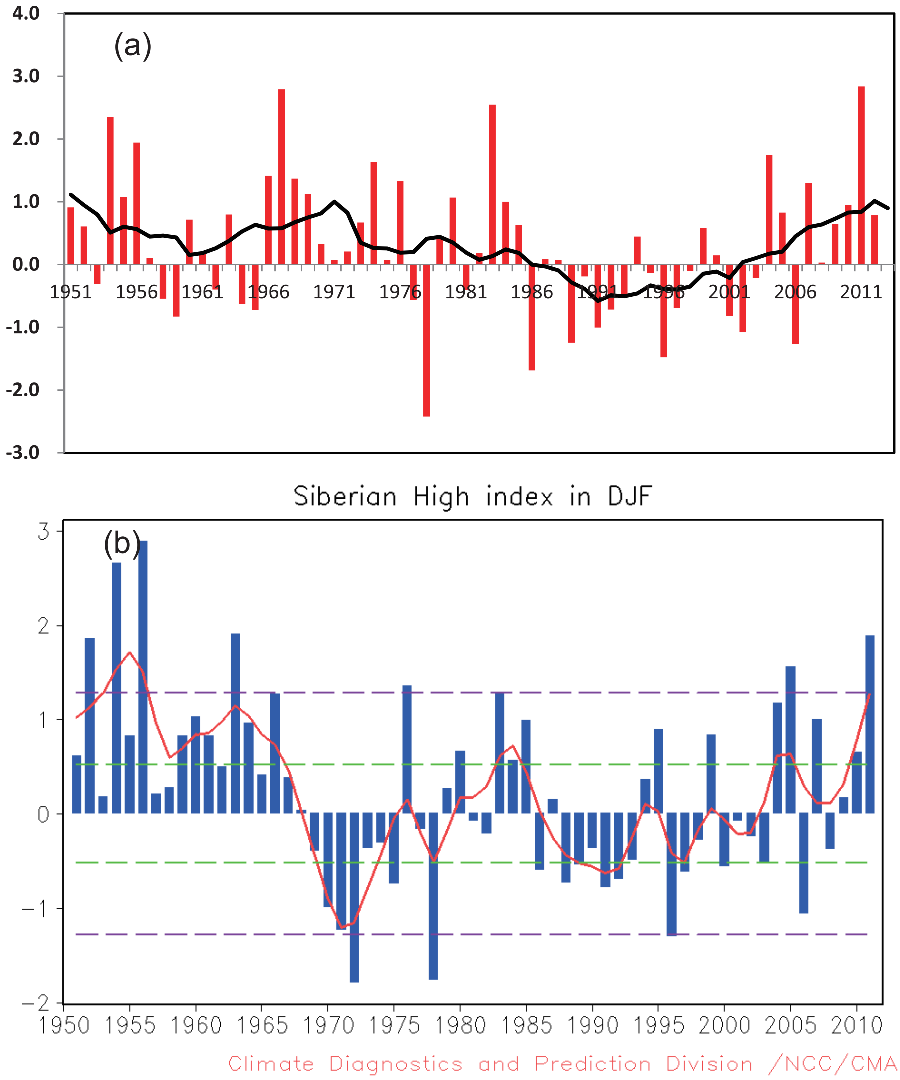

Fig. 9. Temporal variations theEast Asian winter monsoon index and the Siberian high index during 1950/51–2012/13 from (a) the National Climate Center and (b) China Meteorological Administration (http://cmdp.ncc.cma.gov.cn/extreme/lowtemp.php?product=lowtemp_diag). The black solid line in (a) and the red solid line in (b) denote the Gaussian low-pass filtered time series.

fecting quasi-steady planetary wave activity (Wu and Huang, 1999; Gong et al., 2001; Chen and Kang, 2006; Wang et al., 2009b). Wei and Lin (2009) further pointed out that, when the NAM/AO is in negative phase, the East Asian winter monsoon strengthens, usually along with a strong outbreak of cold air; otherwise, when in positive phase, the circulation over East Asia is more zonal and the East Asian winter monsoon becomes weaker.

Based on the above-mentioned results, Liang et al. (2014) produced a schematic diagram to summarize the general features and flow patterns for a cold (strong winter monsoon) and warm (weak winter monsoon) period of PDO (Fig. 11). In cold (warm) periods, the NAM/AO is in negative (positive) phase; high-level zonal wind of the mid and high latitudes weakens (strengthens); the East Asian major trough becomes deeper (more flattened); the Siberian high strengthens (weakens); and the 850 hPa wind field in China in the mid and high latitudes prevails with northerly (southerly) wind anomalies. Meanwhile, the North Pacific is in a negative (positive) phase of PDO; and the anomaly wind field to the east of the Philippines at 850 hPa is cyclonic (anticyclonic). Under such circumstances, northerly (southerly) wind anomalies exist in South China, and such large-scale northerly (southerly) wind anomalies makes it harder for high-latitude cold air to invade into the low latitudes and bring about a relatively colder (warmer) winter for China.

Other studies suggest that the AMO also has an important impact on the East Asian winter monsoon. The AMO's negative phase corresponds to cold periods (stronger winter monsoon), while its positive phase corresponds to warm periods (weaker winter monsoon) (Qu et al., 2006; Li and Bates, 2007; Yan et al., 2008; Wang et al., 2009c; Li et al., 2009).

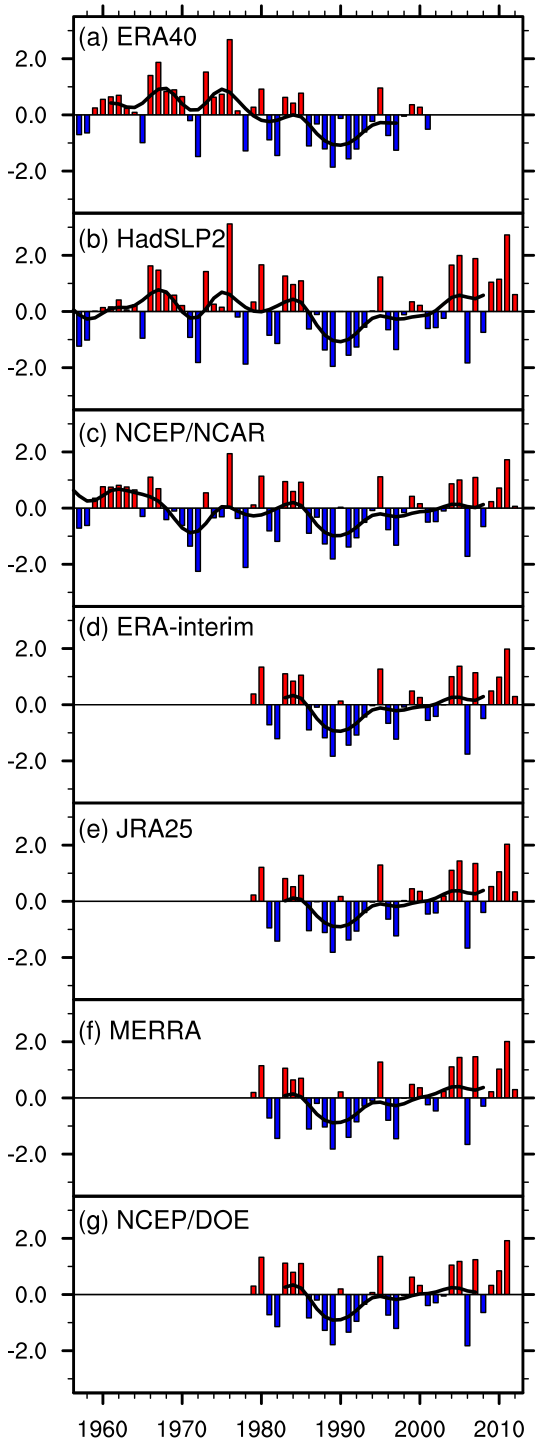

Fig. 10. Normalized winter mean East Asian winter monsoon index based on different data. The black solid lines indicate 9-yr low-pass components using the Lanczos filter. [Reprinted from Wang and Chen (2014)]

Ding et al. (2014a) further point out that the negative and positive phases of the AMO are not completely consistent with winter temperature in China (Fig. 12). Therefore, although the AMO also has an important impact on winter temperature in China, the relative contribution of its impact compared with that of the PDO needs to be further studied.

In addition to the interdecadal variation of the intensity of the East Asian winter monsoon, anthropogenic aerosols are also thought to play a role in enhancing the winter monsoon in the East Asian tropics and subtropics, as well as decreased winter rainfall in southeastern China (Huang et al., 2013).

4. Global monsoon and its link to global climate change

4.1. *Emergence of the concept of the global monsoon*

Traditionally, the regional monsoon climate is characterized by an annual reversal of prevailing surface winds and by the contrast between a rainy summer and dry winter (Ramage, 1971; Webster et al., 1998). It has long been recognized that land–sea thermal contrast is the main driving force for the seasonal shift of regional wind and precipitation fields. However, a key fact is that all regional monsoons are eventually driven and synchronized or coordinated by the annual cycle of solar radiation, and they are bonded or connected by the global divergent circulation, a global-scale persistent overturning of the atmosphere in the tropics and subtropics that varies with the time of year. In essence, the monsoon arises from and manifests as a response of the coupled atmosphere–land–ocean system to annual variation of solar forcing. Therefore, the regional monsoon should not be studied in isolation (Trenberth et al., 2000; Wang et al., 2006). Another reason for examining the global perspective of monsoon and its variability is to detect and attribute the possible impacts of global warming on recent monsoonal changes. As Wang and Ding (2006) pointed out, the literature focuses mostly on specific regions of the world and uses different measures of monsoon strength. Considerable uncertainty exists regarding whether or not global warming has affected monsoon variability (e.g. Kripalani et al., 2003). Bearing in mind the physical principles of the conservation of mass, moisture and energy as it applies to the global atmosphere, and the exchange of energy with underlying surfaces, analysis of overall monsoon variability and change from a global perspective is desirable and advantageous for understanding fundamental monsoon dynamics. Therefore, the global monsoon is an emerging concept (Trenberth et al., 2000; Wang and Ding, 2006). However, the terminology and the related methodology with the global monsoon concept have been accepted increasingly (IPCC, 2013), with its adequate definition, precipitation domain, spatial and temporal modes of variation, strength and relationship to seven regional monsoons (Indian monsoon, East Asian monsoon, Maritime Continent, Australian monsoon, western North Pacific monsoon, North American monsoon, and South American monsoon). With the emergence of the global monsoon, one possible question relates to the American monsoon system, which has not been included in previous definitions of monsoon areas in terms of the seasonal reversal in wind and precipitation. However, from the precipitation changes and divergent wind circulation, the monsoon regions should include the Americas, although the picture is complicated by the meridional

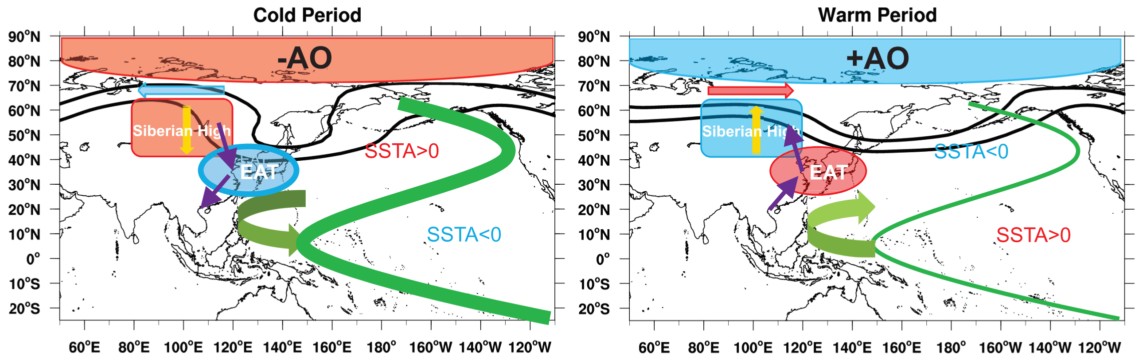

Fig. 11. Schematic diagrams depicting the characteristics and circulation patterns during (a) cold and (b) warm periods. The black thick lines represent the 5200 gpm (northern) and 5300 gpm (southern) contours of geopotential height. Shaded ellipses represent the East Asian trough, with red (blue) shading indicating higher (lower) geopotential height values relative to the climatological mean. Shaded boxes represent the Siberian High, with red (blue) indicating higher (lower) values relative to the climatology. The blue arrow in panel (a) and the red arrow in panel (b) indicate the directions of the mean zonal wind anomalies. Purple arrows represent wind anomalies at 850 hPa. The yellow arrows represent anomalies in vertical velocity, with the downward (upward) arrow indicating stronger (weaker) ascending (descending) motion. The green contour separates positive SST anomalies from negative ones. The green curved arrow indicates the anomalous 850 hPa cyclone (anticyclone) during the cold (warm) period. Red (blue) shading in the polar region indicates positive (negative) sea level pressure anomalies, corresponding to a negative (positive) NAM/AO index. [Reprinted from Liang et al. (2014)]

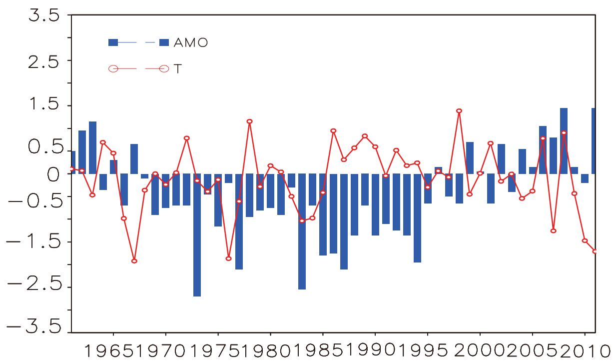

Fig. 12. Detrended time series of winter mean temperature anomalies (red lines; units: °C) and winter AMO index (blue bars, units: °C) from 1961 to 2010. The AMO index is calculated as the SST anomaly averaged over the North Atlantic (0°–60°N, 0°–80°W). [Reprinted from Ding et al. (2014a)]

sectors for North and South America being somewhat offset. Nevertheless, these monsoon regions clearly show a monsoonal (direct) overturning that varies seasonally (Webster et al., 1998; Trenberth et al., 2006).

In order to quantitatively delineate the global monsoon, three fundamental characteristic features are defined by Wang and Ding (2006) and Wang et al. (2006): (1) the primary climatological feature of the tropical precipitation and low-level circulation [global monsoon index (GMI)]; (2) the global monsoon domain or area (GMA); and (3) monsoon precipitation intensity. The first feature can be depicted well by three parameter-metrics: the annual mean, a solstitial mode, and an equinoctial (spring–fall) asymmetric mode (Wang and Ding, 2008). The solstitial mode can be depicted by the June–September mean minus the December–March mean. The equinoctial asymmetric mode can be approximately depicted by the difference between the April–May mean and the October–November mean. Both modes depict an annual cycle and together they account for 84% of the annual variance. The global monsoon precipitation domain can be delineated by the region in which the annual range exceeds 180 mm and the local summer monsoon precipitation exceeds 35% of annual rainfall (Wang and Ding, 2006). The first criterion distinguishes monsoonal from semi-arid and Mediterranean (trade wind) climate, while the second identifies the concentrated summer rain region (Fig. 13). It is interesting to note that this simplistic definition agrees very well with the regional monsoon domains previously defined based on various indices, but the global monsoon domain (Fig. 13a) differs and is complementary to the global mean precipitation (Fig. 13b), which tends to be maximized at the equator and is generally more equatorially symmetric. Wang and Ding (2006) used three complementary methods to measure the monsoon precipitation intensity. Among them, the GMI can more clearly quantify the strength of the global mean monsoon, which is defined by the sum of Northern-Hemisphere-averaged June–July–August (JJA) monsoon precipitation (NHMI) (i.e. the precipitation falling in the Northern Hemisphere monsoon domain) and Southern-Hemisphere-averaged December–January–February (DJF) monsoon precipitation (SHMI). Figure 14 shows the time series of GMI, NHMI and SHMI. The ensemble mean time series indicates a decreasing trend in the GMI across the entire 56 years for the global land area, particularly before 1980 and for the Northern Hemisphere. The SHMI, however, shows no significant trend. The combined global ocean–land monsoon precipitation intensified during 1979–2008, mainly due to an upward trend in the Northern Hemisphere sum-

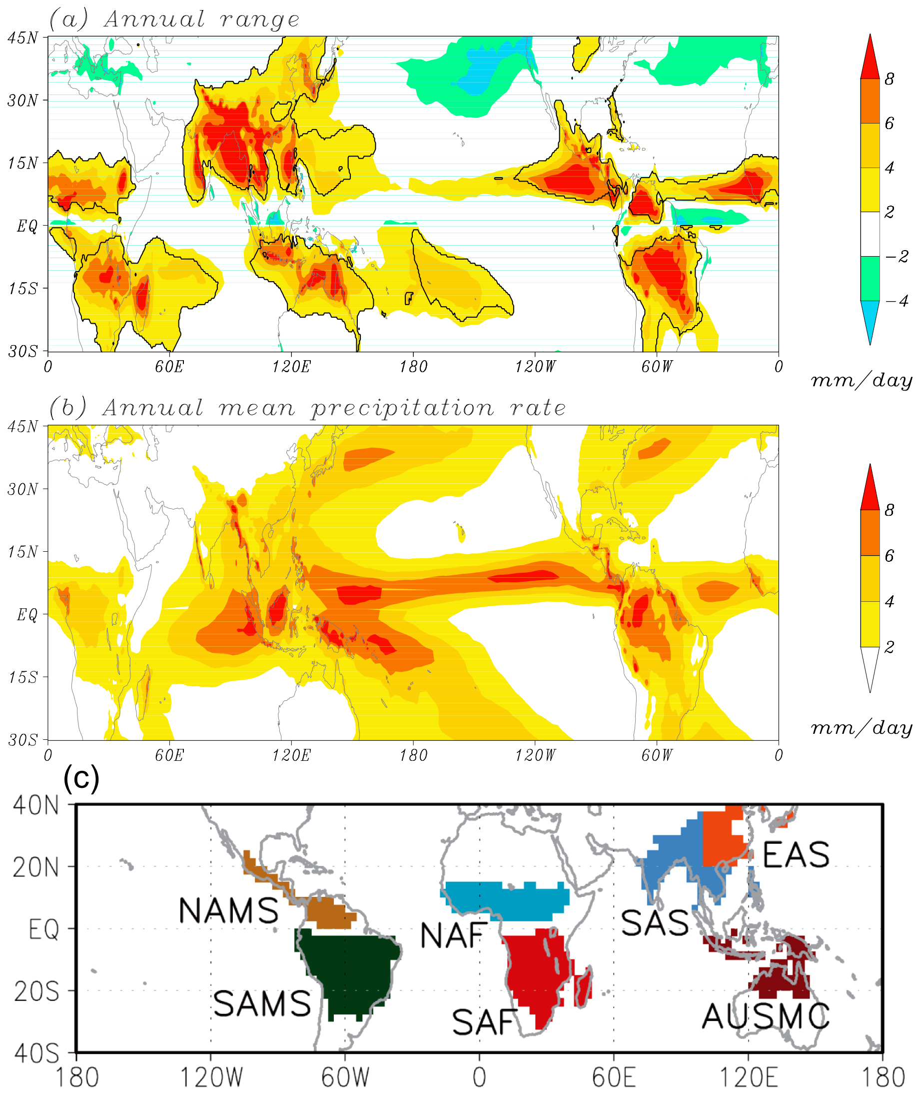

Fig. 13. (a) The climatological mean for the annual range of precipitation, defined by the local summer mean precipitation rate (JJA in the Northern Hemisphere and DJF in the Southern Hemisphere) minus the local winter mean precipitation rate. The bold lines delineate the global monsoon domain. (b) The long-term mean for total annual precipitation. The data used are blended GPCP data (1979–2003) and the means of four precipitation datasets (described in the text) for the period 1948–2003 [Reprinted from Wang and Ding (2006)]. (c) Regional land monsoon domain based on the mean precipitation of 26 CMIP5 models with a common 2.5° × 2.5° grid in the present day (1986–2005). For regional divisions, the equator separates the northern monsoon domains [North American monsoon system (NAMS), North Africa (NAF), southern Asia (SAS) and East Asian summer (EAS)] from the southern monsoon domains [South American monsoon system (SAMS), South Africa (SAF), and Australian–Maritime Continent (AUSMC)], 60°E separates NAF from SAS, and 20°N and 100°E separates SAS from EAS. All the regional domains are within 40°S to 40°N. [Reprinted from Christensen et al. (2013b)]

mer oceanic monsoon precipitation (Fig. 14b) (Wang and Ding, 2006; Zhou et al., 2008; Hsu et al., 2011, Wang et al., 2012; IPCC, 2013). Recently, Lin et al. (2014) further evaluated the changes in global monsoon precipitation based on five kinds of reanalysis datasets. These datasets document well the increasing tendency during 1979–2011, along with a strong interannual variability. This observed increasing trend of global monsoon precipitation is dominated by contributions from the Asian, North American, southern African and Australian monsoon.

CMIP5 models generally reproduce the observed global monsoon domain and global monsoon intensity, but the difference between the best and poorest models is large (Kim et al., 2008; Zhou et al., 2008; Hsu et al., 2011; Wang et al., 2012) (Fig. 15). Their projections suggest that the GMA, global monsoon total precipitation, and GMI will increase by

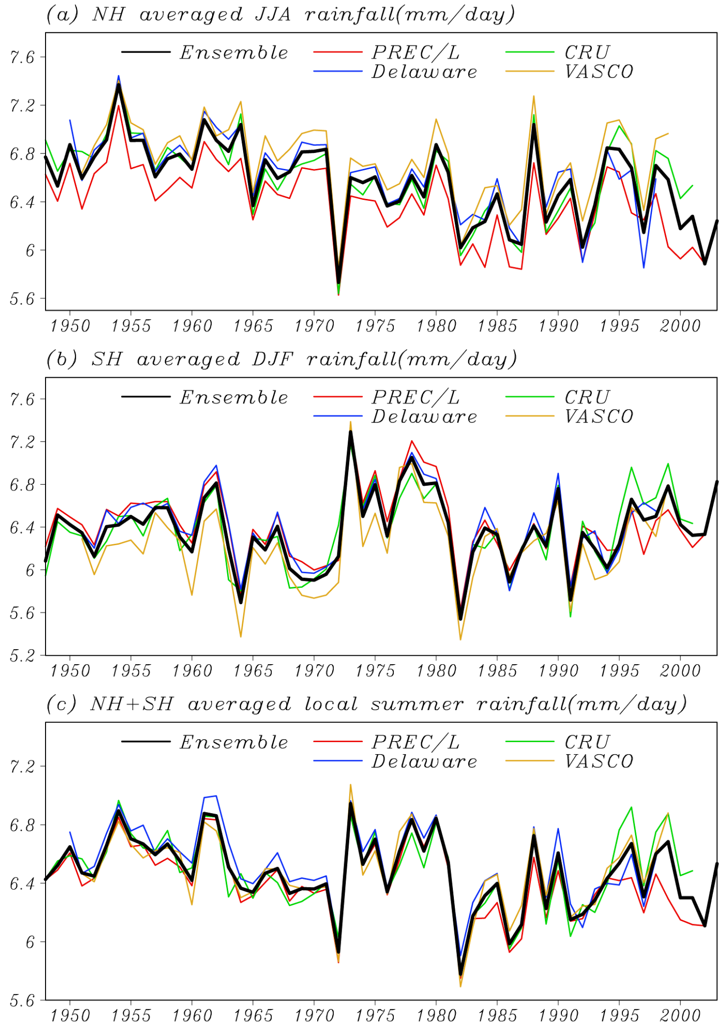

Fig. 14. Time series of the (a) Northern-Hemisphere-averaged JJA precipitation, (b) Southern-Hemisphere-averaged DJF precipitation, and (c) global monsoon index (GMI), or the sum of Figs. 1a and b. (d) The GPCP GMI over land and ocean, respectively. Also shown is the GMI over land regions, derived from ensemble land-based rain-gauge data. [Reprinted from Wang and Ding (2006)]

the end of the 21st century (IPCC, 2013). Indices of precipitation extremes, such as simple daily precipitation intensity index, annual maximum 5-day precipitation total, and consecutive dry days, all indicate that intensive precipitation will increase at higher rates than those of mean precipitation.

4.2. *Influence of climate change on the global monsoon*

An interesting problem is how climate change affects the global monsoon. Based on recent studies (Kim et al., 2008; Zhou et al., 2008; Liu et al., 2009), the major findings in this respect can be summarized as follows:

First, over the past 1000 years before the industrial period, the natural variation in the total amount of effective solar radiative forcing may have played a dominant role (Figure omitted), with simulated global monsoon precipitation being weak during the Little Ice Age (1450–1850) and strong in the Medieval Warm Period (ca. 1030–1240). These results reinforce the importance of the thermal contrast between the ocean and land, as well as between the Northern and Southern Hemisphere, resulting in the millennial-scale variation, quasi-bicentennial oscillation, and interdecadal oscillations (mainly 74-yr oscillation) in the GMI. On the global scale, there have been no integrated observations to verify the above model results; however, the 200-yr oscillation has been noted empirically at many individual sites, and the period of 74 years is observationally very significant (60–80-yr) in the Asian monsoon region for temperature and precipitation (Ding et al., 2008; Liu et al., 2009). From Figs. 16e and f, we can see the almost consistent change in the global monsoon strength and global mean surface temperature. The correlation coefficient between the 31-yr running-mean GMI and global mean temperature that represents the most rapid warming period in the past 100 years is 0.89. However, the global monsoon rainfall follows effective radiative forcing

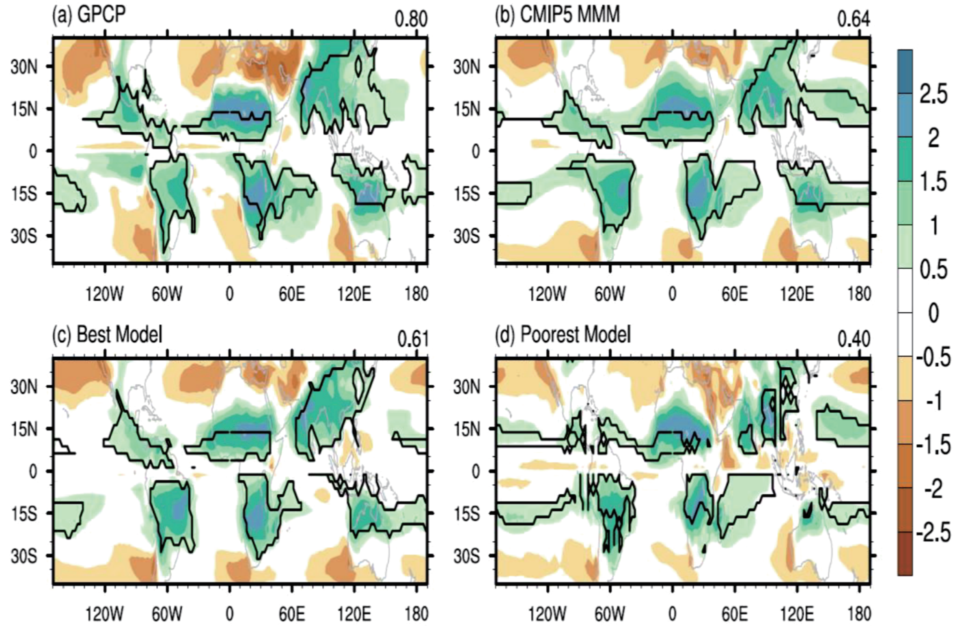

Fig. 15. Monsoon precipitation intensity (shading, dimensionless) and monsoon precipitation domain (lines) for (a) observation-based estimates from GPCP, (b) CMIP5 multi-model mean, (c) the best model and (d) the worst model in terms of the threat score for this diagnostic. These measures are based on the seasonal range of precipitation using hemispheric summer (May–September in the Northern Hemisphere) minus winter (November–March in the Northern Hemisphere) values. The monsoon precipitation domain is defined where the annual range is > 2.5 mm d^{-1}, and the monsoon precipitation intensity is the seasonal range divided by the annual mean. The threat scores (Wilks, 1995) indicate how well the models represent the monsoon precipitation domain compared to the GPCP data. The threat score in panel (a) is between GPCP and CMAP (Climate Prediction Center Merged Analysis of Precipitation) rainfall to indicate observational uncertainty, whereas in the other panels it is between the simulations and the GPCP observational dataset. A threat score of 1.0 would indicate perfect agreement between the two datasets. [Reprinted from Flato et al. (2013)]

more effectively than global mean temperature does, especially in the Little Ice Age. The global monsoon is essentially driven by the interhemispheric temperature and differential heating between the Northern and Southern Hemisphere. It can be seen from Figs. 16f and g that GMI varies obviously with the interhemispheric temperature difference. The interhemispheric temperature difference is related to solar activity change and land–ocean distribution. In response to increased radiative forcing, the Northern Hemisphere, with its greater proportion of land, warms up more than the Southern Hemisphere. As such, the difference between the temperature of the two hemispheres should vary in concert with the effective radiative forcing, and is thus correlated with global monsoon precipitation change. The prominent upward trend in global monsoon precipitation in the last century and the notable strengthening of the global monsoon in the 30 years from 1961 to 1990 were possibly due in part to the increase in atmospheric CO_2 concentration. It is worth noting that the simulated change in global monsoon in the 30 years from 1961 to 1990 has a spatial pattern different from that during the Medieval Warm Period, suggesting that global warming caused by increased GHGs and natural variation of solar forcing have different effects on the characteristics of global monsoon precipitation.

Secondly, another natural external forcing is volcano eruptions. Episodes of volcanic forcing change from year to year, but spectral analysis confirms peaks at 192 and 107 years, respectively. It is uncertain what gives rise to this periodicity, as volcanic eruptions have always been regarded as being chaotic and unpredictable. However, based on an analysis of 21 CMIP3 coupled general circulation models (CGCMs) (Kim et al., 2008), the trends between CGCMs with and without the inclusion of volcanic forcing are statistically different at the $> 90\%$ confidence level based on the t-test, with most CGCMs including volcanic forcing showing a negative NHMI trend for the period 1951–99, and more than half of the CGCMs without volcanic forcing depicting increased rainfall over the Northern Hemisphere land monsoon regions. A possible explanation is the occurrence of abnormally large eruptions after the 1960s, leading to an abrupt increase in stratospheric aerosols.

Thirdly, there is the factor of coupled oceanic forcing. The ocean can be viewed as an internal coupled forcing in the climate system through air–sea interaction. Global warming most likely favors more abundant water vapor in the atmosphere through evaporation due to increased SST, thereby intensifying precipitation over the ocean and even over the land. Wang and Ding (2006) revealed that observed oceanic monsoon precipitation shows increased wetness after 1980 in concert with the rapid global warming (see Fig. 14b). Later, based on the definition of the global ocean monsoon index (GOMI) which is defined by the local summer precipitation fallen into the global oceanic monsoon domain, Kim et al. (2008) obtained an observed increased wetness with a slope of 0.37 mm d^{-1} (20 yr)$^{-1}$ for the period 1980–99. Their study further showed that most CMIP3 CGCMs are able to capture the positive trend of the time series in terms of simulated GOMI. Recently, remote sensing and precipitation observations from the Global Precipitation Climatology Project (GPCP) for 1879–2003 have been used to reconstruct oceanic precipitation for 1900–2008 (75°S–75°N) by employing a statistical technique that makes use of the correlation between precipitation and both SST and sea level pressure (Smith et al., 2009, 2012). The trend from 1900 to 2008 is 1.5 mm month^{-1} (100 yr)$^{-1}$. For the period of overlap, the reconstructed global ocean mean precipitation time series show consistent variability with GPCP. Focusing on the tropical ocean (25°S–25°N) for the recent period of 1979–2005, Gu et al. (2007) identified a precipitation trend of 0.06 mm d^{-1} (10 yr)$^{-1}$ using GPCP data, which is smaller than the trend for 1980–99.

As indicated above, studies of global monsoon precipitation over land reveal an overall weakening over the recent half-century (1950–2000). The study by Zhou et al. (2008) suggested that this significant weakening trend in the global land monsoon precipitation was caused by the warming trend over the central-eastern Pacific and western tropical Indian Ocean. At the interannual scale, global land monsoon precipitation is closely correlated with ENSO. The simulated interannual variation of the global land monsoon index matches well with the observation, indicating that most monsoon precipitation variation arises from ocean forcing. The observed decreasing trend in the NHMI index can be found in the CAM2 simulation. It is, however, slightly weaker than observed. The simulated trend is -0.36 mm d^{-1} (50 yr)$^{-1}$, while the observed decreasing magnitude is -0.59 mm d^{-1} (50 yr)$^{-1}$. The simulated decreasing trend of GMI is comparable to the observation, with a trend of -0.21 mm d^{-1} (50 yr)$^{-1}$ versus the observed value of -0.34 mm d^{-1} (50 yr)$^{-1}$. The simulated SHMI index shows an apparent trend that is consistent with the observation. These results show that the decreasing trend of GMI over land in the past 50 years has oceanic origins.

5. Summary

During the three decades plus since MONEX in 1978–79, substantial achievements have been made in the area of monsoon meteorology. As a significant thrust, MONEX heralded a new era with new findings, concepts and breakthroughs made in monsoon research and prediction. This review paper highlights the most important aspects that have changed or greatly influenced our understanding and the concepts of the monsoon system, which is an extremely important component of the global climate system. In summary, the following conclusions can be drawn:

(1) Compared to the pre-MONEX era, the monsoon system is now viewed as a response of the coupled ocean–land–atmosphere–cryosphere system under the annual cycle of solar heating.

(2) As global climate change or global warming has progressed, the driving force of the monsoon system has changed, with effects of anthropogenic factors (emission of greenhouse gases and aerosols) ever increasingly emerged and other factors to a certain extent altered (ocean and land processes), thus leading to multiple-scale variation in the mean state and variability of monsoons through various thermodynamic processes and circulation changes.

(3) Much observational and model-derived evidence has documented well the changes in the onset, precipitation and circulation modes of the global and/or regional monsoon systems. Global monsoon precipitation over land has decreased since 1960, whereas the global combined ocean and land monsoon precipitation has increased during the most recent three decades. The onset date of the Asian summer monsoon has become earlier during the last 20 years. Although the interannual variability dominated by the monsoon–ENSO relationship has remained largely unchanged, the interdecadal variability of monsoons (especially the Asian summer monsoon) has become more prominent during the last 50–60 years. All of the above evidence demonstrates that we are facing an evolving monsoon system under the impact of global climate change.

(4) The concept of the global monsoon has emerged and its characteristics, driving force, relationship with regional monsoons, and future change have been intensively studied. An advantageous aspect to using this new concept is to provide a better view of the coordinated response of the monsoon as a coupled ocean–land–atmosphere system to global climate change. This global monsoon system is projected to strengthen its global precipitation in the remainder of the 21st century, with an increase in both the area affected and the intensity, while the monsoon circulation will weaken. Generally, monsoon onset dates will become earlier, or not change much, while monsoon decay will become delayed, resulting in a lengthening of the monsoon season.

Acknowledgements. We are grateful to the two anonymous reviewers for their valuable comments and helpful advice. This work was jointly sponsored by the National (Key) Basic Research, Development (973) Program of China (Grant Nos. 2012CB417205 and 2013CB430202), the National Natural Science Foundation of China (Grant No. 41130960) and the China Meteorological Administration Special Public Welfare Research Fund (Grant No. GYHY 201406001).

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