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Global Monsoon Dynamics and Climate Change

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Abstract

This article provides a comprehensive review of the global monsoon that encompasses findings from studies of both modern monsoons and paleomonsoons. We introduce a definition for the global monsoon that incorporates its three-dimensional distribution and ultimate causes, emphasizing the direct drive of seasonal pressure system changes on monsoon circulation and depicting the intensity in terms of both circulation and precipitation. We explore the global monsoon climate changes across a wide range of timescales from tectonic to intraseasonal. Common features of the global monsoon are global homogeneity, regional diversity, seasonality, quasi-periodicity, irregularity, instability, and asynchroneity. We emphasize the importance of solar insolation, Earth orbital parameters, underlying surface properties, and land-air-sea interactions for global monsoon dynamics. We discuss the primary driving force of monsoon variability on each timescale and the relationships among dynamics on multiple timescales. Natural processes and anthropogenic impacts are of great significance to the understanding of future global monsoon behavior.

1. INTRODUCTION

Of all the atmospheric circulation systems, monsoons exhibit the most significant seasonal variations. They are a central component of the global climate system and are large enough to influence it (WCRP 2009). Monsoon anomalies may cause droughts, floods, and other extreme weather or climate events. Monsoon circulations are also critical to the global transport of atmospheric energy and water vapor. Monsoon variability has profound societal and economic influences on the regions they affect, where more than 70% of the world's population lives. Thus, monsoon research is essential not only for understanding global atmospheric circulation and climate change, but also for preventing and mitigating disaster and achieving sustainable development.

The concept of the monsoon is a long-standing one in climatology, and both the awareness and the observation of monsoons have a long history. In English, the term "monsoon" is derived from an Arabic word with several possible spellings (e.g., *mausim*, *mausam*, *mausem*, *mawsim*, *mausin*) (Dash 2005), or from the Malayan *monsin* (Pédelaborde 1963); all these words mean "season." The existence of the monsoon was recognized long before it was given this name. In the twenty-third or twenty-second century BC, the Chinese emperor Shun wrote a poem entitled "Southerly Wind":

> Gently blows the southerly wind,
> That eases my people's resentment.
> Timely comes the southerly wind,
> That makes my people's wealth grow.

This is the earliest written record that implies the main features of the East Asian summer monsoon and its importance for the people's livelihood (Zeng 2005). In addition, a poem entitled "Northerly Wind" in the earliest Chinese poetry anthology, *Shi Jing* (English translation: *Book of Odes*), reads,

> Cold blows the northerly wind,
> Thickly falls the snow . . . .
> The northerly wind whistles,
> The snow falls and drifts about the snow white.

This is the earliest text that describes the typical features of the East Asian winter monsoon. Thus, as early as 3,000 years ago, the Chinese were already aware of the East Asian summer and winter monsoons.

In the Middle Ages, people began to document their observations of the monsoon. Arab navigators found alternate prevailing southwesterlies from April to September and northeasterlies from October to March in the oceans between India and East Africa. In 1554, an Arab author, Sidi-Ali, recorded the dates of monsoon onset and retreat at fifty stations in the Indian Ocean (Pédelaborde 1963).

Halley (1686) was the first to propose a theoretical explanation of the Asian monsoon, in terms of surface thermal mechanics. This classic view regards the monsoon as an enormous sea breeze. Hadley (1735) modified Halley's monsoon model by considering the Coriolis force. Voyeikov (1879) linked seasonal precipitation variations to the monsoon. Subsequently, others set out similar views. Particularly, *Monsoons*, edited by Fein & Stephens (1987), provided an overview that included a summary of modern monsoon theory, by Webster; a historical review, by Singh; a review of monsoon physics, by G. Kutzbach; and an assessment of the possible causative role of the northern summer perihelion, by J. Kutzbach. The paleomonsoon has also become a subject of active interest (Fairbridge 1986). Paleomonsoon variability has been studied extensively using marine and terrestrial records as well as numerical modeling—these studies cover, for example, the Indian monsoon (Kutzbach & Otto-Bliesner 1982, Prell 1984, Kutzbach et al. 1989, Prell & Kutzbach 1992, Clemens et al. 1996, Fleitmann et al. 2003, An et al. 2011), the East Asian monsoon (An et al. 1990, 1991a, 2000, 2001; Liu & Ding 1998; Wang et al. 2008c), and African paleoclimate change (Kutzbach 1980, 1981; Sarnthein et al. 1981; Rossignol-Strick 1983; deMenocal 1995; Kutzbach & Liu 1997). In 2007, the Past Global Changes (PAGES) project established a global monsoon working group to bring the modern global monsoon concept to the paleoclimate community (Wang et al. 2012).

The traditional concept of monsoon dynamics is that the difference in the heat capacities of the land and the upper ocean leads to a seasonal wind reversal caused by greatly amplified temperature change over land compared with over ocean. The traditional monsoon concept is regional, and it refers to certain areas in the tropics that exhibit significant seasonal variation in circulation and precipitation. Recently, however, this concept has been extended to the global scale.

Sankar-Rao (1966, 1970) was the first to use the term "global monsoon"; he explored the impact of land-sea thermal contrast and topography on global monsoon circulation. Charney (1969) noted that the monsoon is driven by a seasonal migration of the intertropical convergence zone (ITCZ). Sikka & Gadgil (1980), Chao & Chen (2001), Gadgil (2003), and Wang (2009) also supported this hypothesis. Hoskins & Rodwell (1995) investigated the Asian summer monsoon in a global context, as a dominant feature of global circulation. Trenberth et al. (2000) defined the global monsoon in terms of a persistent global-scale seasonal overturning of atmospheric circulations throughout the tropics and subtropics. Qian (2000) discussed alternating dry/wet states and their relation to a global-scale monsoon between 40°S and 40°N. Wang & Ding (2006) discussed the global monsoon precipitation domain and defined the global monsoon as a dominant mode of annual variation in the tropics (Wang & Ding 2008, Chang et al. 2011), and Nie et al. (2010) investigated monsoon variations by employing a convective quasi-equilibrium for the diabatic effects of moist convection as well as the regional differences. However, their work did not include extratropical monsoons.

Flohn (1951) proposed a new interpretation of the monsoon as a normal shift in planetary circulation zones. In this framework, some tropical regions (e.g., India, Indo-China, and South China), some subtropical regions (e.g., California, Maghreb, and South Africa), and some frigid regions (e.g., the Siberian coastlands, Alaska, northern Canada, Iceland, and northern Norway) can be considered to be monsoonal regions. However, the frigid monsoon has not yet been fully recognized.

Since 1970s, a series of studies has been published on the dynamics of the axisymmetric Hadley circulation (Schneider & Lindzen 1977, Held & Hou 1980, Schneider 1987). The results of these studies demonstrated that in response to axisymmetric diabatic heating, the atmospheric circulation adopts two distinct regimes: thermal equilibrium and angular momentum conservation. In the tropics, where the planetary vorticity is small and the Rossby radius of deformation (Pierrehumbert et al. 2011) is large, a weak forcing may overcome the in situ planetary vorticity and produce a meridional circulation, forming a nonlinear angular momentum conservation regime. However, in the extratropics, where opposite conditions appear, a weak forcing may not generate meridional circulation, and the temperature is in local thermal equilibrium. Lindzen & Hou (1988) demonstrated that if the heating maximum is shifted a few degrees from the equator, the forced Hadley circulation becomes asymmetric about the equator, causing the dramatic development of a rising arm in the summer hemisphere and a sinking arm in the winter hemisphere. This can be explained by the inertial unstable development of the ITCZ on the summer side of the equator (Tomas & Webster 1997, Tomas et al. 1999, Pierrehumbert 2000). Plumb & Hou (1992) further studied the axisymmetric atmospheric response to an off-equator external forcing centered at 25°N and quantified the theory by identifying a threshold forcing for the transition from the thermal equilibrium to the angular momentum conservation regime. They demonstrated that for a forcing smaller than the threshold, the relative vorticity is weak, the absolute vorticity is determined by the planetary part, and the atmospheric response takes the regular linear thermal equilibrium regime. Conversely, for a forcing stronger than the threshold, the relative vorticity is strong, the absolute vorticity is determined by the relative vorticity part, and the atmospheric response takes the angular momentum conservation regime. Because the threshold is not large, the authors suggested that the angular momentum conservation regime could be relevant to real thermally forced tropical flows, most obviously tropical monsoonal flows.

Li & Zeng (2002) proposed a dynamical normalized seasonality (DNS) index to describe the characteristics of global monsoons and presented a three-dimensional distribution. This index is based on the intensity of the normalized seasonality of the wind field and can be used to depict the seasonal cycle and interannual variability of monsoons over different areas; the papers by Li & Zeng (2002) and Li et al. (2010) provide further details about the physical definition. Global monsoons include global tropical and subtropical monsoons at the surface, and global tropospheric and stratospheric monsoons in the vertical. The DNS index is applicable to any general monsoon circulation (Ellis et al. 2004). Li & Zeng (2005) specified that the global tropical monsoon lies between the boreal summer and winter locations of the ITCZ.

The paleoclimate community has also synthesized global monsoon variability from a geological perspective (An 2000, Wang 2009, Clemens et al. 2010, Ziegler et al. 2010b, Caley et al. 2011a, Cheng et al. 2012). Modeling and geological evidence suggest that a megamonsoon likely occurred during the Phanerozoic as a consequence of the Pangaean megacontinent (Kutzbach et al. 1989, Loope et al. 2001). Paleomonsoon evolution during the Cenozoic has been inferred from early Paleocene tropical rainforest fossils in Colorado (Johnson & Ellis 2002) and from the Eocene monsoon forest in central Australia (Greenwood 1996). The paleomonsoon system in East Asia was probably established during the Eocene and became persistent during the late Oligocene to early Miocene, with further development thereafter (Parrish et al. 1982, Zhou 1984, Gu & Renaut 1994, Sun & Wang 2005, Guo et al. 2008, Qiang et al. 2011, An et al. 2014). The paleomonsoon developed later in Africa and India (Kroon et al. 1991, Hilgen et al. 1995, Sepulchre et al. 2006). Wang et al. (2006) and Cheng et al. (2012) have used high-resolution oxygen isotope records from globally distributed speleothems to analyze orbital to millennial paleomonsoon variability on a global scale. In addition, Liu et al. (2003b, 2004) have studied the evolution of global monsoons in the Holocene using a coupled ocean-atmosphere model.

The study of monsoons has thus been enlarged from the local and regional scale to the global scale, from tropical regions to subtropical regions, and from the troposphere to the stratosphere. It has been extended to overlapping multiscale temporal interactions, and from a land-sea thermal contrast mechanism to a land-air-sea mechanism that incorporates interactions linked to the annual solar radiation cycle. Essentially, monsoon circulation is associated with the variation of large-scale pressure gradients in various monsoon regions. We have noticed that surface monsoons in different regions are a global phenomenon closely connected to the human environment. We refer here to the global monsoon, or global monsoons, as an integration of various regional monsoons (Wang et al. 2012) within a global context in terms of similar dynamics and behaviors. Therefore, we propose the following definition of the global monsoon:

> The global monsoon is the significant seasonal variation of three-dimensional planetary-scale atmospheric circulations forced by seasonal pressure system shifts driven jointly by the annual cycle of solar radiative forcing and land-air-sea interactions, and the associated surface climate is characterized by a seasonal reversal of prevailing wind direction and a seasonal alternation of dry and wet conditions.

The article is organized as follows: In Section 2, we outline the distribution and features of the global monsoon. In Section 3, we discuss multiscale monsoon variability from the orbital, millennial, and centennial to the interdecadal, interannual, and intraseasonal timescales. In Section 4, we emphasize the importance of the Tibetan Plateau for the formation and evolution of Asian monsoons. In Section 5, we summarize and provide perspective on the study of the global monsoon.

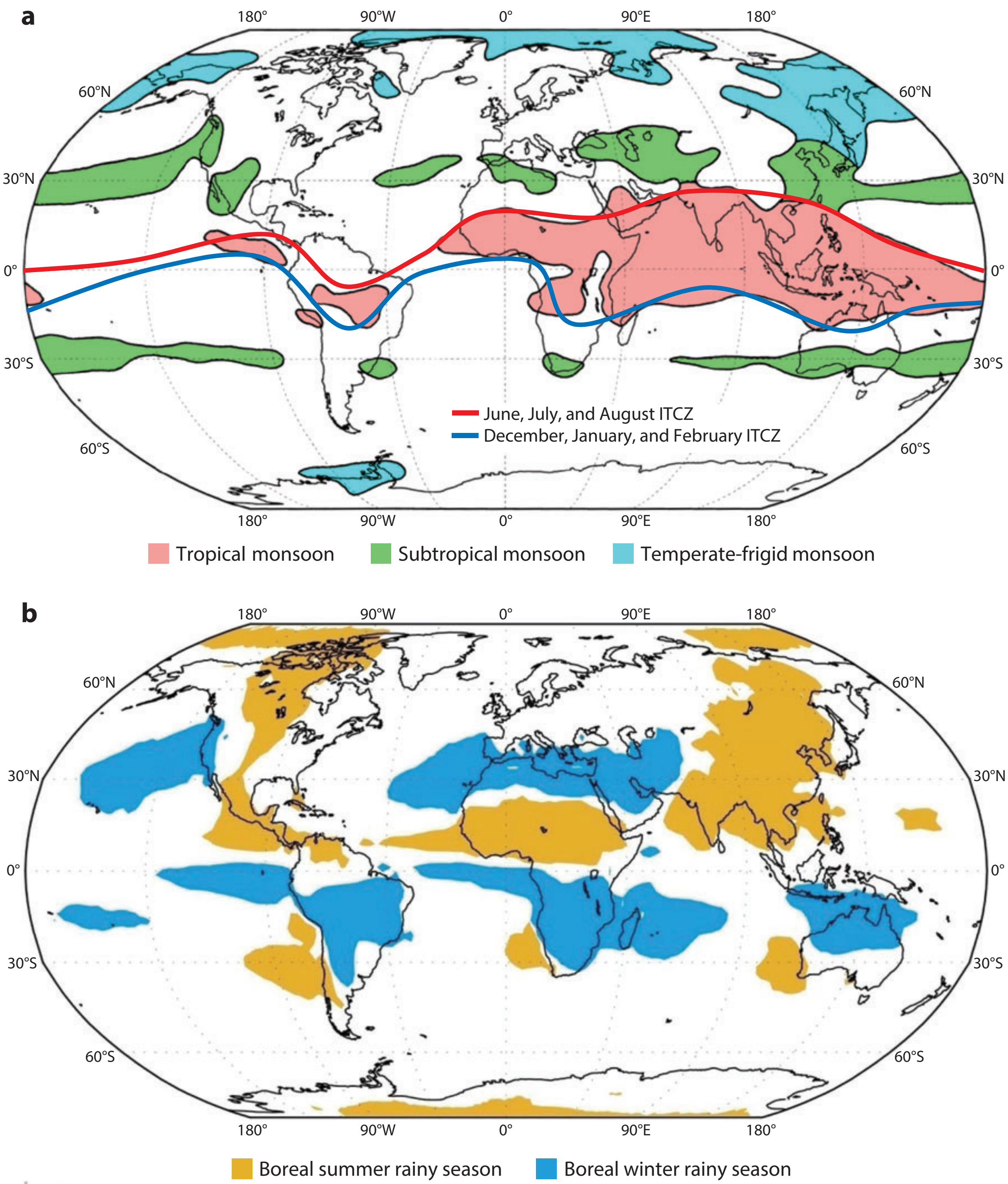
2. DISTRIBUTION AND FEATURES OF THE GLOBAL MONSOON

Here we adopt the DNS index of Li & Zeng (2002) to outline the distribution of the global monsoon (**Figure 1**) and to subdivide the global monsoon according to geographic regions. A large-scale monsoon index can be defined as the averaged DNS values over a monsoon domain. As shown in **Figure 1**, the tropical monsoon essentially lies between the seasonal migration boundaries of the ITCZ (Li & Zeng 2005, Wang 2009). Under the annual cycle of solar radiation, the seasonal migration of the ITCZ, caused by cross-equatorial pressure gradients, produces a strong tropical monsoon (Tomas & Webster 1997, Zeng & Li 2002, Webster & Fasullo 2003). The global tropical monsoon is distributed primarily across the following regions: tropical Asia, Australia, and Africa (Khromov 1957, Ramage 1971); South America (Zhou & Lau 1998); and the tropical eastern Pacific (as well as Central America) (Lau 2003, Li & Zeng 2003). The tropical Asian-Australian monsoon is the archetypal monsoon system and consists of the following main subsystems: the Indian or South Asian monsoon (Goswami et al. 2003), the South China Sea monsoon (Ding & Liu 2001, Feng & Li 2009), the Indochina Peninsula and Western North Pacific monsoon (Tao & Chen 1987, Wang et al. 2008a), the tropical Australian monsoon, and the Maritime Continent monsoon (Webster et al. 1998).

Subtropical monsoons in the Southern and Northern Hemispheres are caused by the seasonal shift of the subtropical high and land-sea distribution and are closely related to large-scale topography, the Rossby radius of deformation, the jet stream, and the interaction between the jet stream and large-scale topography (Molnar et al. 2010). The Northern Hemisphere subtropical monsoon consists of the East Asian (Tao & Chen 1987), North American (Douglas et al. 1993), North African (Khromov 1957, Pédelaborde 1963), and Tibetan Plateau monsoons (Tang & Reiter 1984), as well as the subtropical North Atlantic and North Pacific monsoons (Li & Zeng 2003). The Southern Hemisphere subtropical monsoon includes the Southern Australian (Feng et al. 2010), South African (Khromov 1957, Pédelaborde 1963), and subtropical South Pacific monsoons (Li & Zeng 2003).

Global monsoon systems show a distinct three-dimensional vertical structure that has not been widely considered in the literature. The vertical structure is characterized by significant baroclinicity as well as an obvious asymmetry between the Northern and Southern Hemispheres due to differences in land-sea distribution. In the lower troposphere, the global monsoons include the tropical monsoon and the Southern and Northern Hemisphere subtropical monsoons. In the middle and upper troposphere, only the Southern and Northern Hemisphere subtropical monsoon systems exist, and their extent within the seasonal shift of the subtropical high ridges increases with altitude (see figures 2–4 in Li & Zeng 2000 and figures 3–5 in Li & Zeng 2005). A stratospheric monsoon exists in both the Northern and Southern Hemisphere extratropical regions (Li & Zeng 2000).

Figure 2 illustrates the distributions of semipermanent pressure systems and prevailing circulations in winter and summer, as well as lower tropospheric monsoonal and nonmonsoonal regions. The southwesterly and southeasterly monsoons converge in the Asian-Australian monsoon region (0°–20°N, 140°E–20°W) within the Northern Hemisphere in June, July, and August


a
180°
90°W
0°
90°E
180°
60°N
30°N
0°
30°S
60°S
June, July, and August ITCZ
December, January, and February ITCZ
Tropical monsoon
Subtropical monsoon
Temperate-frigid monsoon
b
Boreal summer rainy season
Boreal winter rainy season

(JJA) and become a northeasterly wind in December, January, and February (DJF), with prevailing northwesterlies in the Southern Hemisphere (0°–10°S, 20°E–160°E). During JJA, the prevailing southeasterlies from Africa, the South Indian Ocean, and Australia (20°S–0°, 20°W–140°E) turn right after crossing the equator and become southwesterlies in tropical North Africa and South Asia. During DJF, the prevailing winds in the above-listed Northern and Southern Hemisphere monsoonal areas reverse. The low-level winds in the subtropical and frigid regions also experience seasonal reversals (**Figure 2**). This feature implies that either the natures of the semipermanent pressure systems associated with these monsoon systems reverse (i.e., a high-pressure system turns into a low-pressure system, and vice versa), or their positions shift considerably.

The seasonal variation in rainfall over the monsoon regions is related to variations in monsoon circulation. To clarify this relationship, we define a normalized seasonal dry-wet index (SDWI), as follows:

$$\mathrm{SDWI} = \frac{R_\mathrm{W} - R_\mathrm{D}}{R_\mathrm{D}} - 2, \tag{1}$$

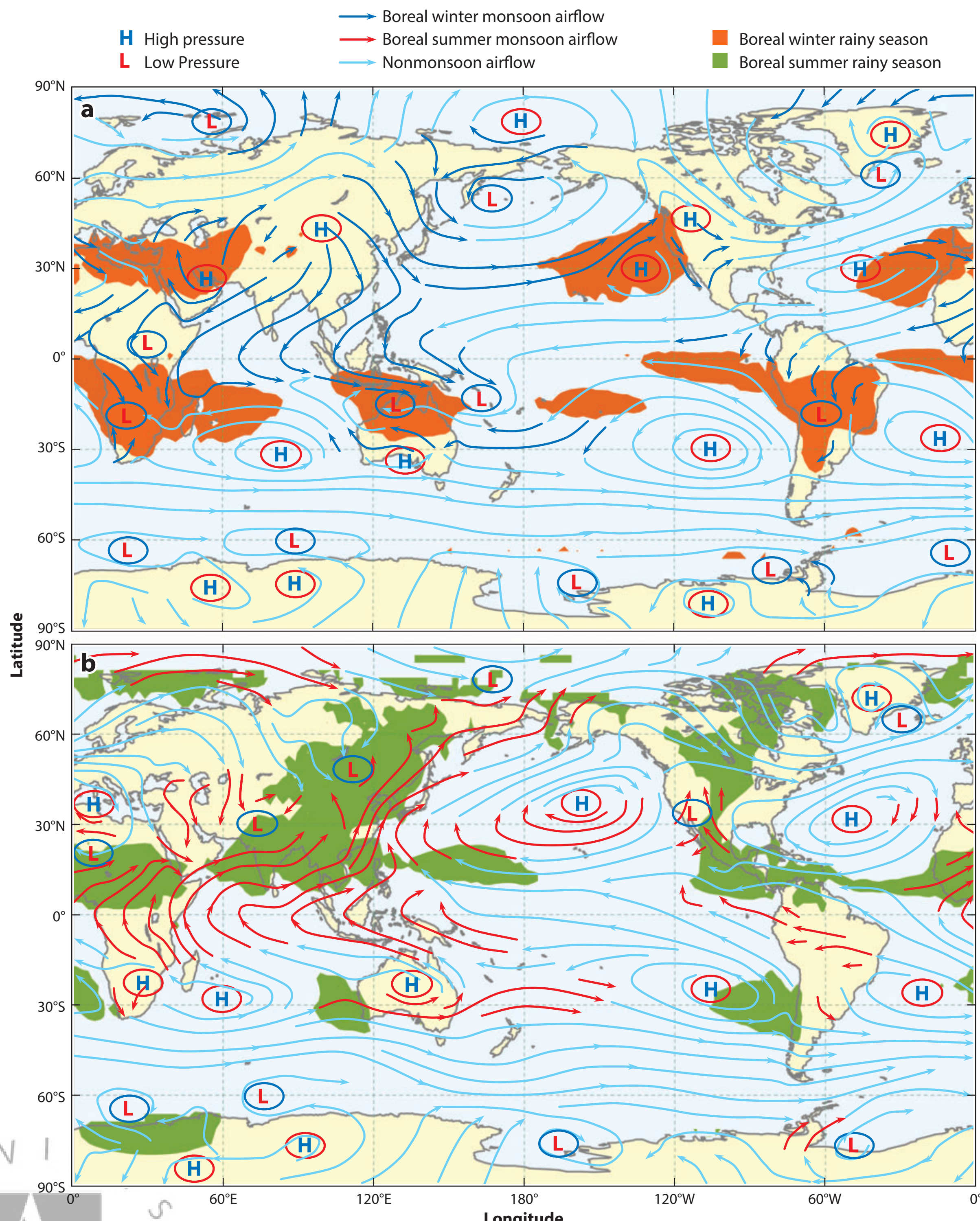
where R_W and R_D indicate the amounts of precipitation in wet and dry seasons, respectively. To simplify, R_W (R_D) here is taken as the seasonal averaged rainfall in JJA (DJF) in the Northern Hemisphere and DJF (JJA) in the Southern Hemisphere. The SDWI, unlike the monsoon precipitation index defined by Wang & Ding (2008), extends the viewpoint to the global and emphasizes the seasonality of rainfall over the monsoonal regions. Domains with positive SDWI occur when the wet season rainfall is at least three times the dry season rainfall, emphasizing the significant seasonal variation of precipitation between dry and wet conditions. This index thus highlights strong seasonality and rainfall concentration. Positive SDWI values in JJA produce patterns that follow most of the Northern Hemisphere monsoon regions and Southern Hemisphere subtropical regions; positive SDWI values in DJF produce patterns that follow most of the Southern Hemisphere tropical monsoon regions and parts of the Northern Hemisphere subtropical regions (**Figure 2**). Overall, the regions with significant seasonal rainfall variation correspond well with the monsoon regions shown in **Figure 1**. Water vapor delivered by monsoon circulation originates primarily from tropical oceans, providing a potential global monsoon link.

3. MULTISCALE MONSOON VARIABILITY

There is a wide range of timescales in monsoon variability. Here we cover monsoon variability on timescales ranging from orbital ($10^5 \sim 10^4$ yr) to intraseasonal scales and address the common features among different monsoon regimes. Orbital to centennial (10^2 yr) variability is mainly inferred from proxies related to monsoonal wind and precipitation in geobiological archives, whereas interdecadal to intraseasonal oscillations are mainly derived from instrumental records. Because monsoons are traditionally described according to geographic associations (e.g., the Indian monsoon, the East Asian monsoon) in historical and published materials, we review monsoon

Figure 1

(*a*) Geographical distribution of global surface monsoon systems. The shading indicates the area where the climatological dynamical normalized seasonality index [National Centers for Environmental Prediction (NCEP) reanalysis data, 1958–2001] is significant. The thick red and blue lines represent the intertropical convergence zone (ITCZ) in boreal summer (June, July, and August) and winter (December, January, and February), respectively. Panel modified with permission from Li & Zeng (2005). (*b*) Distribution of positive seasonal dry-wet index values from Climate Prediction Center Merged Analysis of Precipitation (CMAP) rainfall data, 1979–2008. Yellow shading indicates areas where the boreal summer is the rainy season, and blue shading indicates areas where the boreal winter is the rainy season.



H High pressure
L Low Pressure
Boreal winter monsoon airflow
Boreal summer monsoon airflow
Nonmonsoon airflow
Boreal winter rainy season
Boreal summer rainy season
a
b
90°N
60°N
30°N
0°
30°S
60°S
90°S
Latitude
0°
60°E
120°E
180°
120°W
60°W
0°
Longitude

variability from different geographic regions, with an emphasis on Asian and African monsoon areas.

3.1. Orbital-Scale Monsoon Variability Since the Pliocene

Orbital-scale monsoon variability has been identified from deep-sea, eolian, lacustrine, and speleothem records. Integrating terrestrial and marine monsoon records with modelling results can provide valuable insights into the dynamics of orbital-scale variability.

3.1.1. The Indian summer monsoon. Plio-Pleistocene Indian summer monsoon variations have been reconstructed using biogenic and lithogenic indices from the Arabian Sea (Clemens et al. 1996). A long-term decrease in the lithogenic grain size fraction suggests that the persistent growth of the Northern Hemisphere ice sheets since 3.5 Ma has weakened the intensity of the Indian summer monsoon (**Figure 3**). Since 2.6 Ma, the timing of strong monsoons lagged behind the maximum global ice volume at the precession and obliquity bands by 83° and 124°, respectively. The variability of multiple proxies and associated internal phase relations indicates that the strength of the Indian summer monsoon is sensitive to orbital forcing and the extent of Northern Hemisphere glaciation (Clemens et al. 1996). The phases and timing of multiple monsoon proxies since 350 ka suggest that three mechanisms influence the timing of strong Indian summer monsoons within the precession and obliquity bands: (*a*) sensible heating of the Asian topography, (*b*) glacial boundary conditions, and (*c*) latent heat export from the southern subtropical Indian Ocean (Clemens & Prell 2003).

The phase of the Arabian Sea signals at the precession band was reevaluated by comparing multiple high-resolution productivity and oxygen minimal intensity records from the Arabian Sea with results from a transient climate modeling experiment (Clemens et al. 2010, Ziegler et al. 2010a, Caley et al. 2011b). Three independent proxies (bromine, foraminiferal assemblages, and lithogenic grain size) retrieved in the northern Arabian Sea demonstrated that the productivity and wind strength signals in the northern Arabian Sea were mainly controlled by Indian summer monsoon processes (**Figure 4**). A synthesis of the Indian summer monsoon variability demonstrated similar periodicities and phase lags in the precession and obliquity bands, implying that internal climate forcing might strongly modulate the timing of strong Indian summer monsoons within both the precession and obliquity cycles (Clemens et al. 2010, Caley et al. 2011b).

A high-resolution continental record from the Heqing paleolake indicates the importance of interhemispheric forcing in driving Indian summer monsoon variability at the glacial-interglacial timescale (An et al. 2011). The spectrum of the Heqing proxy records displays primary orbital periods associated with eccentricity (100 kyr) and precession (23 and 19 kyr) (**Table 1**). On the basis of phase and amplitude relationships between the Indian summer monsoon, benthic $\delta^{18}O$, and North Atlantic ice-rafted debris records, Indian summer monsoon variation can be divided into three intervals: 0.92–0.13 Ma, 1.82–0.92 Ma, and 2.60–1.82 Ma (**Figure 3**). During the older and younger intervals, the variations are closely linked to interactions between Northern and Southern Hemisphere dynamic processes, whereas during the middle interval, the northern low is dominant. Change in the Indian summer monsoon at glacial-interglacial timescales is driven by

Figure 2

Circulation patterns (*streamlines*) of the global monsoonal and nonmonsoonal winds at 850 hPa. The areas with positive seasonal dry-wet index values are shaded for (*a*) boreal winter and (*b*) boreal summer. The circulation patterns are modified with permission from Li & Zeng (2005).

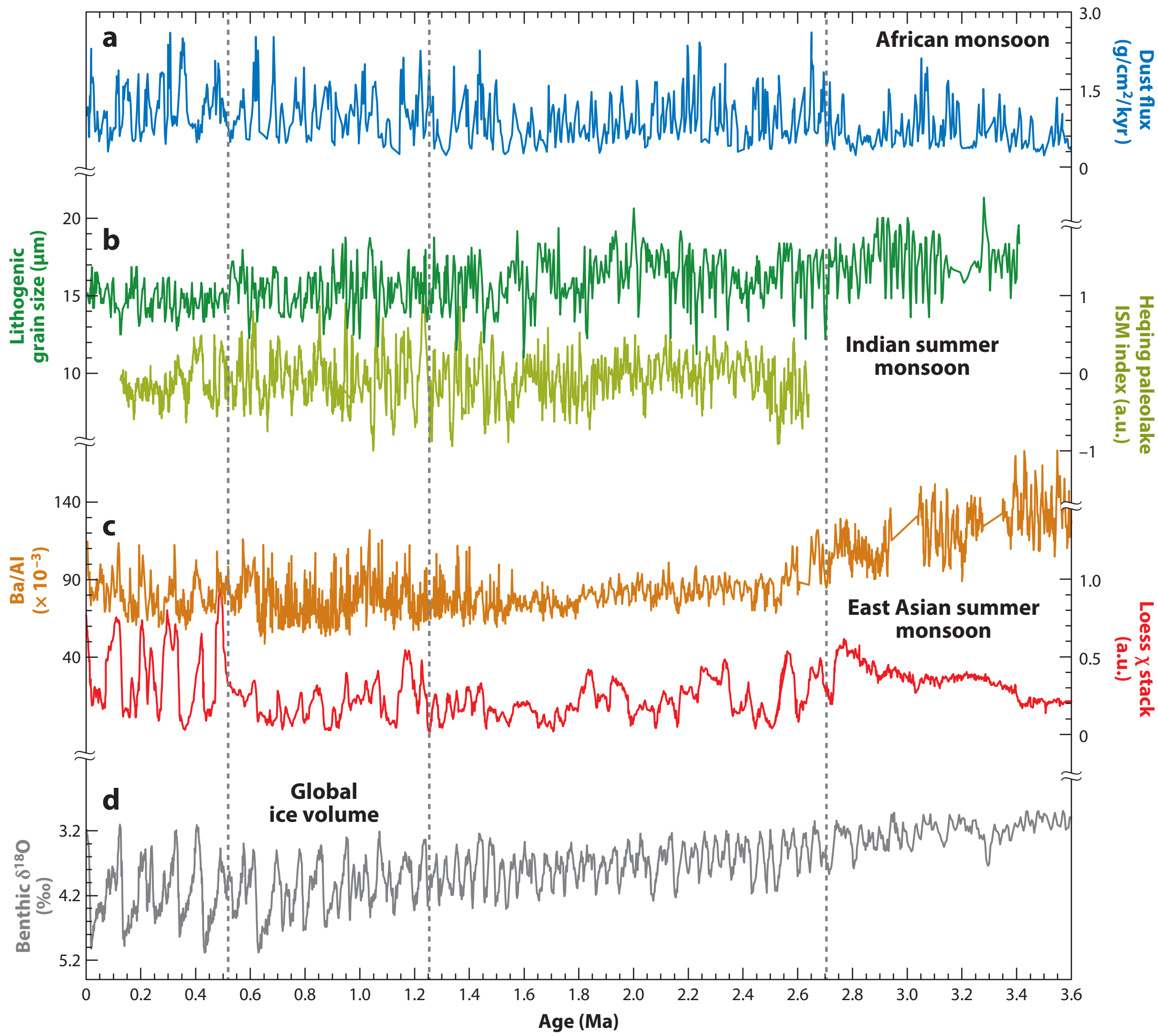


Figure 3

Comparison of monsoon and global ice volume proxies since 3.6 Ma. (*a*) African monsoon proxy: dust flux of Ocean Drilling Program (ODP) site 659 (*blue*, Tiedemann et al. 1994). (*b*) Indian summer monsoon (ISM) proxies: lithogenic grain size of ODP site 722 (*dark green*, Clemens et al. 1996) and ISM index derived from normalized Rb/Sr and total organic carbon content of the Heqing paleolake (*light green*, An et al. 2011). (*c*) East Asian summer monsoon proxies: Ba/Al of ODP site 1146 (*orange*, Clemens et al. 2008) and magnetic susceptibility (χ) stack of the Lingtai and Zhaojiachuan loess sections (*red*, Sun et al. 2006b). (*d*) Global ice volume proxy: marine benthic $\delta^{18}O$ stack (*gray*, Lisiecki & Raymo 2005). Vertical gray dashed lines denote major shifts in these regional monsoon systems.

the relative dominance of both the northern low-pressure and southern high-pressure systems, which is necessary for understanding global monsoon dynamics (An et al. 2011, Caley et al. 2013).

3.1.2. The East Asian summer monsoon. East Asian summer monsoon variations have been reconstructed using land- and marine-based proxies (**Figure 3**) (e.g., An et al. 1990, 1991a,b; Liu

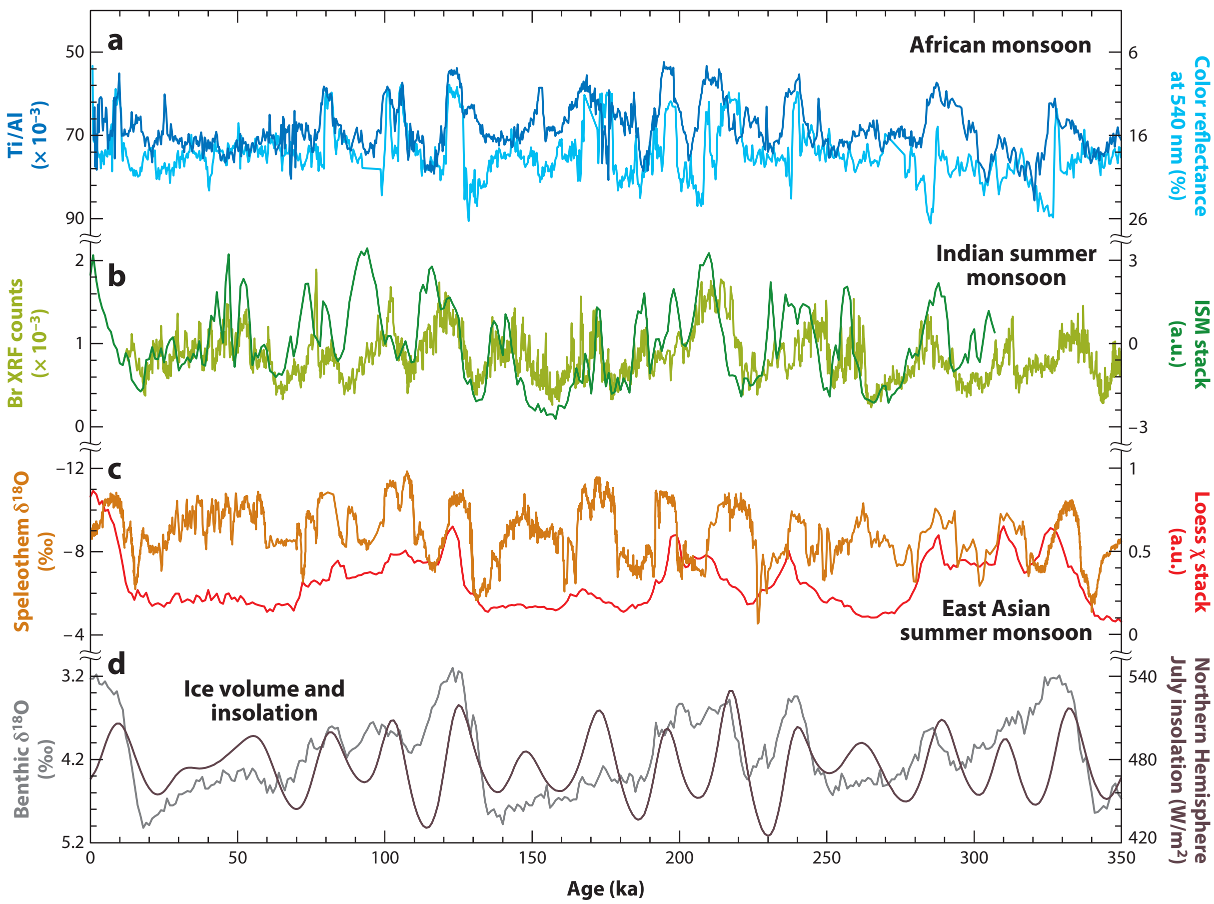


Figure 4

Comparison of monsoon proxies with changes in global ice volume and summer insolation since 350 ka. (*a*) African monsoon proxies: Ti/Al (*dark blue*) and color reflectance (*light blue*) from the Mediterranean Sea (Ziegler et al. 2010b). (*b*) Indian summer monsoon (ISM) proxies: Br X-ray fluorescence (XRF) counts (*light green*) and ISM stack (*dark green*) from the Arabian Sea (Ziegler et al. 2010a, Caley et al. 2011b). (*c*) East Asian summer monsoon proxies: Chinese loess magnetic susceptibility (χ) stack (*red*) and speleothem $\delta^{18}O$ records (*orange*) (Sun et al. 2006a,b; Wang et al. 2008c; Cheng et al. 2009). (*d*) Ice volume and insolation proxies: marine benthic $\delta^{18}O$ stack (*light gray*, Lisiecki & Raymo 2005) and Northern Hemisphere July insolation (*dark gray*, Berger 1978).

& Ding 1998; An 2000; Wang et al. 2005b; Clemens et al. 2008). On the basis of magnetic susceptibility and carbonate content variations in Chinese loess–red clay sequences, the evolutionary process of the East Asian summer monsoon can be divided into three stages: (*a*) an initiation before 3.6 Ma; (*b*) strengthening between approximately 3.6 and 2.7 Ma; (*c*) and large-amplitude fluctuations since 2.7 Ma, followed by two large shifts at approximately 1.2 and 0.5 Ma (An et al. 1990, Xiao & An 1999, Sun et al. 2006b). Interpretation of the Pleistocene South China Sea proxies is complicated by the influences of both monsoon circulations and sea level change on sedimentary geochemistry (Wang et al. 2005a). However, numerous proxies have been employed to infer monsoon-induced oceanic changes (e.g., Wang et al. 1999a, 2003, 2005b). The results suggest a strong East Asian summer monsoon prior to 2.7 Ma, small-amplitude monsoon oscillations between 2.7 and 1.2 Ma, and large-amplitude fluctuations after 1.2 Ma (Clemens et al. 2008).

Table 1 Synthesis of orbital periodicities and Holocene trend of global paleomonsoon variation

Monsoon system	Archive	Orbital periods (kyr)[a]					Non-orbital periods (kyr)	Time range (ka)	Selected references	Chronology	Holocene trend
		~400	~100	~41	~23	~19					
Indian monsoon	Arabian Sea sediments	—		•	•			6–350	Clemens et al. 1991	Benthic $\delta^{18}O$	Weakening
				•	•			0–3,500	Clemens et al. 1996	Benthic $\delta^{18}O$ Biostratigraphy Magnetic polarity	
		—		•	•		66, 16	0–350	Clemens & Prell 2003, Clemens et al. 2010	Benthic $\delta^{18}O$	
		—	•	•	•		33, 11.5	0–450	Ziegler et al. 2010a, Caley et al. 2011b	^{14}C and tuning	
		—		•	•			6–228	Reichart et al. 1998	^{14}C dates Benthic $\delta^{18}O$	
	Heqing paleolake sediments		•	•	•	•	73, 68, 55, 36, 29	0–2,600	An et al. 2011	Astronomical timescale Magnetic polarity	
East Asian monsoon	Chinese loess	•	•	•	•	•	225, 75, 66, 55, 31, 27	0–2,600	Liu et al. 1999, Heslop et al. 2002, Lu et al. 2004, Sun et al. 2006b	Astronomical timescale Magnetic polarity	Weakening
		•		•	•	•	76, 55, 29	1,600–2,600	Sun & An 2004	Astronomical timescale	
			•	•	•	•	72, 55, 29	0–70,000	Sun et al. 2010	Astronomical timescale Magnetostratigraphy	
	South China Sea sediments	—		•	•		67, 53, 35	0–240	Wang et al. 1999a, Jian et al. 2001	Benthic $\delta^{18}O$	
				•	•			2,500–3,200	Wehausen & Brumsack 2002	Astronomical timescale	
			•	•	•	•	11.5	0–5,000	Clemens et al. 2008, Ao et al. 2011	Benthic $\delta^{18}O$ Magnetic polarity Astronomical timescale	
		•	•	•	•			0–250,000	Wang et al. 2003, 2005a	Benthic $\delta^{18}O$ Magnetic polarity Biostratigraphy	
	Chinese speleothem	—			•			0–380	Wang et al. 2008c, Cheng et al. 2009	U-Th dating	

North African monsoon	Tropical Atlantic sediments	•	•	•	•	•		0–5,000	Tiedemann et al. 1994, deMenocal 1995	Astronomical timescale Magnetic polarity	Weakening
	Mediterranean Sea sediments	•	•	•	•	•		0–3,000	Lourens et al. 2001, Larrasoaña et al. 2003	Astronomical timescale Magnetic polarity	
	East Mediter-ranean sapropels	—			•			0–464	Rossignol-Strick 1983	Sapropel and $\delta^{18}O$ chronology	
	Gulf of Guinea sediments	—	•		•			0–155	Weldeab et al. 2007a	^{14}C dates Benthic $\delta^{18}O$	
	Mediterranean Sea sediments	—	•		•			0–350	Ziegler et al. 2010a	Sapropel chronology	
South African monsoon	Lake sediments	—			•			0–200	Partridge et al. 1997	^{14}C dates Sedimentation rate extrapolation	Strengthening
	Buffalo cave speleothem	—		•	•	•	10.7	1,520–1,990	Hopley et al. 2007	Magnetic polarity Astronomical timescale	
Australian monsoon	Banda Sea sediments	—	—	•	•			0–150	Beaufort et al. 2010	^{14}C dates Planktonic-benthic $\delta^{18}O$ correlation	Strengthening
South American monsoon	Brazil speleothem	—	—		•			0–116	Mayle et al. 2000	U-Th dating	Strengthening
	Lake Titicaca sediments	—	—	—	•			0–50	Baker et al. 2001	^{14}C dates	

[a]A dash indicates that the time range of paleoclimate records is too short to identify the corresponding spectral peaks. An empty cell means that the indicated period was not reported in the corresponding selected references. A dot denotes that spectral peaks are evident in the paleoclimate records.

The climate shift at approximately 2.7 Ma was a global phenomenon, and it was particularly dramatic in the monsoonal regions. Both the phase relation and amplitude between different monsoon proxies shifted significantly around that time (An 2000, Ding et al. 2000, Clemens et al. 2008, Sun et al. 2010). At approximately 1.2 and 0.6 Ma, the East Asian summer monsoon was characterized by a two-stage strengthening, with a distinct shift of the dominant 41-kyr periodicity before 1.2 Ma to a dominant 100-kyr periodicity after 0.6 Ma (Liu et al. 1999). Because the East Asian summer monsoon climate is determined by the interactions between the global atmosphere, land, ocean, and ice systems within the specific geographic setting of East Asia, the mid-Pleistocene monsoon shift is likely linked to changing lower boundary conditions—for example, regional tectonic uplift (An et al. 1990, Xiao & An 1999), ice expansion in the high-latitude Northern Hemisphere (Clark et al. 2006, Raymo et al. 2006), and the strong asymmetry of hemispheric climates (Guo et al. 2009). Since 800 ka, a good correlation between loess proxies (magnetic susceptibility and grain size) and benthic $\delta^{18}O$ records suggests that the East Asian monsoon has been strongly coupled to ice volume changes (Ding et al. 1995, Liu et al. 1999).

In contrast, absolute-dated oxygen isotope records from cave speleothems reveal a dominant 23-kyr cycle that lags summer insolation at 65°N by 43°(Wang et al. 2008c, Cheng et al. 2009). The differing expression of orbital-scale monsoon variability between loess magnetic susceptibility and marine records relative to speleothem $\delta^{18}O$ records likely reflects the sensitivity of the different archives to East Asian summer monsoon variation (**Figure 4**) (Clemens et al. 2010, Cheng et al. 2012). Whether loess magnetic susceptibility can record monsoon variability on a precession timescale largely depends on sedimentation rate and postdepositional processes that can attenuate the variability of the loess proxies (Feng et al. 2004, Sun et al. 2006a). Similarly, the effects of summer and winter precipitation, moisture source, and monsoonal wind on speleothem $\delta^{18}O$ variability require further investigation (Wang et al. 2001b, Yuan et al. 2004, Clemens et al. 2010, Pausata et al. 2011, Cheng et al. 2012). Because of these spectral differences—in particular the lack of a distinct precessional signal in Chinese loess and of the 100- and 41-kyr cycles in speleothems—investigators must explore new proxies from high-resolution terrestrial and marine records and their sensitivities to East Asian summer monsoon intensity.

3.1.3. The African monsoon. A characteristic sedimentary expression of changes in African monsoon intensity is the occurrence of cyclic organic-rich layers (sapropels) in the Eastern Mediterranean Sea (Rossignol-Strick 1983, Hilgen et al. 1995). Sapropels S1 to S10 are clearly identifiable in color reflectance records (i.e., percent reflectance at 540 nm) and elemental geochemistry (e.g., Wehausen & Brumsack 2000, Calvert & Fontugne 2001, Lourens et al. 2001, Ziegler et al. 2010b). Wind-borne dust records from deep-sea sediments of the subtropical Atlantic and Mediterranean Sea also provide evidence for monsoon-related Plio-Pleistocene African climate change (e.g., Tiedemann et al. 1994, deMenocal 1995, Larrasoaña et al. 2003).

Marine-based eolian records exhibit significant shifts in flux variability near 2.8, 1.7, and 1 Ma (**Figure 3**). Dust record spectra indicate that before 2.8 Ma, the subtropical African climate varied primarily at precessional periodicity. After 2.8 Ma, marked 41-kyr variability persisted until 1 Ma. The increase at 2.8 Ma in 41-kyr variability coincides with the onset of Northern Hemisphere glaciation, suggesting that the African monsoon was sensitive to remote changes in high-latitude climate (deMenocal & Rind 1993). This sensitivity became further amplified after 1.0 Ma and became characterized by large fluctuations in duration and amplitude of the eolian cycles at a 100-kyr period (deMenocal 1995, Trauth et al. 2009). High-resolution records of bulk sediment geochemistry from the eastern Mediterranean also show distinct 100- and 23-kyr periodicities (**Figure 3**, **Table 1**), implying combined influences of high- and low-latitude processes on African climate (deMenocal & Rind 1993, Ziegler et al. 2010b).

Phase and amplitude spectra of monsoonal proxies reveal that the East African monsoon follows a quasi-direct response to Northern Hemisphere summer insolation with dominant 100- and 23-kyr periodicities (**Table 1**), but the West African monsoon demonstrates distinct lags of several thousand years relative to precession minima and obliquity maxima, with similar weights of precession and obliquity signals (Weldeab et al. 2007a, Caley et al. 2011a). However, the South African monsoon variability inferred from a crater lake is out of phase with the North African monsoon change (Partridge et al. 1997). Transient modeling results that relied only on insolation conditions and on both insolation and ice conditions exhibited an unexpected highly linear response to insolation forcing (Tuenter et al. 2005, Kutzbach et al. 2008, Ziegler et al. 2010b). These simulations indicate that inclusion of variable global ice volume can affect the amplitude of African monsoon strength but has only a small impact in the precession phase (Ziegler et al. 2010b). Different time lags are likely related to the occurrence of precession-paced North Atlantic cold events, which delayed the onset timing of the West and East African monsoon to varying degrees (Weldeab et al. 2007a, Ziegler et al. 2010b).

The dynamics of orbital-scale monsoon variability can be attributed to changes in orbital parameters (eccentricity, obliquity, and precession), surface boundary conditions, and atmospheric CO_2 concentration (e.g., Kutzbach & Otto-Bliesner 1982; Prell & Kutzbach 1987, 1992). The orbital parameters can lead to seasonal changes in solar insolation received at Earth's surface (Berger 1978). The effects of orbital forcing on monsoon strength are due to changes in the amplitude of seasonal insolation and the resulting land-sea thermal and pressure contrasts (Kutzbach & Guetter 1986, Prell & Kutzbach 1987). Obliquity, which varies from 22.1° to 24.5° with a periodicity of approximately 41 kyr, can affect the seasonal variation of incoming insolation in both hemispheres, with a greater effect at high latitudes than at low latitudes (Milanković 1941). Precession, which reflects seasonal changes in the aphelion and perihelion with dominant periods of approximately 23 and 19 kyr, has opposite impacts on insolation changes between the Northern and Southern Hemispheres, with a more significant influence at low latitudes than at high latitudes, resulting in a significant interhemispheric contrast. Changes in net annual insolation can further influence high-latitude sea ice and tropical sea surface temperature (Kutzbach & Gallimore 1988). Therefore, changing obliquity and precession can affect monsoon variability through direct insolation forcing and indirect oceanic feedbacks (Liu et al. 2003b, 2004), as evidenced by distinct obliquity and precession cycles in geological records (**Table 1**). Although mean solar irradiation varies only slightly in response to changes in eccentricity, eccentricity can strongly modulate the precessional amplitude and therefore produce seasonal insolation changes. The eccentricity cycle may have been transmitted into paleomonsoon systems through the nonlinear response of low-latitude carbon cycles (Wang et al. 2003).

Surface boundary conditions and CO_2 changes also affect glacial-interglacial monsoon variability, especially at midlatitudes (Kutzbach & Guetter 1986). The Northern and Southern Hemisphere ice sheets act mainly through pressure and temperature systems, particularly through the cross-equatorial pressure and temperature gradients, to influence the strength of monsoon circulations (Tomas & Webster 1997, An et al. 2011). Variations in greenhouse gases (particularly CO_2) are strongly coupled with global temperature change on orbital timescales and can also influence monsoon variability by affecting temperature and pressure gradients (Felzer et al. 1998, Petit et al. 1999, An et al. 2011). Meanwhile, sea level changes associated with the changing land-ocean distribution and the waxing and waning of ice sheets can lead to shifts in the coastline and marine dynamical conditions. These factors can further influence monsoon variability by changing the land-ocean temperature/pressure contrast and heat/moisture transport from the ocean to the land. Our syntheses indicate that regional monsoons commonly have distinct responses to solar insolation, but that the impacts of obliquity and glacial cycles differ across various locations.

3.2. Millennial-Scale Monsoon Variability Since the Last Glaciation

Suborbital/millennial climate variability is characterized by abrupt changes within decades, followed by stable periods that persist for hundreds to thousands of years (Broecker et al. 1985, Cronin 2009, and references therein). Examples include the Younger Dryas event and Dansgaard–Oeschger cycles identified in the Greenland ice core records (Dansgaard et al. 1993), as well as ice-rafted debris events in the North Atlantic (Heinrich 1988). Impacts of these abrupt events on monsoon variability have been found in both hemispheres (Porter & An 1995, Schulz et al. 1998, Wang et al. 2001b, Altabet et al. 2002, Voelker 2002, Overpeck & Cole 2006, Pisias et al. 2010), with a near-simultaneous pattern in the Northern Hemisphere and opposite changes in the Southern Hemisphere (Augustin et al. 2004, Wolff et al. 2010).

Millennial Indian summer monsoon variability has been observed from multiple proxies studied in Arabian Sea sediments (Sirocko et al. 1993, Overpeck et al. 1996, Schulz et al. 1998, Altabet et al. 2002). These proxies reflect Indian summer monsoon–induced processes such as upwelling, surface productivity, and ventilation within the intermediate waters of intense oxygen-minimum zones. They are well correlated with Dansgaard–Oeschger cycles in Greenland ice cores, where weak monsoon events are associated with cold events in the Northern Hemisphere during the last glaciation (**Figure 5**). On the continent, numerous speleothem $\delta^{18}O$ records show that dry conditions correlate with cold periods in Greenland (Neff et al. 2001, Burns et al. 2003, Fleitmann et al. 2003, Sinha et al. 2005, Cai et al. 2006). These marine and terrestrial records together yield a coherent picture of Indian summer monsoon behavior on millennial timescales during the late Pleistocene.

In East Asia, a strong teleconnection between the East Asian winter monsoon and the North Atlantic climate has been identified by correlation of loess grain size oscillations and ice-rafting events in the North Atlantic (Porter & An 1995, An & Porter 1997), implying that the northern westerlies play an important role in transmitting this temperature signal from the North Atlantic to the Asian monsoon regions. Subsequently, similar correlations have been established based on numerous loess-based monsoonal proxies (Xiao et al. 1995, Guo et al. 1996, Chen et al. 1997, Zhang et al. 1997, Ding et al. 1998, Fang et al. 1999, Wu et al. 2006). Millennial-scale summer monsoon variability has also been inferred from South China Sea sediment records (e.g., Wang et al. 1999b, Oppo & Sun 2005). Recently, integration of loess grain size and speleothem $\delta^{18}O$ records with modeling results has suggested that the Atlantic Meridional Overturning Circulation (AMOC) is likely a driver of abrupt change in East Asian monsoon systems (Sun et al. 2012).

The well-dated speleothem $\delta^{18}O$ record from Hulu cave, eastern China, provides a robust correlation of abrupt monsoon changes with Dansgaard–Oeschger cycles and Heinrich events in high-latitude regions during the last glaciation (Wang et al. 2001b), yielding a steady mechanical connection between North Atlantic events and Asian summer monsoon changes. This correlation—that is, weak East Asian summer monsoons coincident with cold periods in the North Atlantic, and vice versa—has been confirmed by other speleothem records from central and southern China (**Figure 5**) (Yuan et al. 2004, Dykoski et al. 2005, Cheng et al. 2006, Kelly et al. 2006, Wang et al. 2008c). Moreover, lake records reveal an antiphase relationship between the Asian summer monsoon and westerlies on millennial timescales, indicating the interplay of these two systems at Lake Qinghai (An et al. 2012).

The interpretation of the speleothem $\delta^{18}O$ in China is still under debate. A few researchers have suggested that the cave $\delta^{18}O$ in eastern China reflects changes in moisture transport and precipitation upstream over the Indian Ocean and the Indian summer monsoon region (Maher 2008, Dayem et al. 2010, Pausata et al. 2011), whereas more recent modeling results indicate that it does provide a proxy for East Asian summer monsoon intensity (Liu et al. 2014). It is not currently feasible to elucidate the relative contributions of local rainfall, various moisture sources,

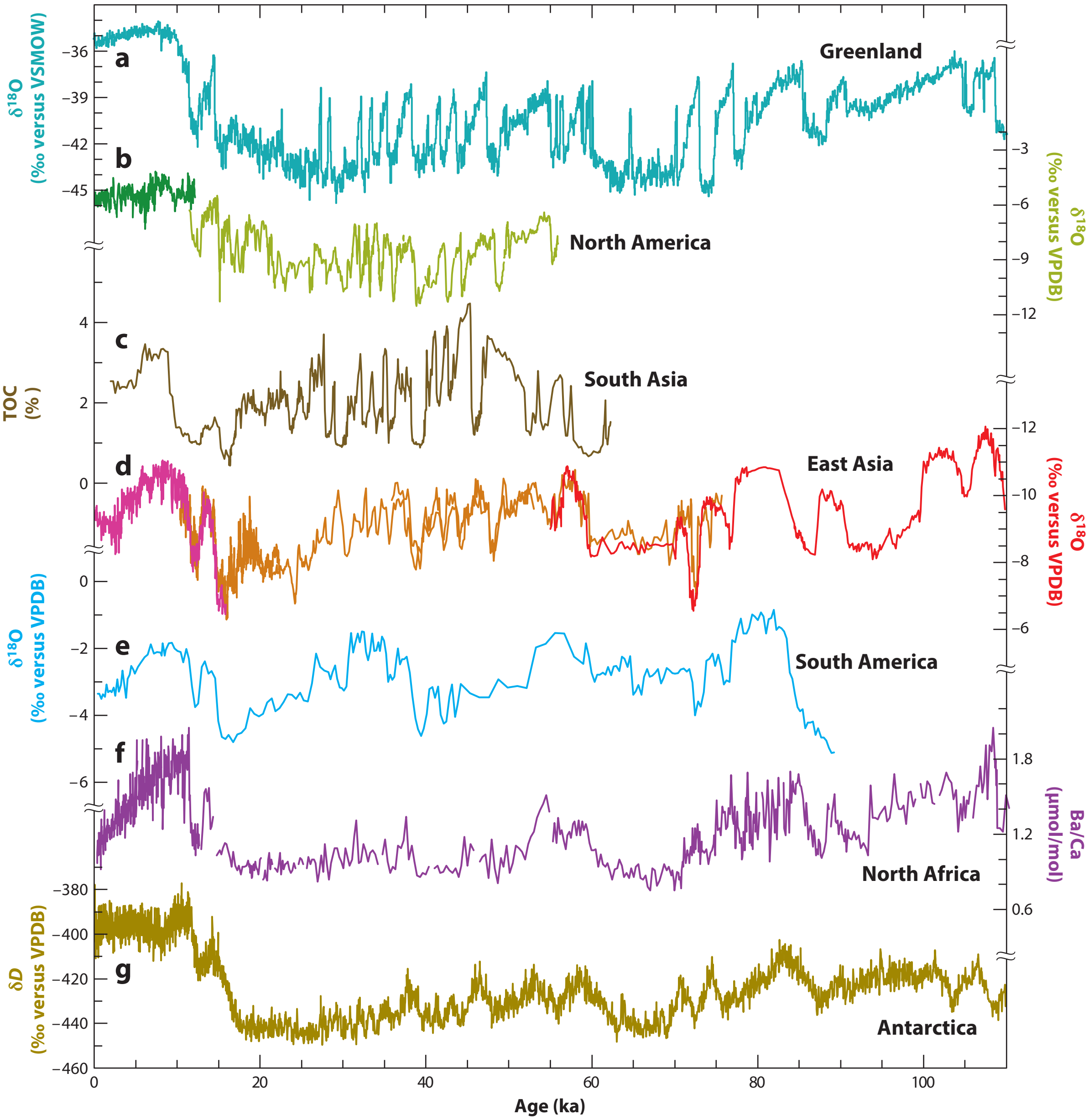


Figure 5

Comparison of different monsoon records with ice-core records since 110 ka. (*a*) Greenland: $\delta^{18}O$ from the North Greenland Ice Core Project (NGRIP) ice core (*aqua*, Andersen et al. 2004). (*b*) North America: speleothem $\delta^{18}O$ records from the southwestern United States (*dark green*, Asmerom et al. 2007; *light green*, Asmerom et al. 2010). (*c*) South Asia: total organic carbon (TOC) from the Arabian Sea (*brown*, Schulz et al. 1998). (*d*) East Asia: speleothem $\delta^{18}O$ records from East Asia (*pink*, Dykoski et al. 2005; *orange*, Wang et al. 2001b; *red*, Wang et al. 2008c). (*e*) South America: speleothem $\delta^{18}O$ records from southern Brazil (*light blue*, Wang et al. 2006). (*f*) North Africa: Ba/Ca ratios from eastern Gulf of Guinea (*purple*, Weldeab et al. 2007a). (*g*) Antarctica: δD of the European Project for Ice Coring in Antarctica (EPICA) ice core (*dark yellow*, Augustin et al. 2004). Standards: VPDB, Vienna Pee Dee Belemnite; VSMOW, Vienna Standard Mean Ocean Water.

and distant fractionation and to what extent the temperature influences the speleothem $\delta^{18}O$. New records from key regions and sophisticated isotope-enabled global circulation model simulations will be necessary to inform further deliberations on this issue.

In South American monsoon regions, speleothem records also document millennial-scale changes, which are correlated with abrupt events in the North Atlantic but exhibit an antiphase relationship with speleothem $\delta^{18}O$ records from Asian monsoon regions (**Figure 5**) (Wang et al. 2004, 2006; Cruz et al. 2005; Kanner et al. 2012; Cheng et al. 2013). This antiphase relationship indicates a close interaction between the Northern and Southern Hemispheres and suggests that a shift in the ITCZ may be responsible for the mirroring across hemispheres of abrupt events (Wang et al. 2004, 2006). In southwestern North America, millennial-scale climate changes during the late Pleistocene have been identified from speleothem $\delta^{18}O$ records (**Figure 5**) (Asmerom et al. 2007, 2010; Wagner et al. 2010). Speleothem $\delta^{18}O$ values decreased during cold stadials and increased during warm interstadials, however, due to the changing fraction of moisture from the Gulfs of Mexico and California (heavier $\delta^{18}O$ during the summer monsoon season) relative to that from the North Pacific (lighter $\delta^{18}O$ during winter) (Asmerom et al. 2010, Wagner et al. 2010), a pattern contrary to that in the Asian monsoon region.

Marine and terrestrial records have shown that in the northern Africa monsoon domain, monsoonal rainfall decreased during the Younger Dryas event, Heinrich event 1, and the intervals from 8.4 to 8 ka and from 4.2 to 4 ka, indicating the response of the northern African monsoon climate to temperature changes in the North Atlantic (deMenocal et al. 2000, Gasse 2000). Subsequently, a growing body of records has confirmed that the northern African monsoon tends to weaken at intervals corresponding to Dansgaard–Oeschger stadials and Holocene cold spells, resulting in drier and dustier conditions in the Sahel region (**Figure 5**) (Talbot et al. 2007; Weldeab et al. 2007a,b; Itambi et al. 2009; Niedermeyer et al. 2010; Zarriess & Mackensen 2010). In contrast, South Africa was wet, with significant seasonal variation during the stadials. This anticorrelation may be associated with a southward shift of the rainfall belt due to ITCZ migration during North Atlantic cold events, which in turn is linked with the bipolar seesaw response of the sea surface temperature to the AMOC (Garcin et al. 2007, Moernaut et al. 2010).

Because millennial-scale monsoon variability in various monsoon regions indicates significant coherence, and is well correlated with abrupt events in the North Atlantic region, the ocean thermohaline circulation hypothesis is widely accepted as the mechanism responsible for the teleconnection (Broecker et al. 1992, Alley et al. 1999, Clement & Peterson 2008, and references therein). Both observational and modeling studies suggest that freshwater input and/or a greater extent of sea ice in the North Atlantic may lead to significant changes in North Atlantic Deep Water and the AMOC, resulting in decreased temperatures across the northern high latitudes and a southward shift of the ITCZ (e.g., Zhang & Delworth 2005, Broccoli et al. 2006, Menviel et al. 2008, Zhang et al. 2010). In addition, sea ice changes over the North Atlantic regions may also play an important role in amplifying and transmitting the signal, and the resulting southward shift of the ITCZ, which in turn lead to weakening of the Asian summer monsoon and increased convective precipitation over Brazil and South Africa (Chiang & Bitz 2005). In Asian monsoon regions, it is also plausible that decreased temperatures across the northern high latitudes may lead to a strengthened East Asian winter monsoon and increased snow cover over the Tibetan Plateau (Barnett et al. 1989, Overpeck et al. 1996), which may in turn weaken the Asian summer monsoon in a coupled response.

Although the AMOC hypothesis can explain abrupt millennial-scale climate change and its global linkages, the recurrence of Dansgaard–Oeschger events approximately every 1,500 yr on average seems to be linked with changes in external insolation and internal interactions within the ocean-ice-climate system (Ghil et al. 1987, Maasch & Saltzman 1990, Mayewski et al. 1997). Modeling results suggest that Dansgaard–Oeschger cycles may initially be driven by periodic freshwater inputs into the North Atlantic Ocean induced by external solar activity (Braun et al. 2005, Clemens 2005, Li et al. 2005). Recently, Petersen et al. (2013) hypothesized that the abrupt warming at the onset of a Dansgaard–Oeschger cycle is caused by the rapid retreat of sea ice after the collapse of an ice shelf in the Nordic Sea, whereas gradual cooling during the subsequent interstadial phase is determined by the timescale of ice-shelf regrowth. In any case, the AMOC in the North Atlantic is most likely at the crux of these abrupt events, and land-air-sea interaction probably plays an important role in amplifying the impact of solar activity and modulating monsoon changes on millennial timescales.

Millennial-scale events may have differing characteristics within different monsoon regions. For example, Zhou et al. (2001) found that the Younger Dryas event in East Asia manifested globally in different ways, suggesting that El Niño–Southern Oscillation (ENSO)-like variability and other tropical factors may have been superimposed on this cold event. An (2000) proposed that strengthening of the southern trade winds and associated Southern Oscillation may increase the precipitation in East Asia through cross-equatorial flow. Speleothem $\delta^{18}O$ records from southwestern China demonstrate that climate change in the Southern Hemisphere has an important influence on Asian summer monsoon variations on millennial timescales (Cai et al. 2006).

During the Holocene, the Asian summer monsoon has exhibited substantial oscillations on millennial to decadal timescales as well (**Figure 6**) (Neff et al. 2001; Fleitmann et al. 2003; Gupta et al. 2003, 2005; Dykoski et al. 2005; Wang et al. 2005b; Cai et al. 2012), with weak monsoons corresponding to cold events in the North Atlantic (Bond et al. 1997). An episodic slowdown of the AMOC, and the consequent strengthening of Northern Hemisphere circulations, is considered a likely cause for a southward ITCZ shift that results in changes in low-latitude monsoonal precipitation patterns (Barber et al. 1999, Murton et al. 2010, Yu et al. 2010). On the one hand, a multiproxy, annual-layered stalagmite from central China recently showed that the duration and evolution of the precipitation during the 8.2-ka event are indistinguishable from those observed in the Greenland ice cores, suggesting an effective and rapid atmospheric teleconnection exists between the North Atlantic and East Asian summer monsoon regions (Liu et al. 2013). On the other hand, the periodicities of solar activity and monsoon records are highly correlated with $\Delta^{14}C$ records, suggesting that solar output changes are partly responsible for fine-scale monsoon variability in the Holocene (**Figure 6**) (Shindell et al. 2001; Fleitmann et al. 2003; Gupta et al. 2003, 2005; Wang et al. 2005b; Asmerom et al. 2007; An et al. 2012; Cai et al. 2012). However, other feedback processes may be involved to amplify solar output effect (Kodera 2004), as solar irradiation changes that occur on millennial to decadal timescales are very small (Vieira et al. 2011).

3.3. Centennial-Scale Monsoon Variability over the Past Millennium

The primary features of global climate during the past millennium include the most recent warm period (the Medieval Climate Anomaly, 800–1300), a cold period (the Little Ice Age, 1400–1850), and the subsequent warming (1850–present). Centennial and multidecadal variability of the global monsoon during the past millennium has been investigated extensively to ascertain characteristic patterns of natural versus anthropogenic forcings (Kumar et al. 1999, Verschuren et al. 2000, Russell & Johnson 2005, Stager et al. 2005, Newton et al. 2006, Tan et al. 2008, Liu et al. 2009, Sachs et al. 2009).

3.3.1. The Asian-Australian monsoon. Many archives, including stalagmites, lake sediments, marine sediments, and historic documents, have been studied to understand centennial-scale

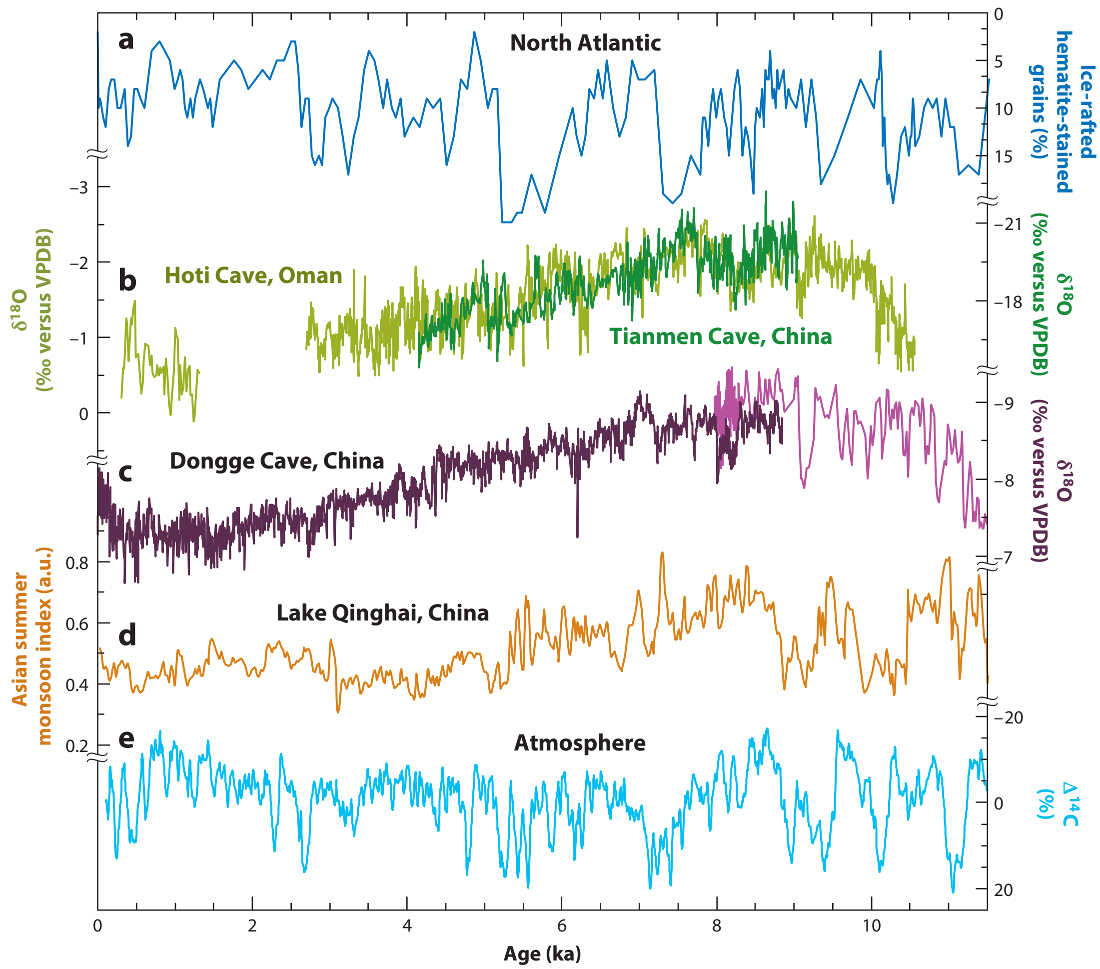


Figure 6

Comparison of Asian monsoon records with other records during the Holocene. (*a*) North Atlantic ice-rafted hematite-stained grains (*dark blue*, Bond et al. 1997). (*b*) Speleothem $\delta^{18}O$ records from Hoti Cave, Oman (*light green*, Fleitmann et al. 2003) and Tianmen Cave, southern Tibetan Plateau, China (*dark green*, Cai et al. 2012). (*c*) Speleothem $\delta^{18}O$ records from Dongge Cave, southern China (*dark purple*, Wang et al. 2005b; *light purple*, Dykoski et al. 2005). (*d*) Asian summer monsoon index from Lake Qinghai, China (*orange*, An et al. 2012). (*e*) Atmospheric $\Delta^{14}C$ (*light blue*, Reimer et al. 2009).

hydrological changes in the Asian-Australian monsoon area over the past millennium (**Figure 7**). A large body of evidence suggests that the area located at the northern limit of the Asian-Australian summer monsoon region, including northern China (Tan et al. 2008, Zhang et al. 2008, Liu et al. 2011) and India (Sinha et al. 2011), experienced a drier climate during the Little Ice Age than during the Medieval Climate Anomaly. In contrast, areas in the southern part of the Asian-Australian monsoon region, including Indonesia (Newton et al. 2006, Oppo et al. 2009, Tierney et al. 2010), entered the Little Ice Age with a wetter climate. This antiphased variation between the Asian and Australian summer monsoons during the past millennium has been attributed to the southward migration of the ITCZ (Newton et al. 2006, Sachs et al. 2009, Tierney et al. 2010). However, some hydrological records from southern China (Chu et al. 2002, Tan et al. 2009, Yan et al. 2011, Zeng

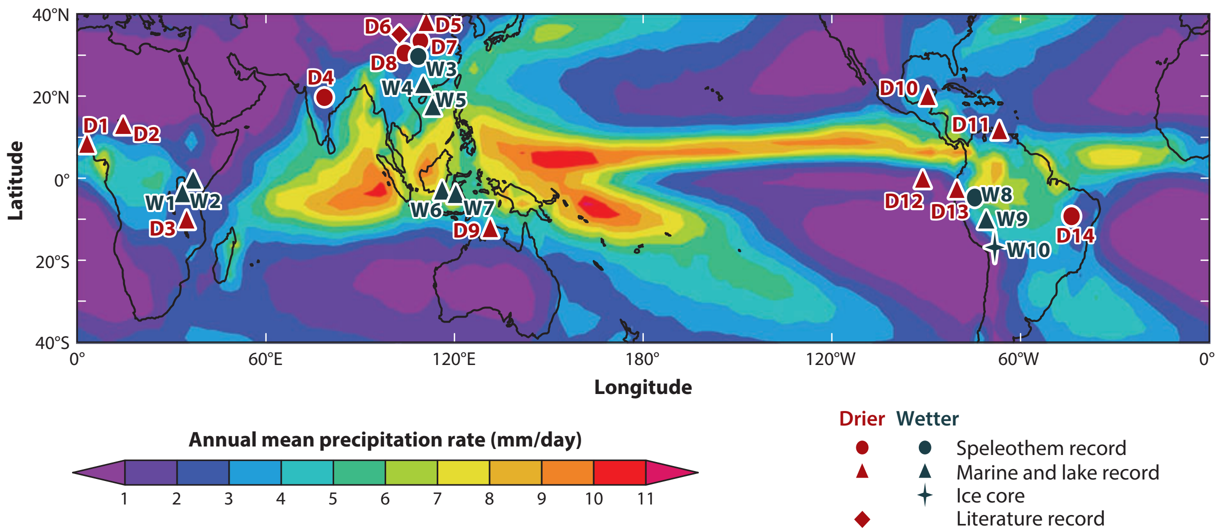


Figure 7

Centennial-scale variations of the global monsoon. The base map is the annual mean precipitation rate (mm/day) in the global tropics and subtropics derived from National Centers for Environmental Prediction (NCEP) reanalysis 2 data from January 1979 to December 2010. Locations of hydrological records in the global monsoon area covering the past millennium are also marked. Locations that were drier during the Little Ice Age (1400–1850) than during the Medieval Climate Anomaly (800–1300) are marked in dark red: D1 (Talbot & Delibrias 1977), D2 (Maley 1981), D3 (Johnson et al. 2001), D4 (Sinha et al. 2011), D5 (Liu et al. 2011), D6 (Tan et al. 2008), D7 (Zhang et al. 2008), D8 (Tan et al. 2011), D9 (Wasson & Bayliss 2010), D10 (Hodell et al. 2005), D11 (Haug et al. 2001), D12 (Conroy et al. 2008), D13 (Moy et al. 2002), and D14 (Stríkis et al. 2011, Novello et al. 2012). Locations that were wetter during the Little Ice Age than during the Medieval Climate Anomaly are marked in dark blue: W1 (Stager et al. 2005), W2 (Verschuren et al. 2000), W3 (Tan et al. 2009), W4 (Chu et al. 2002, Zeng et al. 2011), W5 (Yan et al. 2011), W6 (Newton et al. 2006, Oppo et al. 2009), W7 (Tierney et al. 2010), W8 (Reuter et al. 2009), W9 (Bird et al. 2011), and W10 (Thompson et al. 1986).

et al. 2011) (**Figure 7**) indicating a wetter Little Ice Age, along with the record from northern Australia (Wasson & Bayliss 2010) (**Figure 7**) indicating a drier Little Ice Age, are difficult to explain by the southward ITCZ migration and might be caused by other factors (e.g., Yan et al. 2011).

Recent studies have also highlighted multidecadal oscillations of the Asian summer monsoon during the last millennium (Zhang et al. 2008, Tan et al. 2011). Power spectrum analyses indicate that the speleothem $\delta^{18}O$ time series show consistent solar cycles, with periodicities at 80–120, 27–35, ~20, and ~11 yr, suggesting that solar activity may influence on the variability of the Asian summer monsoon on multidecadal timescales. Solar activity may affect the thermal contrast between Asia and the North Pacific, and thus the intensity of the East Asian summer monsoon (Zhao et al. 2007b, Tan et al. 2011). Modeling results suggest that volcanic activity may also have contributed to multidecadal global monsoon variability during the past millennium (Liu et al. 2009). Other factors, such as the ENSO and sea surface temperature in other tropical oceans, may also affect multidecadal variations of the Asian-Australian summer monsoon (Kumar et al. 1999, Oppo et al. 2009). In addition, a stalagmite $\delta^{18}O$ record from Wanxiang Cave, China, shows similarities to the North Atlantic ice-rafted debris record and the North Atlantic Oscillation record during the past 1,800 yr, suggesting a potential link between the East Asian summer monsoon and the North Atlantic Oscillation (Zhang et al. 2008).

3.3.2. African and American monsoons. Compared with those in the Asian-Australian monsoon area, hydrological variations in African monsoon regions over the past millennium are less well established, and high-resolution climatic records are concentrated in the lakes of the eastern Africa (Verschuren et al. 2000, Johnson et al. 2001, Russell & Johnson 2005, Stager et al. 2005, Wolff et al. 2011, Tierney et al. 2013). Lake level and lake sediment records do not show clear hydrological changes between the Medieval Climate Anomaly and Little Ice Age, but they do show widespread multidecadal oscillations associated with solar activity and a connection with Indian Ocean sea surface temperature (Verschuren et al. 2000, Russell & Johnson 2005, Stager et al. 2005, Tierney et al. 2013).

Numerous studies have focused on hydrological changes within American monsoon regions over the past millennium, revealing pronounced regional differences (**Figure 7**). Hydrological records from the northern limit of the ITCZ (e.g., Cariaco Basin and Yucatán Peninsula) document similar variability (Haug et al. 2001, Hodell et al. 2005). This region was drier during the Little Ice Age than during the Medieval Climate Anomaly, supporting a southward migration of the ITCZ during the Little Ice Age (**Figure 7**). The southward migration of the ITCZ is also supported by results from ice cores, stalagmites, and lake sediments from the South American Andes, which indicate wet conditions during the Little Ice Age (Thompson et al. 1986, Reuter et al. 2009, Bird et al. 2011). However, some hydrological records from both the eastern (Novello et al. 2012) and western (Moy et al. 2002, Conroy et al. 2008) coasts of South America indicate dry conditions during the Little Ice Age, which cannot be explained by an ITCZ migration (**Figure 7**). Other factors, such as the ENSO and the Atlantic Multidecadal Oscillation, may also contribute to the centennial- and multidecadal-scale hydrological changes in South America (Moy et al. 2002, Novello et al. 2012).

In summary, the centennial-scale southward migration of the ITCZ during the relatively cold Little Ice Age can explain most of the hydrological records in global monsoon areas (Sachs et al. 2009), although some monsoon records and simulated results suggest different patterns and mechanisms (Liu et al. 2009, Yan et al. 2011). Multidecadal monsoon variability is widely recognized in hydrological records of regions affected by the global monsoon, and these records are probably influenced by external changes in solar activity and internal oscillations within Earth's climate system (Kumar et al. 1999, Verschuren et al. 2000, Russell & Johnson 2005, Stager et al. 2005, Tan et al. 2011).

3.4. Interdecadal Variability

Global monsoon precipitation has intensified over the past three decades, mainly due to a significant increasing trend in the Northern Hemisphere summer monsoon. The intensification of global monsoon precipitation results primarily from an enhanced east-west thermal contrast between the Pacific and Indian Oceans that features rising pressure in the subtropical eastern Pacific and decreasing pressure over the Indo-Pacific warm pool. This mechanism tends to amplify both the Northern and Southern Hemisphere summer monsoons. However, the stronger warming trend in the Northern Hemisphere and the weaker warming trend in the Southern Hemisphere create a hemispheric thermal contrast that enhances the Northern Hemisphere summer monsoon but weakens the Southern Hemisphere summer monsoon. The strengthened Pacific zonal thermal contrast results from both natural variability and global warming, and anthropogenic forcing can intensify the hemispherical thermal contrast (Luo et al. 2012, Wang et al. 2012). However, some have suggested that the increased west-east sea surface temperature contrast is the result of observational bias (Tokinaga et al. 2012); thus, the mechanism by which the increased zonal sea surface temperature gradient in the Pacific may increase global monsoon precipitation requires further study.

3.4.1. The Asian summer monsoon. Indian summer monsoon rainfall shows coherent multidecadal variability with an approximate periodicity of 55 to 60 yr, without a trend or a climate change signal. Generally, Indian summer monsoon rainfall was above normal from 1891 to 1900 and from 1930 to 1960 and was below normal from 1901 to 1930 and from 1971 to 2000 (Goswami 2006). The interdecadal variability of Indian summer monsoon rainfall is attributed to the complex ENSO-like interdecadal variability of the Pacific sea surface temperature (Graham 1994, Kawamura 1994). A significant negative relationship between the interdecadal variability of Indian summer monsoon rainfall and Niño 3 sea surface temperature indicates a close association on interdecadal timescales (Parthasarathy et al. 1994, Torrence & Webster 1999, Krishnamurthy & Goswami 2000). An empirical inverse relationship between the Indian summer monsoon and the ENSO has weakened remarkably since the late 1970s (**Figure 8**) (e.g., Kumar et al. 1999). In addition to oceanic forcing, aerosols can also alter cloud density and affect the radiative balance in the atmosphere, leading to changes in cloud microphysics and atmospheric stability, which can either suppress or foster clouds and precipitation. Recent studies (e.g., Ganguly et al. 2012) suggested that the widespread summertime drying over South Asia during the second half of the twentieth century can be primarily attributed to human-influenced aerosol emissions—that is, the drying results from a slowdown in the tropical meridional overturning circulation, which compensates for the aerosol-induced energy imbalance between the Northern and Southern Hemispheres.

The traditional strong East Asian summer monsoon signifies the northward extension of the rainfall belt and above-normal rainfall over northern China. Since the end of the 1970s, the East Asian summer monsoon has followed an evident weakening trend with a "southern China flood and northern China drought" rainfall pattern (Zhou et al. 2009a). One possible reason is that warming in the Indian Ocean and far western Pacific Ocean has induced a westward extension of the western Pacific subtropical high by enhancing convection over the equatorial Indian Ocean and Maritime Continent (Zhou et al. 2009b). Meanwhile, an increase in the extent and depth of snow cover and a weakening in the sensible heat source over the Tibetan Plateau (Zhao et al. 2007a, Duan et al. 2013) have apparently weakened monsoon circulation and produced a southward retreat of the monsoon rainfall belt. The subtropical jet stream is also an important factor to consider in relation to the East Asian summer monsoon (Molnar et al. 2010). Relatively large-amplitude warming in the higher latitudes of the Eurasian continent could decelerate the westerly jet and produce a decadal-timescale change in the East Asian summer monsoon. All of these mechanisms are related, to a certain extent, to anthropogenic global warming. Moreover, anthropogenic aerosols might be a significant factor responsible for such decadal changes. Jiang et al. (2013) suggest that anthropogenic aerosols have suppressed precipitation in North China but enhanced precipitation in South China and adjacent oceanic regions. However, some studies (e.g., Ding et al. 2008, Lei et al. 2011) argue that decadal changes in the East Asian summer monsoon likely result from internal oscillations in the climate system. Recent decadal variability in the Asian summer monsoon may thus result from forcings related to both natural variability and human activity in the climate system.

3.4.2. The African summer monsoon. The interdecadal variability of the West African monsoon is well documented and has motivated considerable research efforts (e.g., Fontaine & Janicot 1996, Le Barbé et al. 2002). The dramatic change from wet conditions in the 1950s and 1960s to much drier conditions in the 1970s, 1980s, and 1990s over the whole region represents one of the strongest interdecadal signals on the planet in the twentieth century. The decadal-scale rainfall deficit from 1970 to 1990 is dominated by lower rainfall during the peak of the rainy season in August and early September (Le Barbé & Lebel 1997, Le Barbé et al. 2002). Furthermore, the

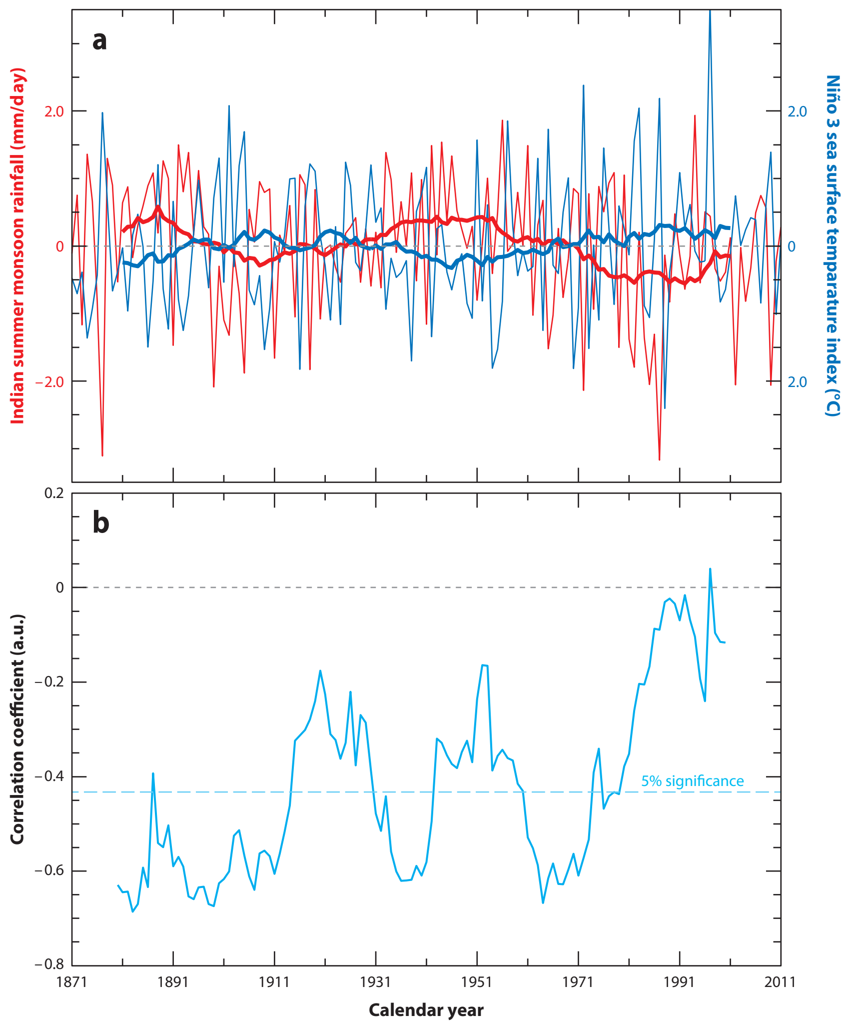


Figure 8

(*a*) The standardized time series of Indian summer monsoon rainfall (*thin red line*) and the June-July-August Niño 3 sea surface temperature index (averaged from 5°S to 5°N and 150°W to 90°W) (*thin blue line*) for the period 1871–2011. The thick lines show corresponding 21-yr running means. (*b*) The 21-yr sliding correlation between the Indian summer monsoon rainfall and the June-July-August Niño 3 sea surface temperature index. The light blue horizontal dashed line indicates the 95% confidence level (5% significance).

Sahel rainfall variability from July to September associated with the northern African summer monsoon declines distinctly from the wetter-than-average 1950s to the progressively drier 1960s, 1970s, and 1980s.

Earlier studies have attributed the long drying trend to human activity and the associated positive land-atmosphere feedback (Charney 1975). More recent modeling studies propose that such rainfall variability in the Sahel is a response of the African summer monsoon to oceanic forcing, amplified by land-atmosphere interaction (Giannini et al. 2008). The role of positive land-atmosphere interaction is also controversial. In spite of some earlier modeling evidence of positive land-atmosphere feedback through surface albedo and soil moisture, there has been no observational evidence of positive land feedback on African rainfall. Instead, a recent analysis of remote sensing observations did not find statistically significant evidence of positive vegetation feedback on African rainfall (Liu et al. 2006). Therefore, the role of land-atmosphere feedback and human effects on the African monsoon remains controversial.

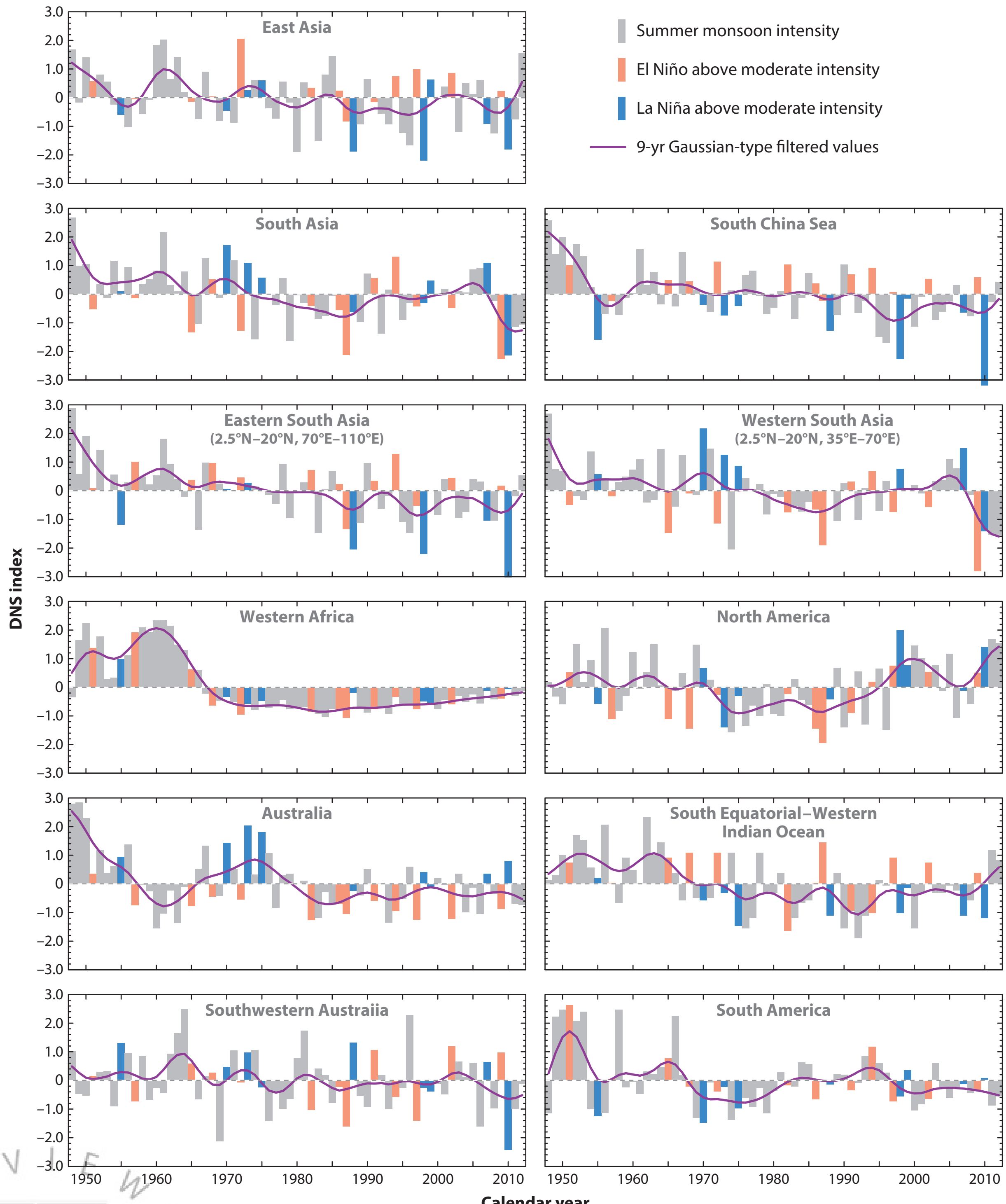
3.4.3. The American summer monsoon. The domain of the North American monsoon is large, extending from the western United States to northwestern Mexico (Adams & Comrie 1997). The North American monsoon system's strength, onset, and retreat over northwestern Mexico exhibited multidecadal variations from 1948 to 2009. Two dry regimes, associated with late onsets, early retreat, and weaker rainfall rates, occurred from 1948 to 1970 and from 1991 to 2005, whereas a wet regime, associated with early onset, late retreat, and stronger rainfall rates, occurred from 1971 to 1990. Several studies suggest that North American monsoon variability is modulated by the Pacific Decadal Oscillation (Higgins & Shi 2000, Castro et al. 2007), the Atlantic Multidecadal Oscillation (Hu & Feng 2008), and the Arctic Oscillation/Northern Hemisphere Annular Mode (Hu & Feng 2010). Arias et al. (2012) argue that the multidecadal variation is linked to the variability of sea surface temperature anomalies in conjunction with the Atlantic Multidecadal Oscillation and the warming sea surface temperature trends. However, for the South American summer monsoon, the decadal variability featured above-normal rainfall over northeastern Brazil and from the central Andes to Gran Chaco and below-normal rainfall north of the equator and over the southern Amazon Basin during most of the 1980s. The reverse trend appears to have occurred during the late 1980s and early 1990s (Zhou & Lau 2001). Decadal- and longer-timescale variability is related to ocean surface changes on those timescales in the Pacific and Atlantic Oceans (Zhou & Lau 2001).

Overall, the global monsoon exhibited decadal variation during the past century, but the temporal-spatial characteristics differed substantially among specific monsoon regions. Oceanic forcing seems to play a very important role, although the effects over specific oceans are distinct. Decadal ENSO-like variability determines long-term changes in the Asian-Australian monsoon, whereas decadal variability over the tropical Atlantic and Indian Oceans seems to determine changes in the African summer monsoon. In addition, anthropogenic aerosols may suppress precipitation in the Asian summer monsoon regions. Specifically, in the case of the East Asian summer monsoon, aerosols decrease sensible heating over the Tibetan Plateau, weakening monsoon circulation and resulting in a southward retreat of the summer rainfall belt in East China. For the American summer monsoon, decadal variability is primarily influenced by the Pacific Decadal Oscillation, Atlantic Multidecadal Oscillation (Hu & Feng 2008), and the Northern Hemisphere Annular Mode.

3.5. Interannual Variability and Intraseasonal Oscillations

Among the wide range of timescales for monsoon variability, the interannual and intraseasonal timescales are the most extensively studied due to their socioeconomic impacts on monsoon regions.

3.5.1. Interannual variability. The interannual variability of the global monsoon is a dominant year-to-year fluctuation. Rainfall over the Indian and Indonesian-Australian monsoon regions exhibits biennial variability during certain decades (Yasunari 1991, Webster et al. 1998), and similarly, a dominant oscillation at a period of 2–3 yr exists over East Asia (Chang et al. 2001, Li & Zeng 2005). **Figure 9** shows the interannual variability of monsoon intensity measured by the DNS index (Li & Zeng 2002) for some major monsoon subsystems. Besides distinct


East Asia
Summer monsoon intensity
El Niño above moderate intensity
La Niña above moderate intensity
9-yr Gaussian-type filtered values
South Asia
South China Sea
Eastern South Asia
(2.5°N–20°N, 70°E–110°E)
Western South Asia
(2.5°N–20°N, 35°E–70°E)
Western Africa
North America
Australia
South Equatorial–Western
Indian Ocean
Southwestern Austraiia
South America
DNS index
Calendar year
3.0
2.0
1.0
0
−1.0
−2.0
−3.0
1950
1960
1970
1980
1990
2000
2010

quasi-biennial fluctuations, the anomalous monsoon for each subsystem is apparently related to an ENSO event. This is because ENSO events usually induce anomalous vertical motions in the equatorial Indian Ocean via modification of the Walker circulation, and thus they affect the regional Hadley circulation associated with the Indian summer monsoon (Krishnamurthy & Goswami 2000, Kumar et al. 2006). The Indian summer monsoon and the ENSO were negatively correlated on the interannual timescale prior to the late 1970s (Kumar et al. 1999). An anomalous Indian summer monsoon may influence the interannual variability of the Australian summer monsoon: A weak Indian summer monsoon may be followed by a delayed Australian summer monsoon, and a strong Indian summer monsoon by an early Australian summer monsoon (Joseph et al. 1991). The Indian summer monsoon is also teleconnected with the West African summer monsoon on the interannual timescale; the physical link in this case is the low-level inflow into the ascending branch of the African ITCZ over the eastern Mediterranean region (Rodwell & Hoskins 1996, Raicich et al. 2003). The interannual variability of the North American summer monsoon is controlled by various anomalies in ocean and land surface conditions (e.g., sea surface temperature, soil moisture) associated with tropical climate anomalies such as the ENSO (Higgins et al. 1998). The East Asian summer monsoon, characterized by Meiyu-Baiu-Changma (the major rainy season from central China across South Korea to Japan, brought by a zonally elongated rainband from June to July), is far more complicated than other monsoons; thus, its interannual variability is forced by multiple factors including the ENSO, the Indian summer monsoon, winter snow cover in Eurasia and the Tibetan Plateau, and the Northern and Southern Hemisphere Annular Modes (Wang et al. 2001a, Nan & Li 2003, Ding & Chan 2005).

In summary, ENSO-related sea surface temperature anomalies in the Pacific and Indian Oceans are believed to be a primary factor responsible for the interannual variability of the global monsoon. Snow cover in Eurasia, thermal forcing of the Tibetan Plateau, and the Northern and Southern Hemisphere Annular Modes may also contribute to the interannual activity of the Asian summer monsoon.

3.5.2. Intraseasonal oscillations. Intraseasonal oscillations are an important component of monsoon variability; they are characterized by several active and break sequences occurring within each summer monsoon season. They significantly affect the local weather and climate as well as the global atmospheric circulation.

The predominant intraseasonal oscillation in a broad period ranging from 30 to 60 days was identified in the Asian summer monsoon regime (Mao & Chan 2005), propagating both eastward and northward (Lawrence & Webster 2002, Mao et al. 2010). Another dominant intraseasonal oscillation that modulates Asian summer monsoon activity is a 10–20-day westward-propagating oscillation (Krishnamurti & Ardanuy 1980, Mao & Chan 2005). The active and break sequences of the Indian summer monsoon are associated with shifts of the monsoon trough position (Webster et al. 1998). The active-break cycles in rainfall over the Indian summer monsoon zone are linked with fluctuations in the intensity of the continental tropical convergence zone (TCZ) (Sikka & Gadgil 1980, Gadgil 2003); the continental TCZ results from northward migration of the Indian Ocean TCZ during late boreal spring to boreal summer. This northward propagation is driven by

Figure 9

Time series of area-averaged summer monsoon intensity (*bars*) indicated by the dynamical normalized seasonality (DNS) index (Li & Zeng 2002) over different submonsoon regions. The corresponding El Niño and La Niña years with El Niño–Southern Oscillation events above moderate intensity are colored orange and blue, respectively. The solid purple lines indicate the 9-yr Gaussian-type filtered values.

the north-south gradient of total heating that results from the meridional gradient of convective stability and moisture availability; the maximum convective heating is displaced toward the north of the maximum vertical ascent (Gadgil & Srinivasan 1990). The active and break sequences in Meiyu rainfall of the East Asian summer monsoon are related to convective anomalies over the South China–Philippine Sea, which result from a Rossby wave–like coupled circulation-convection system that propagates north-northwestward from the equatorial western Pacific (Mao et al. 2010).

The timing of summer monsoon onset is one of the most significant manifestations of intraseasonal variability (Webster et al. 1998). In the Australian-Asian monsoon regime, the yearly summer monsoon onset is often associated with an eastward-propagating tropical intraseasonal oscillation event (Madden & Julian 1994). In northern Australia, the monsoon onset date is defined to be the beginning of the first westerly wind burst in each wet season (Wheeler & McBride 2012). The Asian summer monsoon onset arises primarily from the arrival of a wet phase in the tropical intraseasonal oscillation or a phase lock between intraseasonal oscillation modes (Ding & Chan 2005). The onset of the Asian summer monsoon occurs in the eastern Bay of Bengal in early May, in the South China Sea in mid-May, and in South Asia in early June (Wu & Zhang 1998, Mao & Wu 2007). It may be concurrent with the reversal of the meridional temperature gradient (MTG) in the upper troposphere south of the Tibetan Plateau (Flohn 1957, Li & Yanai 1996). Thus, the area-averaged MTG in the middle-upper troposphere has been proposed as an index to measure the onset of the Asian summer monsoon for different regions (Mao & Wu 2007). Rajagopalan & Molnar (2012) suggest that the MTG-based Indian summer monsoon onset and withdrawal dates (Goswami & Xavier 2005) depend on the ENSO much more clearly than those derived from other variables, such as rainfall and large-scale circulation (Joseph et al. 2006).

Although the anomalous convection during each active phase of the Asian and Australian summer monsoons is initiated from the equatorial Indian and western Pacific Oceans, where air-sea interactions may help generate the convective disturbances, the intraseasonal oscillations of summer monsoons are considered to represent an atmospheric internal variability with large-scale coupling between atmospheric circulation and deep convection. The poleward migration of such anomalous convection into the continental monsoon zone supports an intrinsic dynamic mechanism—such as cloud-radiation feedbacks on the position of the TCZ (Gadgil 2003) or instability of the coupled ocean-atmosphere system (Webster et al. 1998)—for the intraseasonal variability of summer monsoons. Extratropical disturbances such as the passage of midlatitude cold surges can also help trigger the tropical convection. The intraseasonal oscillations of the Asian-Australian monsoon may be related to tropical-extratropical interactions (Hsu 2012, Wheeler & McBride 2012).

4. THE TIBETAN PLATEAU AND THE CENOZOIC ASIAN MONSOON

4.1. Cenozoic Asian Monsoon Formation

The formation and development of the Asian summer monsoon during the Cenozoic were closely linked to land-sea redistribution (including the retreat of the Paratethys Sea) and Tibetan Plateau growth; they were also affected by factors such as changes in global ice volume, sea level, and atmospheric CO_2 (Prell & Kutzbach 1992, deMenocal & Rind 1993, Kutzbach et al. 1993, Ramstein et al. 1997, Liu & Yin 2002). The earliest Cenozoic Asian summer monsoon was previously thought to have formed during the very late Oligocene to early Miocene (Qiang et al. 2001, Sun & Wang 2005, Guo et al. 2008). During the Eocene, however, coal and oil shale were widely deposited along the southern margins of the subtropical arid zone (Gu & Renaut 1994), evergreen trees began to appear in South China (Guo 1965, 1983), and mammals preferring moist and forested habitats became common in southeastern China (Qiu & Li 2005). These observations imply that the Asian paleomonsoon may have initiated in the southern tropical zone as early as the Eocene, associated with a commensurate northward retreat of the arid zone and westerly planetary wind system (Guo 1965, Qiu & Li 2005, Huber & Goldner 2012).

Numerical experiments demonstrate that when there is only extratropical continent located between 0° and 120°E and between 20°N–30°N and the North Pole, a weak monsoon rainband appears along the southern border of the continent (Liang et al. 2005, Privé & Plumb 2007, Wu et al. 2012a). Paleogeographic reconstructions suggest that the Indo-Eurasia collision produced a large extratropical continent in the Northern Hemisphere during the Eocene (Molnar & Stock 2009). Thus, according to this line of thought, the Cenozoic Asian summer monsoon should have initiated in the Eocene. Furthermore, numerical experiments show that a strong Asian summer monsoon circulation can be induced by solar forcing when the Tibetan Plateau elevation is half that of today (Prell & Kutzbach 1992). The appearance of high-altitude vegetation after 38 Ma in the Xining Basin (Dupont-Nivet et al. 2008), rapidly increased sedimentation rates at ~40 Ma in the Hoh Xil Basin (Wang et al. 2008b), isotope-based paleoaltimetry in the Lunpola Basin (Rowley & Currie 2006), and increased exhumation and erosion at ~35 Ma in the Kunlun Mountains (Clark et al. 2010) indicate that the Tibetan Plateau had already reached a high elevation during the late Eocene. In addition, the Paratethys Sea retreated westward from the Tarim Basin since the Eocene (Bosboom et al. 2011), and this may have also contributed to the initiation of the Asian summer monsoon (Ramstein et al. 1997).

4.2. Tibetan Plateau Growth and Asian Monsoon Evolution on Tectonic Timescales

The Tibetan Plateau began to grow after India collided with Eurasia at 55–45 Ma (Molnar et al. 2010). The expansion of the Tibetan Plateau followed a multiphase process of northward and eastward growth and uplift from its central and southern parts. At 35–20 Ma, the central Tibetan Plateau was raised to a height of 3,000–4,500 m (Rowley & Currie 2006, DeCelles et al. 2007). At 15–8 Ma, the Tibetan Plateau extended eastward and northeastward, and its height and range approached present-day conditions (Yuan et al. 2013). Since the Pliocene, limited growth may have occurred in its northern, northeastern, and eastern margins (e.g., Zheng et al. 2000, An et al. 2001). Tibetan Plateau growth and uplift during the Cenozoic had profound net effects on the Asian summer monsoon and global climate change.

The thermal and mechanical effects of the Tibetan Plateau on general atmospheric circulations were first recognized from meteorological observations (Bolin 1950, Yeh 1952, Flohn 1957, Yeh et al. 1957) and from the "with/without mountains" experiments using earlier numerical models (Kasahara et al. 1973, Manabe & Terpstra 1974, Hahn & Manabe 1975). With more complicated models, sensitivity experiments demonstrate that the Tibetan Plateau intensifies both the winter and summer monsoons in Asia (e.g., Kutzbach et al. 1989, 1993; Ruddiman et al. 1989; Prell & Kutzbach 1992; An et al. 2001). Because the Tibetan Plateau blocks northward moisture delivery from the tropical Indian Ocean and forces strong subsidence air flows, inland Asian aridity (Manabe & Broccoli 1990, Broccoli & Manabe 1992) and the associated atmospheric dust cycle (Shi et al. 2011) are significantly intensified. An experiment with phased uplift of the Tibetan Plateau by an ideal 10% increase in all regions indicated that the response of the East Asian summer monsoon is more sensitive than that of the Indian monsoon (Liu & Yin 2002). This response can be intensified by oceanic feedbacks (Kitoh 2004). Furthermore, Tibetan Plateau uplift can amplify the response of the East Asian summer monsoon to orbital forcing (Liu et al. 2003a).

Kutzbach and colleagues were among the first to simulate the impacts of plateau growth on Asian summer monsoon evolution with four different "realistic" scenarios of temporal and spatial expansions based on limited geological evidence (An et al. 2001). As the Tibetan Plateau grew, the East Asian summer and winter monsoons, inland aridity, and the westerly jet were strengthened. This supports a coupling between Asian climate evolution and phased Tibetan Plateau growth. The distinct responses of the East Asian and Indian summer monsoons to the uplift of different parts of the Tibetan Plateau have also been verified by recent numerical experiments (Zhang & Liu 2010, Tang et al. 2013). Furthermore, a recent study (Shi et al. 2014) demonstrated that Mongolian Plateau uplift also contributes substantially to the strengthened westerly jet.

Using geological evidence and simulation results, An et al. (2001) showed a close coupling between the evolution of the Asian summer monsoon and the phased uplift of the Tibetan Plateau since the Late Miocene. Subsequently, An et al. (2006) further suggested a four-stage increase in summer monsoon intensity and Asian interior aridification at 25–22, 16–14, 10–7, and 4–2.6 Ma, based on a broad range of paleoclimatic records. The global ice volume remained relatively stable, or even increased in certain stages, and global CO_2 concentration remained at a relatively consistent low level after ~25 Ma (Zachos et al. 2001, Tipple & Pagani 2007). The four stages of monsoon intensification can hardly be explained by changes in ice volume and CO_2 concentration. Therefore, they could be attributed to the stepwise Tibetan Plateau growth revealed by sedimentary, geochemical, and tectonic evidence (An et al. 2006). They are also supported by synchronous enhanced chemical weathering of South China Sea (Wei et al. 2006). In particular, eolian red clay consisting of alternating loess and paleosol layers on the west Loess Plateau was deposited after ~25 Ma (Qiang et al. 2011). Molnar (2005) suggested an important Tibetan Plateau uplift at ~8 Ma based on ecological, tectonic, and sedimentary records. This is consistent with the eastward expansion of red clay from the west Loess Plateau after 8 Ma (Sun et al. 1998, Ding et al. 1999, Qiang et al. 2001), which indicates further intensification of the Asian summer monsoon and aridification. In addition, the onset of major Northern Hemisphere glaciation after 2.6 Ma appears to have influenced the development of the Asian summer monsoon. The magnetic susceptibility of Chinese loess indicates that after 2.6 Ma, the Asian summer monsoon began to fluctuate in a pattern closely following that of glacial cycles.

4.3. The Tibetan Plateau and the Asian Monsoon

Monsoons are not controlled by land-sea thermal contrast alone; zonal asymmetric diabatic heating and large-scale orography can also significantly affect them (e.g., Hahn & Manabe 1975, Molnar et al. 1993, Chakraborty et al. 2002, Liu et al. 2007). Mechanical forcing by the Tibetan Plateau is dominant in winter, whereas heating is more important than topography in forcing the summer stationary waves in the tropical and subtropical monsoons (Wu et al. 2005, 2007). Recently, Wu et al. (2012a,b) reinvestigated the influences of the land-sea distribution and large-scale orography on the formation of the modern Asian summer monsoon system. They emphasized the significance of tropical land on the genesis of cross-equatorial flow. In aquaplanet experiments, no monsoon is observed (Liang et al. 2005). When only extratropical land is present in the model, a weak summer monsoon is produced in association with an intense ITCZ. When tropical land is also prescribed, the ITCZ over the longitude domain of the extratropical land disappears in response to strong surface cross-equatorial flow from the winter hemisphere to the summer hemisphere, and a prototype of the Asian summer monsoon—especially the tropical Asian summer monsoon—is produced. However, this monsoon remains in southern India and southern China and cannot extend to the north (Wu et al. 2012a).

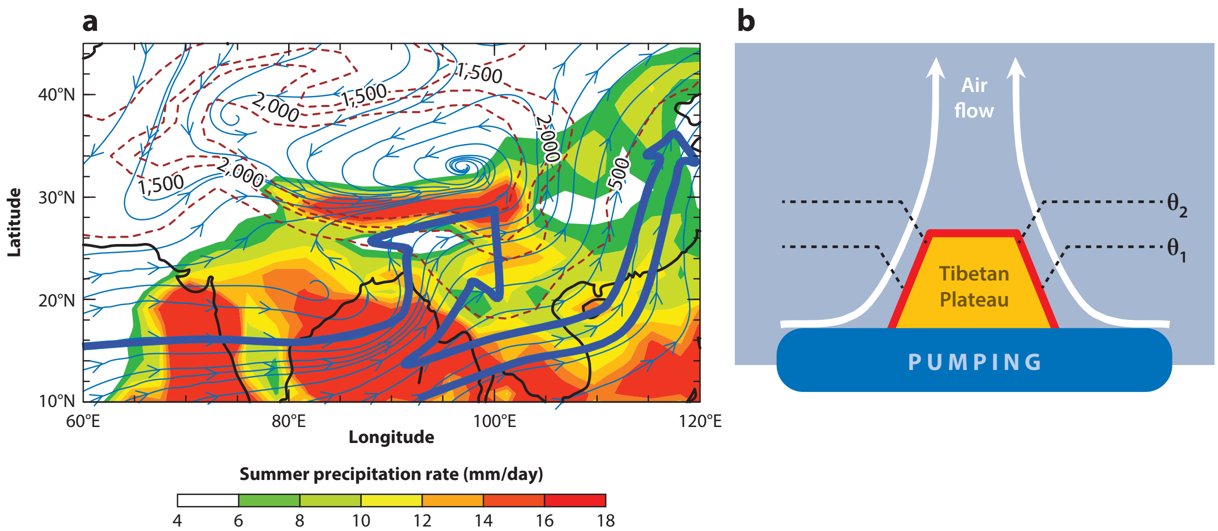


Figure 10

(*a*) The summer precipitation rate (*colored shading*) and streamlines at the $\sigma = 0.89$ model level for the full Tibetan Plateau experiment in which surface heating exists. Dashed contours denote elevations at 500, 1,500, 2,000, and 2,500 m. Dark blue open arrows denote the main atmospheric flows that impinge on the Tibetan Plateau, either climbing up or moving around it. (*b*) The corresponding mechanism for the Tibetan Plateau surface sensible heat–driven air pump (TP-SHAP). The yellow trapezoid and thick red line represent the plateau and its surface sensible heating, respectively. Dashed black lines denote the stratification of the isentropic surfaces θ_1 and θ_2. The white vector indicates the ascending air flow penetrating the isentropic surfaces from smaller θ_1 to larger θ_2 due to the pumping of the TP-SHAP (Wu et al. 2007, 2012b). The heated air particles at the sloping surface penetrate the isentropic surfaces θ_1 and θ_2 and slide upward, creating a strong rising motion and even heavy rainfall over the Tibetan Plateau.

The surface sensible heating on large mountain slopes in summer can pump up the surrounding atmosphere to produce surface convergent flow, and cooling in winter can pump it down to produce divergent flow; together, these actions combine to form a sensible heat–driven pump and thereby affect monsoon circulation (Wu et al. 1997, 2007). The uplifted thermal forcing in summer strengthens the coupling between the subtropical and tropical circulation, and between the lower and upper troposphere. The integration of the Iran Plateau into the Tibetan Plateau generated an extra cyclonic circulation in the lower troposphere, which contributes to the dryness in North Africa and the heavy precipitation over the Arabian Sea and northern India, enhances the Indian and East Asian summer monsoons, and spurs the development of a mid-Asia desert (**Figure 10**) (Wu et al. 2012a).

Boos & Kuang (2010, 2013) recently proposed that Himalayan mechanical forcing exerts a block effect on the colder and drier air advection from the north and suggested that the high surface entropy over North India plays a more important role than the Tibetan Plateau does in producing the local monsoon rainfall and upper warm center over North India. Wu et al. (2007, 2012b), in contrast, emphasize that the sensible heat–driven pump effect of large mountains drives low-level moisture advection from the Indian Ocean to the southern Tibetan Plateau to support the high surface entropy over North India and maintain the north branch of the Asian summer monsoon. Because high surface entropy requires high surface potential temperature as well as high specific humidity, both local surface heating and water vapor transport from ocean to land are required. In any case, both results demonstrate that the Asian summer monsoon is thermally controlled.

The intensity of Tibetan Plateau heating contributes to monsoon variability. Duan et al. (2006) and Duan & Wu (2008) found that the amount of temporal change in boreal spring surface air temperature and in boreal spring and summer land temperature over the Tibetan Plateau has increased significantly in recent decades. Surface sensible heating over the Tibetan Plateau decreased steadily from the mid-1970s to the end of the twentieth century, because surface wind speed decreased during that time and the change in wind speed is much greater than the change in temperature difference between the land surface and the air (Liu et al. 2012). The weakening trend in forcing by the Tibetan Plateau during summer weakens the surface cyclonic circulation and contributes significantly—at least in terms of the underlying mechanism—to a decreasing JJA precipitation trend over North China but an increasing trend over northwestern China and South China (Wu et al. 2012a,b). Numerical experiments based on an atmospheric general circulation model and an ocean-atmosphere coupled model have verified that the variation in thermal forcing by the Tibetan Plateau contributes to the rainfall pattern with wetter climate in the south and drier climate in the north on decadal timescales (Liu et al. 2012).

5. SUMMARY AND PERSPECTIVES

5.1. Common Features of Monsoon Climate

The monsoon definition we put forth in Section 1 emphasizes the common characteristics and dynamic coherence within the global monsoon. Changes in the climates of the various monsoon regimes all have significant seasonality and similar changes in periods and trends on orbital timescales (**Table 1**). For example, the precession cycle is distinctive in all the monsoonal proxies, implying that the global monsoon shares the coherent variability induced by solar radiation. However, owing to regional diversity, such as differences in underlying surface properties, land-sea distribution, and cloud-radiation forcing, regional monsoons can respond differently even under the same forcing. For example, the vast Eurasian continent, the Tibetan Plateau, and the Indo-Pacific warm pool have made the Asian monsoon the strongest and largest monsoon on the planet.

The periodicity of monsoon changes on different timescales should be attributed to responses to external and/or internal forcings; irregular monsoon changes also develop in response to nonlinear processes and stochastic perturbations. Instability or abrupt changes in the global monsoon are identified as a rapid jump from one stable state to another, reflecting a nonlinear response to the continuing variation of external forcing parameters. For example, abrupt monsoon weakening events recorded in the last glacial stalagmite and loess deposits occurred in response to the North Atlantic cooling events. The onset date and rainfall belt migrations of the Asian summer monsoon (Ding & Chan 2005, Li & Zhang 2009) and the southward retreat of maximum precipitation belts over East Asia in the Holocene (An et al. 2000) exemplify asynchronous monsoon properties. The irregularity, abruptness, and asynchroneity of monsoons increase the complexity and limit the predictability of monsoon evolution.

5.2. Multiscale Monsoon Dynamics

Monsoon dynamics can be considered as the combined effects of external (orbital parameters and solar activities) and internal (surface boundary conditions and land-air-sea interactions) forces at different timescales. Monsoon variability encompasses a wide range of timescales, expressed as intraseasonal to interannual cycles and interdecadal, multidecadal, centennial, suborbital (millennial), orbital, and tectonic variability (**Figure 11**). However, the primary driving forces

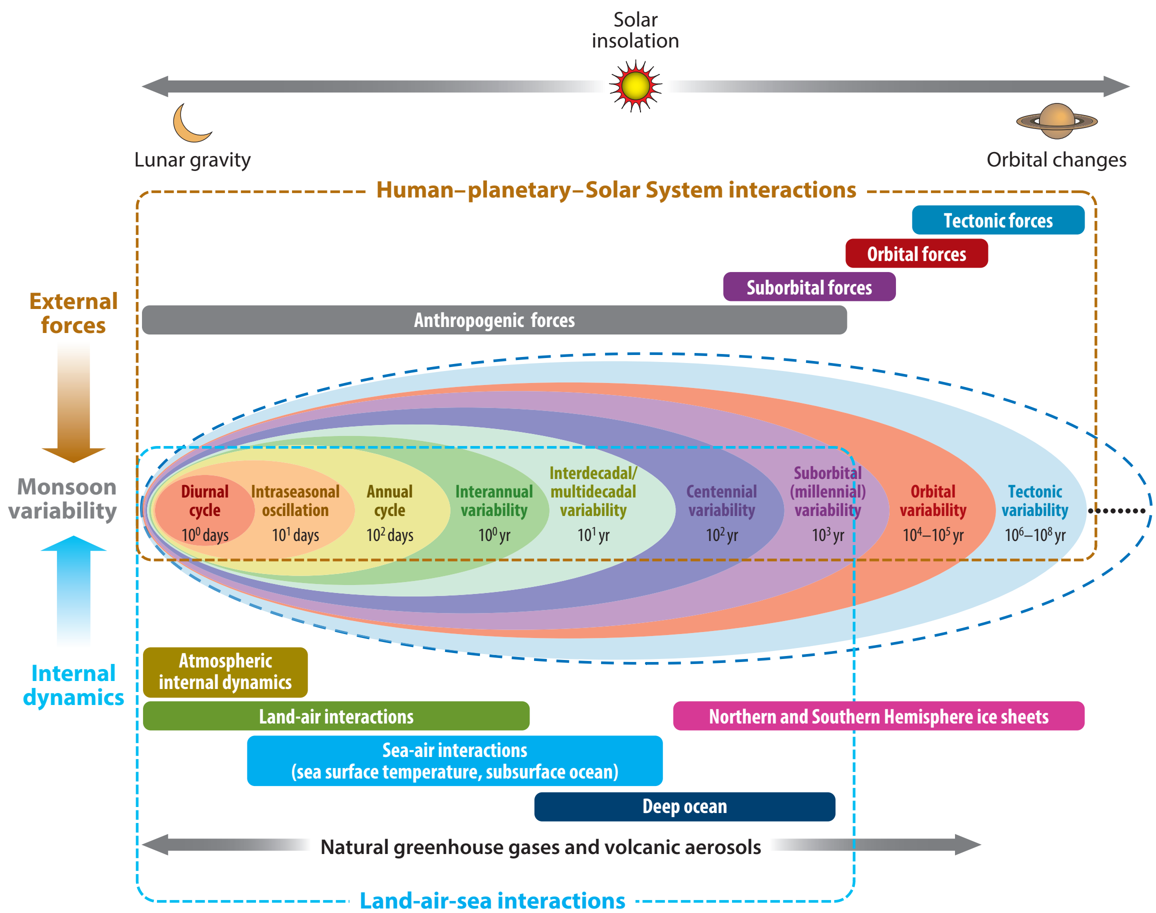


Figure 11

Monsoon variability and multiscale monsoon dynamics. The numbers under each label represents the corresponding timescale. The Sun, Moon, and planet icons respectively denote solar insolation, lunar gravity, and orbital changes associated with the gravitational pull of Solar System stars on Earth.

of monsoon variability on different timescales are not unique. Moreover, multiscale monsoon dynamics involves inherent interactions among individual timescales.

Tectonic-scale monsoon evolution and variability are forced predominantly by mountain uplift/growth and changes in sea-land configuration (e.g., Tibetan Plateau uplift and Paratethys Sea retreat), as well as Northern and Southern Hemisphere ice-sheet changes. Orbital-timescale monsoon variations are primarily controlled by seasonal variations in insolation imparted through changing orbital parameters (e.g., obliquity, precession, and eccentricity) and are strongly modulated by changes in ice volume, greenhouse gas concentrations, and interhemispheric contrast. Millennial-scale monsoon variability is related to North Atlantic climate changes induced by external solar activity and internal interactions. Internal land-air-sea interactions (e.g., thermohaline circulation, sea-ice extent, sea surface temperature), which might be related to solar output and anthropogenic forcing (e.g., greenhouse gases, aerosols, and land use), apparently impact centennial to intraseasonal monsoon variability.

Multiscale interactions increase the complexity of monsoon variability. Longer-timescale monsoon variability usually provides a background for change on shorter timescales. For instance, Tibetan Plateau uplift and Paratethys Sea retreat since the Eocene set the sea-land configuration in Asia, which provided a background for the ensuing monsoon variations. Orbital-timescale monsoon variability may be amplified by Tibetan Plateau growth (Prell & Kutzbach 1992, Liu et al. 2003a). After the onset of the Northern Hemisphere glaciations, the amplitude and frequency of monsoon variations were modulated by changes in ice volume and atmospheric greenhouse gases. The amplitude and frequency of millennial climatic events apparently differ between glacials and interglacials, indicating that ice volume might modulate millennial-scale events (Sima et al. 2004, Wang & Mysak 2006, Wang et al. 2008b). Asian summer monsoon weakening events at centennial to multidecadal timescales were smaller in amplitude than millennial monsoon weakening events within Holocene (Wang et al. 2005b), implying a constraint that millennial forcing places on centennial to multidecadal variability.

5.3. Perspectives on Future Monsoon Research

Monsoonal climate changes are driven by external and internal forces that primarily manifest as changes in circulation and precipitation. These external and internal factors can influence monsoon variability by affecting the thermal difference in the underlying surface and the resultant pressure gradient. Owing to the intrinsic mechanisms, the responses of the global monsoon to these forcings show some coherent features in spatial-temporal variations. However, deciphering the dynamics of the global monsoon remains challenging due to unresolved issues, such as the roles of low-latitude processes and the interactions between high- and low-latitude climates and between Southern and Northern Hemisphere climates across various timescales. Particularly, further investigation is needed into the relationship of monsoon variability on tectonic timescales with the growth of mountains and the opening and closure of ocean gateways; with ice sheet expansions in the Northern and Southern Hemispheres; and with atmospheric CO_2 changes. In the context of orbital timescales, assessing the physical meaning of different monsoonal indices and identifying the common proxies for global monsoon changes will help to resolve discrepancies among multiple monsoon proxies and elucidate the underlying monsoon dynamics. North Atlantic climate change is believed to contribute to global monsoon changes on millennial to centennial timescales. However, the ultimate cause for abrupt climate change in the North Atlantic, and whether the impacts of such change on monsoon dynamics are conveyed through atmospheric and/or oceanic circulations, is still unclear. Monsoonal climate variability on decadal and shorter timescales is mainly associated with land-air-sea interactions and anthropogenic forcings; however, the relative contributions are not well understood.

Challenges also remain in uncovering the connections among the forces that drive monsoons across different spatial-temporal scales, as well as connections among monsoonal and other atmospheric circulations. The inhomogeneous and asynchronous properties of monsoonal circulation and related precipitation changes also remain to be elucidated. It will be essential to carry out integrated studies that combine past and present monsoon changes and compare data from observations, proxies, and models. Research on monsoon changes since the Industrial Revolution, especially in the past several decades, should also be emphasized. Priorities for future research should include numerical simulation, impact assessment, and prediction and projection of monsoon variability on different temporal-spatial scales—particularly on seasonal to centennial timescales. Changes in global and regional monsoon trends and environmental effects must be evaluated from the perspective of both natural and anthropogenic contributions, especially under global warming conditions. The role of anthropogenic forcings such as greenhouse gases, aerosols, vegetation, and land use need to be understood in the context of the larger land-air-sea coupling system.

DISCLOSURE STATEMENT

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LITERATURE CITED

Adams DK, Comrie AC. 1997. The North American monsoon. *Bull. Am. Meteorol. Soc.* 78:2197–213

Alley R, Clark P, Keigwin L, Webb R. 1999. Making sense of millennial-scale climate change. *Geophys. Monogr.* 112:385–94

Altabet MA, Higginson MJ, Murray, DW. 2002. The effect of millennial-scale changes in Arabian Sea denitrification on atmospheric CO_2. *Nature* 415:159–62

An ZS. 2000. The history and variability of the East Asian paleomonsoon climate. *Quat. Sci. Rev.* 19:171–87

An ZS, Clemens SC, Shen J, Qiang X, Jin Z, et al. 2011. Glacial-interglacial Indian summer monsoon dynamics. *Science* 333:719–23

An ZS, Colman SM, Zhou W, Li X, Brown ET, et al. 2012. Interplay between the Westerlies and Asian monsoon recorded in Lake Qinghai sediments since 32 ka. *Sci. Rep.* 2:619

An ZS, Kukla G, Porter SC, Xiao JL. 1991a. Magnetic susceptibility evidence of monsoon variation on the loess plateau of central China during the last 130000 years. *Quat. Res.* 36:29–36

An ZS, Kutzbach JE, Prell WL, Porter SC. 2001. Evolution of Asian monsoons and phased uplift of the Himalaya-Tibet Plateau since Late Miocene times. *Nature* 411:62–66

An ZS, Liu TS, Lu YC, Porter SC, Kukla G, et al. 1990. The long-term paleomonsoon variation recorded by the loess-paleosol sequence in central China. *Quat. Int.* 7–8:91–95

An ZS, Porter SC. 1997. Millennial-scale climatic oscillations during the last interglaciation in central China. *Geology* 25:603–6

An ZS, Porter SC, Kutzbach JE, Wu X, Wang S, et al. 2000. Asynchronous Holocene optimum of the East Asian monsoon. *Quat. Sci. Rev.* 19:743–62

An ZS, Sun YB, Chang H, Zhang PZ, Liu XD, et al. 2014. Late Cenozoic climate in monsoon-arid Asia and global changes. In *Late Cenozoic Climate Change in Asia: Loess, Monsoon and Monsoon-Arid Environment Evolution*, ed. ZS An, pp. 491–582. Dordrecht, Neth.: Springer

An ZS, Wu XH, Wang PX, Wang SM, Dong GR, et al. 1991b. Paleomonsoons of China over the last 130,000 years. *Sci. China B* 34:1007–24

An ZS, Zhang PZ, Wang E, Wang SM, Qiang XK, et al. 2006. Changes of the monsoon-arid environment in China and growth of the Tibetan Plateau since the Miocene. *Quat. Sci.* 26:678–93 (in Chinese, with English abstract)

Andersen KK, Azuma N, Barnola JM, Bigler M, Biscaye P, et al. (NGRIP members). 2004. High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* 431:147–151

Ao H, Dekkers MJ, Qin L, Xiao GQ. 2011. An updated astronomical timescale for the Plio-Pleistocene deposits from South China Sea and new insights into Asian monsoon evolution. *Quat. Sci. Rev.* 30:1560–75

Arias PA, Fu R, Mo KC. 2012. Decadal variation of rainfall seasonality in the North American monsoon region and its potential causes. *J. Clim.* 25:4258–74
Asmerom Y, Polyak V, Burns S, Rassmussen J. 2007. Solar forcing of Holocene climate: new insights from a speleothem record, southwestern United States. *Geology* 35:1–4
Asmerom Y, Polyak VJ, Burns SJ. 2010. Variable winter moisture in the southwestern United States linked to rapid glacial climate shifts. *Nat. Geosci.* 3:114–17
Augustin L, Barbante C, Barnes PRF, Barnola JM, Bigler M, et al. (EPICA community memb.). 2004. Eight glacial cycles from an Antarctic ice core. *Nature* 42:623–28
Baker PA, Rigsby CA, Seltzer GO, Fritz SC, Lowenstein TK, et al. 2001. Tropical climate changes at millennial and orbital timescales on the Bolivian Altiplano. *Nature* 409:698–701
Barber DC, Dyke A, Hillaire-Marcel C, Jennings AE, Andrews JT, et al. 1999. Forcing of the cold event of 8,200 years ago by catastrophic drainage of Laurentide Lakes. *Nature* 400:344–48
Barnett TP, Dümenil L, Schlese U, Roeckner E, Latif M. 1989. The effect of Eurasian snow cover on regional and global climate variations. *J. Atmos. Sci.* 46:661–86
Beaufort L, Van der Kaars S, Bassinot F, Moron V. 2010. Past dynamics of the Australian monsoon: precession, phase and links to the global monsoon concept. *Clim. Past* 6:695–706
Berger A. 1978. Long term variations of daily insolations and Quaternary climatic changes. *J. Atmos. Sci.* 35:2362–67
Bird BW, Abbott MB, Vuille M, Rodbell DT, Stansell ND, et al. 2011. A 2,300-year-long annually resolved record of the South American summer monsoon from the Peruvian Andes. *Proc. Natl. Acad. Sci. USA* 108:8583–88
Bolin B. 1950. On the influence of the Earth's orography on the general character of the Westerlies. *Tellus* 2:184–95
Bond G, Showers W, Cheseby M, Lotti R, Almasi P, et al. 1997. A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates. *Science* 278:1257–66
Boos WR, Kuang Z. 2010. Dominant control of the South Asian monsoon by orographic insulation versus plateau heating. *Nature* 463:218–22
Boos WR, Kuang Z. 2013. Sensitivity of the South Asian monsoon to elevated and non-elevated heating. *Sci. Rep.* 3:1192
Bosboom RE, Dupont-Nivet G, Houben AJP, Brinkhuis H, Villa G, et al. 2011. Late Eocene sea retreat from the Tarim Basin (west China) and concomitant Asian paleoenvironmental change. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 299:385–98
Braun H, Christl M, Rahmstorf S, Ganopolski A, Mangini A, et al. 2005. Possible solar origin of the 1,470-year glacial climate cycle demonstrated in a coupled model. *Nature* 438:208–11
Broccoli AJ, Dahl KA, Stouffer RJ. 2006. Response of the ITCZ to Northern Hemisphere cooling. *Geophys. Res. Lett.* 33:1–4
Broccoli AJ, Manabe S. 1992. The effects of orography on midlatitude Northern Hemisphere dry climates. *J. Clim.* 5:1181–201
Broecker WS, Bond G, Klas M, Clark E, McManus J. 1992. Origin of the northern Atlantic's Heinrich events. *Clim. Dyn.* 6:265–73
Broecker WS, Peteet DM, Rind D. 1985. Does the ocean-atmosphere system have more than one stable mode of operation? *Nature* 315:21–26
Burns SJ, Fleitmann D, Matter A, Kramers J, Al-Subbary AA. 2003. Indian Ocean climate and an absolute chronology over Dansgaard/Oeschger events 9 to 13. *Science* 301:1365–67
Cai Y, An Z, Cheng H, Edwards RL, Kelly MJ, et al. 2006. High-resolution absolute-dated Indian monsoon record between 53 and 36 ka from Xiaobailong Cave, southwestern China. *Geology* 34:621–24
Cai Y, Zhang H, Cheng H, An ZS, Lawrence Edwards R, et al. 2012. The Holocene Indian monsoon variability over the southern Tibetan Plateau and its teleconnections. *Earth Planet. Sci. Lett.* 335–36:135–44
Caley T, Malaizé B, Kageyama M, Landais A, Masson-Delmotte V. 2013. Bi-hemispheric forcing for Indo-Asian monsoon during glacial terminations. *Quat. Sci. Rev.* 59:1–4
Caley T, Malaizé B, Revel M, Ducassou E, Wainer K, et al. 2011a. Orbital timing of the Indian, East Asian and African boreal monsoons and the concept of a "global monsoon." *Quat. Sci. Rev.* 30:3705–15

Caley T, Malaizé B, Zaragosi S, Rossignol L, Bourget J, et al. 2011b. New Arabian Sea records help decipher orbital timing of Indo-Asian monsoon. *Earth Planet. Sci. Lett.* 308:433–44

Calvert SE, Fontugne MR. 2001. On the late Pleistocene-Holocene sapropel record of climatic and oceanographic variability in the eastern Mediterranean. *Paleoceanography* 16:78–94

Castro CL, Pielke RA, Adegoke JO, Schubert SD, Pegion PJ. 2007. Investigation of the summer climate of the contiguous United States and Mexico using the Regional Atmospheric Modeling System (RAMS). Part II: Model climate variability. *J. Clim.* 20:3866–87

Chakraborty A, Nanjundiah RS, Srinivasan J. 2002. Role of Asian and African orography in Indian summer monsoon. *Geophys. Res. Lett.* 29:50–51

Chang CP, Ding YH, Lau GNC, Johnson RH, Wang B, Yasunari T, eds. 2011. *The Global Monsoon System: Research and Forecast*. pp. 594. Singapore: World Scientific Publishing. 2nd ed.

Chang CP, Zhang Y, Li T. 2001. Interannual and interdecadal variations of the East Asian summer and tropical Pacific SSTs. Part I: Roles of the subtropical ridge. *J. Clim.* 13:4310–25

Chao WC, Chen BD. 2001. The origin of monsoon. *J. Atmos. Sci.* 58:3497–507

Charney JG. 1969. The intertropical convergence zone and the Hadley circulation of the atmosphere. In *Proceedings of the WMO/IUGG Symposium on Numerical Weather Prediction in Tokyo, Nov. 26–Dec. 4, 1968*, pp. 73–79. Tokyo: Jpn. Meteorol. Agency

Charney JG. 1975. Dynamics of deserts and drought in the Sahel. *Q. J. R. Meteorol. Soc.* 101:193–202

Chen FH, Bloemendal J, Wang JM, Li JJ, Oldfield F, et al. 1997. High-resolution multi-proxy climate records from Chinese loess: evidence for rapid climatic changes over the last 75 kyr. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 130:323–35

Cheng H, Edwards RL, Broecker WS, Denton GH, Kong X, et al. 2009. Ice age terminations. *Science* 326:248–52

Cheng H, Edwards RL, Wang Y, Kong X, Ming Y, et al. 2006. A penultimate glacial monsoon record from Hulu Cave and two-phase glacial terminations. *Geology* 34:217–20

Cheng H, Sinha A, Cruz FW, Wang X, Edwards RL, et al. 2013. Climate change patterns in Amazonia and biodiversity. *Nat. Commun.* 4:1411

Cheng H, Sinha A, Wang XF, Cruz FW, Edwards RL. 2012. The global paleomonsoon as seen through speleothem records from Asia and the Americas. *Clim. Dyn.* 39:1045–62

Chiang JCH, Bitz CM. 2005. Influence of high latitude ice cover on the marine intertropical convergence zone. *Clim. Dyn.* 25:477–96

Chu G, Liu J, Sun Q, Lu H, Gu Z, et al. 2002. The "Mediaeval Warm Period" drought recorded in Lake Huguangyan, tropical South China. *Holocene* 12:511–16

Clark MK, Farley KA, Zheng DW, Wang ZC, Duvall AR. 2010. Early Cenozoic faulting of the northern Tibetan Plateau margin from apatite (U-Th)/He ages. *Earth Planet. Sci. Lett.* 296:78–88

Clark P, Archer D, Pollard D, Blum JD, Rial JA, et al. 2006. The middle Pleistocene transition: characteristics, mechanisms, and implications for long-term changes in atmospheric pCO_2. *Quat. Sci. Rev.* 25:3150–84

Clemens SC. 2005. Millennial-band climate spectrum resolved and linked to centennial-scale solar cycles. *Quat. Sci. Rev.* 24:521–31

Clemens SC, Murray DW, Prell WL. 1996. Non-stationary phase of the Plio-Pleistocene Asian monsoon. *Science* 274:943–48

Clemens SC, Prell WL. 2003. A 350,000-year summer monsoon multiproxy stack from the Owen Ridge, Northern Arabian Sea. *Mar. Geol.* 201:35–51

Clemens SC, Prell W, Murray D, Shimmield G, Weedon G. 1991. Forcing mechanisms of the Indian Ocean monsoon. *Nature* 353:720–25

Clemens SC, Prell WL, Sun Y. 2010. Orbital-scale timing and mechanisms driving Late Pleistocene Indo-Asian summer monsoons: reinterpreting cave speleothem $\delta^{18}O$. *Paleoceanography* 25:PA4207

Clemens SC, Prell WL, Sun Y, Liu Z, Chen G. 2008. Southern Hemisphere forcing of Pliocene $\delta^{18}O$ and the evolution of Indo-Asian monsoons. *Palaeoceanography* 23:PA4210

Clement AC, Peterson LC. 2008. Mechanisms of abrupt climate change of the last glacial period. *Rev. Geophys.* 46:RG4002

Conroy JL, Overpeck JT, Cole JE, Shanahan TM, Steinitz-Kannan M. 2008. Holocene changes in eastern tropical Pacific climate inferred from a Galapagos lake sediment record. *Quat. Sci. Rev.* 27:1166–80

Cronin TM. 2009. *Paleoclimates: Understanding Climate Change Past and Present*. New York: Columbia Univ. Press

Cruz FW, Burns SJ, Karmann I, Sharp WD, Vuille M, et al. 2005. Insolation-driven changes in atmospheric circulation over the past 116,000 years in subtropical Brazil. *Nature* 434:63–66

Dansgaard W, Johnsent SJ, Clausen HB, Dahl-Jensen D, Gundestrup NS, et al. 1993. Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature* 364:218–20

Dash S. 2005. Monsoons and monsoon climate. In *Encyclopedia of World Climatology*. ed. J Oliver, pp. 509–16. Dordrecht, Neth.: Springer

Dayem KE, Molnar P, Battisti DS, Roe GH. 2010. Lessons learned from oxygen isotopes in modern precipitation applied to interpretation of speleothem records of paleoclimate from eastern Asia. *Earth Planet. Sci. Lett.* 295:219–30

DeCelles PG, Quade J, Kapp P, Fan M, Dettman DL, et al. 2007. High and dry in central Tibet during the Late Oligocene. *Earth Planet. Sci. Lett.* 253:389–401

deMenocal PB. 1995. Plio-Pleistocene African climate. *Science* 270:53–59

deMenocal P, Ortiz J, Guilderson T, Adkins J, Sarnthein M, et al. 2000. Abrupt onset and termination of the African Humid Period: rapid climate responses to gradual insolation forcing. *Quat. Sci. Rev.* 19:347–61

deMenocal P, Rind D. 1993. Sensitivity of Asian and African climate to variations in seasonal insolation, glacial ice cover, sea surface temperature, and Asian orography. *J. Geophys. Res.* 98(D4):7265–87

Ding Y, Chan JC. 2005. The East Asian summer monsoon: an overview. *Meteorol. Atmos. Phys.* 89:117–42

Ding YH, Liu YJ. 2001. Onset and the evolution of the summer monsoon over the South China Sea during the SCSMEX field experiment in 1998. *J. Meteorol. Soc. Jpn.* 79:255–76

Ding YH, Wang Z, Sun Y. 2008. Inter-decadal variation of the summer precipitation in East China and its association with decreasing Asian summer monsoon. Part I: Observed evidences. *Int. J. Climatol.* 28:1139–61

Ding ZL, Liu TS, Rutter RW, Yu ZW, Guo ZT, et al. 1995. Ice-volume forcing of East Asian winter monsoon variations in the past 800,000 years. *Quat. Res.* 44:149–59

Ding ZL, Rutter NW, Liu TS, Sun JM, Ren JZ, et al. 1998. Correlation of Dansgaard-Oeschger cycles between Greenland ice and Chinese loess. *Paleoclimates* 4:281–91

Ding ZL, Rutter NW, Sun JM, Yang SL, Liu TS. 2000. Re-arrangement of atmospheric circulation at about 2.6 Ma over northern China: evidence from grain size records of loess-palaeosol and red clay sequences. *Quat. Sci. Rev.* 19:547–58

Ding ZL, Xiong SF, Sun JM, Yang SL, Gu ZY, et al. 1999. Pedostratigraphy and paleomagnetism of a ~7.0 Ma eolian loess–red clay sequence at Lingtai, Loess Plateau, north-central China and the implications for paleomonsoon evolution. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 152:49–66

Douglas MW, Maddox RA, Howard K, Reyes S. 1993. The Mexican monsoon. *J. Clim.* 6:1665–77

Duan AM, Wang MR, Lei YH, Cui YF. 2013. Trends in summer rainfall over China associated with the Tibetan Plateau sensible heat source during 1980–2008. *J. Clim.* 26:261–75

Duan AM, Wu GX. 2008. Weakening trend in the atmospheric heat source over the Tibetan Plateau during recent decades. Part I: Observations. *J. Clim.* 21:3149–64

Duan AM, Wu GX, Zhang Q, Liu YM. 2006. New proofs of the recent climate warming over the Tibetan Plateau as a result of the increasing greenhouse gases emissions. *Chin. Sci. Bull.* 51:1396–400

Dupont-Nivet G, Hoorn C, Konert M. 2008. Tibetan uplift prior to the Eocene-Oligocene climate transition: evidence from pollen analysis of the Xining Basin. *Geology* 36:987–90

Dykoski CA, Edwards RL, Cheng H, Yuan D, Cai Y, et al. 2005. A high-resolution, absolute-dated Holocene and deglacial Asian monsoon record from Dongge Cave, China. *Earth Planet. Sci. Lett.* 233:71–86

Ellis AW, Saffell EM, Hawkins TW. 2004. A method for defining monsoon onset and demise in the southwestern USA. *Int. J. Climatol.* 24:247–65

Fang XM, Ono Y, Fukusawa H, Bao-Tian P, Li JJ, et al. 1999. Asian summer monsoon instability during the past 60,000 years: magnetic susceptibility and pedogenic evidence from the western Chinese Loess Plateau. *Earth Planet. Sci. Lett.* 168:219–32

Fairbridge RW. 1986. Monsoons and paleomonsoons. *Episodes* 9:143–49

Fein JS, Stephens PL. 1987. *Monsoons*. New York: Wiley

Felzer B, Webb T III, Oglseby RJ. 1998. The impact of ice sheets, CO_2, and orbital insolation on late quaternary climate: sensitivity experiments with a general circulation model. *Quat. Sci. Rev.* 17:507–34

Feng J, Li JP. 2009. Variation of the South China Sea summer monsoon and its association with the global atmosphere circulation and sea surface temperature. *Chin. J. Atmos. Sci.* 33:568–80 (in Chinese)

Feng J, Li JP, Li Y. 2010. A monsoon-like southwest Australian circulation and its relation with rainfall in southwest Western Australia. *J. Clim.* 23:1334–53

Feng ZD, Wang HB, Olson CG. 2004. Pedogenic factors affecting magnetic susceptibility of the last interglacial palaeosol S1 in the Chinese Loess Plateau. *Earth Surf. Processes Landf.* 29:1389–402

Fleitmann D, Burns SJ, Mudelsee M, Neff U, Kramers J, et al. 2003. Holocene forcing of the Indian monsoon recorded in a stalagmite from southern Oman. *Science* 300:1737–39

Flohn H. 1951. Passatzirkulation und äquatoriale Westwindzone. *Arch. Meteorol. Geophys. Bioklimatol. B* 3:3–15

Flohn H. 1957. Large-scale aspects of the "summer monsoon" in south and east Asia. *J. Meteorol. Soc. Jpn.* 75:180–86

Fontaine B, Janicot S. 1996. Sea surface temperature fields associated with West African rainfall anomaly types. *J. Clim.* 9:2935–40

Gadgil S. 2003. The Indian monsoon and its variability. *Annu. Rev. Earth Planet. Sci.* 31:429–67

Gadgil S, Srinivasan J. 1990. Low frequency variation of tropical convergence zones. *Meteorol. Atmos. Phys.* 44:119–32

Ganguly D, Rasch PJ, Wang H, Yoon JH. 2012. Climate response of the South Asian monsoon system to anthropogenic aerosols. *J. Geophys. Res.* 117:D13209

Garcin Y, Vincens A, Williamson D, Buchet G, Guiot J. 2007. Abrupt resumption of the African Monsoon at the Younger Dryas–Holocene climatic transition. *Quat. Sci. Rev.* 26:690–704

Gasse F. 2000. Hydrological changes in the African tropics since the Last Glacial Maximum. *Quat. Sci. Rev.* 19:189–211

Ghil M, Mullhaupt A, Pestiaux P. 1987. Deep water formation and Quaternary glaciations. *Clim. Dyn.* 2:1–10

Giannini A, Biasutti M, Verstraete MM. 2008. A climate model–based review of drought in the Sahel: desertification, the re-greening and climate change. *Glob. Planet. Change* 64:119–28

Goswami BN. 2006. The Asian monsoon: interdecadal variability. In *The Asian Monsoon*, ed. B Wang, pp. 295–327. London: Springer

Goswami BN, Ajayamohan R, Xavier PK, Sengupta D. 2003. Clustering of synoptic activity by Indian summer monsoon intraseasonal oscillations. *Geophys. Res. Lett.* 30:1431

Goswami BN, Xavier PK. 2005. ENSO control on the south Asian monsoon through the length of the rainy season. *Geophys. Res. Lett.* 32:L18717

Graham NE. 1994. Decadal-scale climate variability in the tropical and North Pacific during the 1970s and 1980s: observations and model results. *Clim. Dyn.* 10:135–62

Greenwood DR. 1996. Eocene monsoon forests in central Australia? *Aust. Syst. Bot.* 9:95–112

Gu C, Renaut RW. 1994. The effect of Tibetan uplift on the formation and preservation of Tertiary lacustrine source-rocks in eastern China. *J. Paleolimnol.* 11:31–40

Guo S. 1965. On discovery of fossil palms from Tertiary formation of Kwangtung and Kwangsi. *Acta Palaeontol. Sin.* 13:598–609

Guo S. 1983. Note on phytogeographic province and ecological environment of Late Cretaceous and Tertiary floras in China. In *Palaeobiogeography Provinces of China*, ed. Editorial Committee of Fundamental Theory of Palaeontology Book Series, pp. 164–77. Beijing: Science (in Chinese)

Guo ZT, Berger A, Yin QZ, Qin L. 2009. Strong asymmetry of hemispheric climates during MIS-13 inferred from correlating China loess and Antarctica ice records. *Clim. Past* 5:21–31

Guo ZT, Liu TS, Guiot J, Wu NQ, Lu HY, et al. 1996. High frequency pulses of East Asian monsoon climate in the last two glaciations: link with the North Atlantic. *Clim. Dyn.* 12:701–9

Guo ZT, Sun B, Zhang ZS, Peng SZ, Xiao GQ, et al. 2008. A major reorganization of Asian climate by the early Miocene. *Clim. Past* 4:153–74

Gupta AK, Anderson DM, Overpeck JT. 2003. Abrupt changes in the Asian southwest monsoon during the Holocene and their links to the North Atlantic Ocean. *Nature* 421:354–57

Gupta AK, Das M, Anderson DM. 2005. Solar influence on the Indian summer monsoon during the Holocene. *Geophys. Res. Lett.* 32:L17703

Hadley G. 1735. Concerning the cause of the general trade winds. *Philos. Trans. R. Soc.* 39:58–62

Hahn D, Manabe S. 1975. The role of mountains in the South Asian monsoon circulation. *J. Atmos. Sci.* 32:1515–41

Halley E. 1686. An historical account of the trade winds, and monsoons, observable in the seas between and near the tropicks, with an attempt to assign the phisical cause of the said winds. *Philos. Trans. R. Soc.* 16:153–68

Haug GH, Hughen KA, Sigman DM, Peterson LC, Röhl U. 2001. Southward migration of the Intertropical Convergence Zone through the Holocene. *Science* 293:1304–8

Heinrich H. 1988. Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years. *Quat. Res.* 29:142–52

Held IM, Hou AY. 1980. Nonlinear axially symmetric circulations in a nearly inviscid atmosphere. *J. Atmos. Sci.* 37:515–33

Heslop D, Dekkers MJ, Langereis CG. 2002. Timing and structure of the mid-Pleistocene transition: records from the loess deposits of northern China. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 185:133–43

Higgins RW, Mo KC, Yao Y. 1998. Interannual variability of the US summer precipitation regime with emphasis on the southwestern monsoon. *J. Clim.* 11:2582–606

Higgins RW, Shi W. 2000. Dominant factors responsible for interannual variability of the summer monsoon in the southwestern United States. *J. Clim.* 13:759–76

Hilgen FJ, Krijgsman W, Langereis CG, Lourens LJ, Santarelli A, et al. 1995. Extending the astronomical (polarity) time scale into the Miocene. *Earth Planet. Sci. Lett.* 136:495–510

Hodell DA, Brenner M, Curtis JH, Medina-González R, Ildefonso-Chan Can E, et al. 2005. Climate change on the Yucatan Peninsula during the Little Ice Age. *Quat. Res.* 63:109–21

Hopley PJ, Weedon GP, Marshall JD, Herries AI, Latham AG, et al. 2007. High- and low-latitude orbital forcing of early hominin habitats in South Africa. *Earth Planet. Sci. Lett.* 256:419–32

Hoskins BJ, Rodwell MJ. 1995. A model of the Asian summer monsoon. Part I: The global scale. *J. Atmos. Sci.* 52:1329–40

Hsu HH. 2012. Intraseasonal variability of the atmosphere-ocean-climate system: East Asian monsoon. In *Intraseasonal Variability in the Atmosphere-Ocean Climate System*, ed. WK Lau, DE Waliser, pp. 73–110. Berlin: Springer-Verlag

Hu Q, Feng S. 2008. Variation of the North American summer monsoon regimes and the Atlantic multidecadal oscillation. *J. Clim.* 21:2371–83

Hu Q, Feng S. 2010. Influence of the Arctic oscillation on central United States summer rainfall. *J. Geophys. Res.* 115:D01102

Huber M, Goldner A. 2012. Eocene monsoons. *J. Asian Earth Sci.* 44:3–23

Itambi AC, Von Dobeneck T, Mulitza S, Bickert T, Heslop D. 2009. Millennial-scale northwest African droughts related to Heinrich events and Dansgaard-Oeschger cycles: evidence in marine sediments from offshore Senegal. *Paleoceanography* 24:PA1205

Jian ZM, Huang BQ, Kuhnt W, Lin HL. 2001. Late Quaternary upwelling intensity and East Asian monsoon forcing in the South China Sea. *Quat. Res.* 55:363–70

Jiang YQ, Liu XH, Yang XQ, Wang MH. 2013. A numerical study of the effect of different aerosol types on East Asian summer clouds and precipitation. *Atmos. Environ.* 70:51–63

Johnson KR, Ellis B. 2002. A tropical rainforest in Colorado 1.4 million years after the Cretaceous–Tertiary boundary. *Science* 296:2379–83

Johnson TC, Barry SL, Chan Y, Wilkinson P. 2001. Decadal record of climate variability spanning the past 700 yr in the Southern Tropics of East Africa. *Geology* 29:83–86

Joseph PV, Liebmann B, Hendon HH. 1991. Interannual variability of the Australian summer monsoon onset: possible influence of Indian summer monsoon and El Niño. *J. Clim.* 4:529–38

Joseph PV, Sooraj KP, Rajan CK. 2006. The summer monsoon onset process over South Asia and an objective method for the date of monsoon onset over Kerala. *Int. J. Climatol.* 26:1871–93

Kanner LC, Burns SJ, Cheng H, Edwards RL. 2012. High-latitude forcing of the South American summer monsoon during the last glacial. *Science* 335:570–73

Kasahara A, Sasamori T, Washington WM. 1973. Simulation experiments with a 12-layer stratospheric global circulation model. I. Dynamical effect of the Earth's orography and thermal influence of continentality. *J. Atmos. Sci.* 30:1229–51

Kawamura R. 1994. A rotated EOF analysis of global sea surface temperature variability with interannual and interdecadal scales. *J. Phys. Oceanogr.* 24:707–15

Kelly MJ, Edwards RL, Cheng H, Yuan D, Cai Y, et al. 2006. High resolution characterization of the Asian Monsoon between 146,000 and 99,000 years BP from Dongge Cave, China and global correlation of events surrounding Termination II. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 236:20–38

Khromov S. 1957. Die geographische verbreitung der Monsune. *Petermanns Geogr. Mitt.* 101:234–37

Kitoh A. 2004. Effects of mountain uplift on East Asian summer climate investigated by a coupled atmosphere-ocean GCM. *J. Clim.* 17:783–802

Kodera K. 2004. Solar influence on the Indian Ocean monsoon through dynamical processes. *Geophys. Res. Lett.* 31:L24209

Krishnamurthy V, Goswami BN. 2000. Indian monsoon–ENSO relationship on interdecadal timescale. *J. Clim.* 13:579–95

Krishnamurti TN, Ardanuy P. 1980. The 10 to 20-day westward propagating mode and "Breaks in the Monsoons." *Tellus* 32:15–26

Kroon D, Steens T, Troelstra SR. 1991. Onset of monsoonal related upwelling in the western Arabian Sea as revealed by planktonic foraminifers. *Proc. Ocean Drill. Program Sci. Results* 117:257–64

Kumar KK, Rajagopalan B, Cane MA. 1999. On the weakening relationship between the Indian monsoon and ENSO. *Science* 284:2156–59

Kumar KK, Rajagopalan B, Hoerling M, Bates G, Cane M. 2006. Unraveling the mystery of Indian monsoon failure during El Niño. *Science* 314:115–19

Kutzbach JE. 1980. Estimates of past climate at Paleolake Chad, North Africa, based on a hydrological and energy-balanced model. *Quat. Res.* 14:210–23

Kutzbach JE. 1981. Monsoon climate of the early Holocene: climate experiment with the Earth's orbital parameters for 9000 years ago. *Science* 214:59–61

Kutzbach JE, Gallimore RG. 1988. Sensitivity of a coupled atmosphere/mixed layer ocean model to changes in orbital forcing at 9000 years BP. *J. Geophys. Res.* 93(D1):803–21

Kutzbach JE, Guetter PJ. 1986. The influence of changing orbital parameters and surface boundary conditions on climate simulations for the past 18 000 years. *J. Atmos. Sci.* 43:1726–59

Kutzbach JE, Guetter PJ, Ruddiman WF, Prell WL. 1989. Sensitivity of climate to late Cenozoic uplift in southern Asia and the American west: numerical experiments. *J. Geophys. Res.* 94(D15):18393–407

Kutzbach JE, Liu X, Liu Z, Chen G. 2008. Simulation of the evolutionary response of global summer monsoons to orbital forcing over the past 280,000 years. *Clim. Dyn.* 30:567–79

Kutzbach JE, Liu Z. 1997. Oceanic feedback on the western African monsoon at 6000 BP. *Science* 278:440–43

Kutzbach JE, Otto-Bliesner BL. 1982. The sensitivity of the African-Asian monsoonal climate to orbital parameter changes for 9000 years BP in a low-resolution general circulation model. *J. Atmos. Sci.* 39:1177–88

Kutzbach JE, Prell WL, Ruddiman WF. 1993. Sensitivity of Eurasian climate to surface uplift of the Tibetan Plateau. *J. Geol.* 101:177–90

Larrasoaña JC, Roberts AP, Rohling EJ, Winklhofer M, Wehausen R. 2003. Three million years of monsoon variability over the northern Sahara. *Clim. Dyn.* 21:689–98

Lau K. 2003. ENSO-monsoon interactions. In *Encyclopedia of Atmosphere Sciences*, ed. J Holton, pp. 1386–91. London: Elsevier

Lawrence DM, Webster PJ. 2002. The boreal summer intraseasonal oscillation: relationship between northward and eastward movement of convection. *J. Atmos. Sci.* 59:1593–606

Le Barbé L, Lebel T. 1997. Rainfall climatology of the HAPEX-Sahel region during the years 1950–1990. *J. Hydrol.* 43:188–89

Le Barbé L, Lebel T, Tapsoba D. 2002. Rainfall variability in West Africa during the years 1950–90. *J. Clim.* 15:187–202

Lei YH, Hoskins B, Slingo J. 2011. Exploring the interplay between natural decadal variability and anthropogenic climate change in summer rainfall over China. Part I: Observational evidence. *J. Clim.* 24:4584–99

Li C, Battisti DS, Schrag DP, Tziperman E. 2005. Abrupt climate shifts in Greenland due to displacements of the sea ice edge. *Geophys. Res. Lett.* 32:L19702

Li C, Yanai M. 1996. The onset and interannual variability of the Asian summer monsoon in relation to land-sea thermal contrast. *J. Clim.* 9:358–75

Li JP, Wu ZW, Jiang ZH, He JH. 2010. Can global warming strengthen the East Asian summer monsoon? *J. Clim.* 23:6696–705

Li JP, Zeng QC. 2000. Significance of the normalized seasonality of wind field and its rationality for characterizing the monsoon. *Sci. China D* 43:646–53

Li JP, Zeng QC. 2002. A unified monsoon index. *Geophys. Res. Lett.* 29:1274

Li JP, Zeng QC. 2003. A new monsoon index and the geographical distribution of the global monsoons. *Adv. Atmos. Sci.* 20:299–302

Li JP, Zeng QC. 2005. A new monsoon index, its interannual variability and relation with monsoon precipitation. *Clim. Environ. Res.* 10:351–65 (in Chinese)

Li JP, Zhang L. 2009. Wind onset and withdrawal of Asian summer monsoon and their simulated performance in AMIP models. *Clim. Dyn.* 32:935–68

Liang X, Liu Y, Wu G. 2005. The role of land-sea distribution in the formation of the Asian summer monsoon. *Geophys. Res. Lett.* 32:L03708

Lindzen RS, Hou AY. 1988. Hadley circulations for zonally averaged heating centered off the equator. *J. Atmos. Sci.* 45:2416–27

Lisiecki LE, Raymo ME. 2005. A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}O$ records. *Paleoceanography* 20:PA1003

Liu J, Chen F, Chen J, Xia D, Xu Q, et al. 2011. Humid Medieval Warm Period recorded by magnetic characteristics of sediments from Gonghai Lake, Shanxi, North China. *Chin. Sci. Bull.* 56:2464–74

Liu J, Wang B, Ding Q, Kuang X, Soon W, et al. 2009. Centennial variations of the global monsoon precipitation in the last millennium: results from ECHO-G model. *J. Clim.* 22:2356–71

Liu TS, Ding ZL. 1998. Chinese loess and the paleomonsoon. *Annu. Rev. Earth Planet. Sci.* 26:111–45

Liu TS, Ding ZL, Rutter N. 1999. Comparison of Milankovitch periods between continental loess and deep sea records over the last 2.5 Ma. *Quat. Sci. Rev.* 18:1205–12

Liu XD, Kutzbach JE, Liu ZY, An ZS, Li L. 2003a. The Tibetan Plateau as amplifier of orbital-scale variability of the East Asian monsoon. *Geophys. Res. Lett.* 30:1839

Liu XD, Yin ZY. 2002. Sensitivity of East Asian monsoon climate to the uplift of the Tibetan Plateau. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 183:223–45

Liu Y, Henderson G, Hu C, Mason A, Charnley N, et al. 2013. Links between the East Asian monsoon and North Atlantic climate during the 8,200 year event. *Nat. Geosci.* 6:117–20

Liu Y. Hoskins BJ, Blackburn M. 2007. Impacts of the Tibetan topography and heating on the summer flow over Asia. *J. Meteorol. Soc. Jpn.* 85B:1–19

Liu Y, Wu G, Hong J, Dong B, Duan A, et al. 2012. Revisiting Asian monsoon formation and change associated with Tibetan Plateau forcing. II: Change. *Clim. Dyn.* 39:1183–95

Liu Z, Harrison S, Kutzbach J, Otto-Bliesner B. 2004. Global monsoons in the mid-Holocene and oceanic feedback. *Clim. Dyn.* 22:157–82

Liu Z, Notaro M, Kutzbach J, Liu N. 2006. Assessing global vegetation-climate feedbacks from observation. *J. Clim.* 19:787–814

Liu Z, Otto-Bliesner B, Kutzbach J, Li L, Shields C. 2003b. Coupled climate simulation of the evolution of global monsoons in the Holocene. *J. Clim.* 16:2472–90

Liu Z, Wen X, Brady EC, Otto-Bliesner B, Yu G, Lu H, et al. 2014. Chinese cave records and the East Asia summer monsoon. *Quat. Sci. Rev.* 83:115–28

Loope DB, Rowe CM, Joeckel RM. 2001. Annual monsoon rains recorded by Jurassic dunes. *Nature* 412:64–66

Lourens LJ, Wehausen R, Brumsack HJ. 2001. Geological constraints on tidal dissipation and dynamical ellipticity of the Earth over the past three million years. *Nature* 409:1029–33

Lu HY, Zhang FQ, Liu XD, Duce RA. 2004. Periodicities of palaeoclimatic variations recorded by loess-paleosol sequences in China. *Quat. Sci. Rev.* 23:1891–900

Luo JJ, Sasaki W, Masumoto Y. 2012. Indian Ocean warming modulates Pacific climate change. *Proc. Natl. Acad. Sci. USA* 109:18701–6

Maasch KA, Saltzman B. 1990. A low-order dynamical model of global climatic variability over the full Pleistocene. *J. Geophys. Res.* 95(D2):1955–63

Madden RA, Julian PR. 1994. Observations of the 40–50-day tropical oscillation—a review. *Mon. Weather Rev.* 122:814–37

Maher BA. 2008. Holocene variability of the East Asian summer monsoon from Chinese cave records: a re-assessment. *Holocene* 18:861–66

Maley J. 1981. Etudes palynologiques dans le bassin du Tchad et paléoclimatologie de l'Afrique nord-tropicale de 30 000 ans à l'époque actuelle. Paris: ORSTOM

Manabe S, Broccoli A. 1990. Mountains and arid climates of middle latitudes. *Science* 247:192–95

Manabe S, Terpstra TB. 1974. The effects of mountains on the general circulation of the atmosphere as identified by numerical experiments. *J. Atmos. Sci.* 31:3–42

Mao JY, Chan JC. 2005. Intraseasonal variability of the South China Sea summer monsoon. *J. Clim.* 18:2388–402

Mao JY, Sun Z, Wu GX. 2010. 20–50-day oscillation of summer Yangtze rainfall in response to intraseasonal variations in the subtropical high over the western North Pacific and South China Sea. *Clim. Dyn.* 34:747–61

Mao JY, Wu GX. 2007. Interannual variability in the onset of the summer monsoon over the eastern Bay of Bengal. *Theor. Appl. Climatol.* 89:155–70

Mayewski PA, Meeker LD, Twickler MS, Whitlow S, Yang Q, et al. 1997. Major features and forcing of high-latitude Northern Hemisphere atmospheric circulation using a 110,000-year-long glaciochemical series. *J. Geophys. Res.* 102(C12):26345–66

Mayle FE, Burbridge R, Killeen TJ. 2000. Millennial-scale dynamics of southern Amazonian rain forests. *Science* 290:2291–94

Menviel L, Timmermann A, Mouchet A, Timm O. 2008. Meridional reorganizations of marine and terrestrial productivity during Heinrich events. *Paleoceanography* 23:PA1203

Milanković M. 1941. Kanon der Erdbestrahlung und seine Anwendung auf das Eiszeitenproblem. Belgrade, Serb.: Mihaila Ćurčića

Moernaut J, Verschuren D, Charlet F, Kristen I, Fagot M, et al. 2010. The seismic-stratigraphic record of lake-level fluctuations in Lake Challa: hydrological stability and change in equatorial East Africa over the last 140 kyr. *Earth Planet. Sci. Lett.* 290:214–23

Molnar P. 2005. Mio-Pliocene growth of the Tibetan Plateau and evolution of East Asian climate. *Palaeontol. Electron.* 8:1–23

Molnar P, Boos WR, Battisti DS. 2010. Orographic controls on climate and paleoclimate of Asia: thermal and mechanical roles for the Tibetan Plateau. *Annu. Rev. Earth Planet. Sci.* 38:77–102

Molnar P, England P, Martinod J. 1993. Mantle dynamics, uplift of the Tibetan Plateau, and the Indian monsoon. *Rev. Geophys.* 31:357–96

Molnar P, Stock JM. 2009. Slowing of India's convergence with Eurasia since 20 Ma and its implications for Tibetan mantle dynamics. *Tectonics* 28:TC3001

Moy CM, Seltzer GO, Rodbell DT, Anderson DM. 2002. Variability of El Niño/Southern Oscillation activity at millennial timescales during the Holocene epoch. *Nature* 420:162–65

Murton JB, Bateman MD, Dallimore SR, Teller JT, Yang Z. 2010. Identification of Younger Dryas outburst flood path from Lake Agassiz to the Arctic Ocean. *Nature* 464:740–43

Nan S, Li J. 2003. The relationship between the summer precipitation in the Yangtze River valley and the boreal spring Southern Hemisphere annular mode. *Geophys. Res. Lett.* 30:2266

Neff U, Burns S, Mangini A, Mudelsee M, Fleitmann D, et al. 2001. Strong coherence between solar variability and the monsoon in Oman between 9 and 6 kyr ago. *Nature* 411:290–93

Newton A, Thunell R, Stott L. 2006. Climate and hydrographic variability in the Indo-Pacific Warm Pool during the last millennium. *Geophys. Res. Lett.* 33:L19710

Nie J, Boos WR, Kuang Z. 2010. Observational evaluation of a convective quasi-equilibrium view of monsoons. *J. Clim.* 23:4416–28

Niedermeyer EM, Schefuß E, Sessions AL, Mulitza S, Mollenhauer G, et al. 2010. Orbital-and millennial-scale changes in the hydrologic cycle and vegetation in the western African Sahel: insights from individual plant wax δD and δ^{13}C. *Quat. Sci. Rev.* 29:2996–3005

Novello VF, Cruz FW, Karmann I, Burns SJ, Stríkis NM, et al. 2012. Multidecadal climate variability in Brazil's Nordeste during the last 3000 years based on speleothem isotope records. *Geophys. Res. Lett.* 39:L23706

Oppo DW, Rosenthal Y, Linsley BK. 2009. 2,000-year-long temperature and hydrology reconstructions from the Indo-Pacific warm pool. *Nature* 460:1113–16

Oppo DW, Sun Y. 2005. Amplitude and timing of sea-surface temperature change in the northern South China Sea: dynamic link to the East Asian monsoon. *Geology* 33:785–88

Overpeck J, Anderson D, Trumbore S, Prell W. 1996. The southwest Indian Monsoon over the last 18,000 years. *Clim. Dyn.* 12:213–25

Overpeck JT, Cole JE. 2006. Abrupt change in Earth's climate system. *Annu. Rev. Environ. Resour.* 31:1–31

Parrish JT, Ziegler A, Scotese CR. 1982. Rainfall patterns and the distribution of coals and evaporites in the Mesozoic and Cenozoic. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 40:67–101

Parthasarathy B, Munot A, Kothawale D. 1994. All-India monthly and seasonal rainfall series: 1871–1993. *Theor. Appl. Climatol.* 49:217–24

Partridge T, deMenocal P, Lorentz S, Paiker M, Vogel J. 1997. Orbital forcing of climate over South Africa: a 200,000-year rainfall record from the Pretoria Saltpan. *Quat. Sci. Rev.* 16:1125–33

Pausata FS, Battisti DS, Nisancioglu KH, Bitz CM. 2011. Chinese stalagmite $\delta^{18}O$ controlled by changes in the Indian monsoon during a simulated Heinrich event. *Nat. Geosci.* 4:474–80

Pédelaborde P. 1963. *The Monsoon*, transl. MJ Clegg. London: Methuen

Petersen S, Schrag D, Clark P. 2013. A new mechanism for Dansgaard-Oeschger cycles. *Paleoceanography* 28:24–30

Petit JR, Jouzel J, Raynaud D, Barkov N, Barnola JM, et al. 1999. Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399:429–36

Pierrehumbert RT. 2000. Climate change and the tropical Pacific: the sleeping dragon wakes. *Proc. Natl. Acad. Sci. USA* 97:1355–58

Pierrehumbert RT, Abbot DS, Voigt A, Koll D. 2011. Climate of the Neoproterozoic. *Annu. Rev. Earth Planet. Sci.* 39:417–60

Pisias NG, Clark PU, Brook EJ. 2010. Modes of global climate variability during marine isotope stage 3 (60–26 ka). *J. Clim.* 23:1581–88

Plumb RA, Hou AY. 1992. The response of a zonally symmetric atmosphere to subtropical thermal forcing: threshold behavior. *J. Atmos. Sci.* 49:1790–99

Porter SC, An ZS. 1995. Correlation between climate events in the North Atlantic and China during the last glaciation. *Nature* 375:185–88

Prell WL. 1984. Monsoonal climate of the Arabian Sea during the Late Quaternary: a response to changing solar radiation. In *Milankovitch and Climate*, ed. AL Berger, pp. 349–66. Dordrecht, Neth.: Reidel

Prell WL, Kutzbach JE. 1987. Monsoon variability over the past 150,000 years. *J. Geophys. Res.* 92(D7):8411–25

Prell WL, Kutzbach JE. 1992. Sensitivity of the Indian monsoon to forcing parameters and implications for its evolution. *Nature* 360:647–52

Privé NC, Plumb RA. 2007. Monsoon dynamics with interactive forcing. Part II: Impact of eddies and asymmetric geometries. *J. Atmos. Sci.* 64:1431–42

Qian WH. 2000. Dry/wet alternation and global monsoon. *Geophys. Res. Lett.* 27:3679–82

Qiang X, An Z, Song Y, Chang H, Sun Y, et al. 2011. New eolian red clay sequence on the western Chinese Loess Plateau linked to onset of Asian desertification about 25 Ma ago. *Sci. China D* 54:136–44

Qiang X, Li Z, Powell CM, Zheng H. 2001. Magnetostratigraphic record of the Late Miocene onset of the East Asian monsoon, and Pliocene uplift of northern Tibet. *Earth Planet. Sci. Lett.* 187:83–93

Qiu Z, Li C. 2005. Evolution of Chinese mammalian faunal regions and elevation of the Qinghai-Xizang (Tibet) Plateau. *Sci. China D* 48:1246–58

Raicich F, Pinardi N, Navarra A. 2003. Teleconnections between Indian monsoon and Sahel rainfall and the Mediterranean. *Int. J. Climatol.* 23:173–86

Rajagopalan B, Molnar P. 2012. Pacific Ocean sea-surface temperature variability and predictability of rainfall in the early and late parts of the Indian summer monsoon season. *Clim. Dyn.* 39:1543–57

Ramage CS. 1971. *Monsoon Meteorology*. New York: Academic

Ramstein G, Fluteau F, Besse J, Joussaume S. 1997. Effect of orogeny, plate motion and land-sea distribution on Eurasian climate change over the past 30 million years. *Nature* 386:788–95

Raymo M, Lisiecki L, Nisancioglu KH. 2006. Plio-Pleistocene ice volume, Antarctic climate, and the global δ^{18}O record. *Science* 313:492–95

Reichart GJ, Lourens LJ, Zachariasse WJ. 1998. Temporal variability in the northern Arabian Sea oxygen minimum zone (OMZ) during the last 225,000 years. *Paleoceanography* 13:607–21

Reimer PJ, Baillie MG, Bard E, Bayliss A, Beck JW, et al. 2009. IntCal09 and Marine09 radiocarbon age calibration curves, 0–50,000 years cal BP. *Radiocarbon* 51:1111–50

Reuter J, Stott L, Khider D, Sinha A, Cheng H, Edwards RL. 2009. A new perspective on the hydroclimate variability in northern South America during the Little Ice Age. *Geophys. Res. Lett.* 36:L21706

Rodwell MJ, Hoskins BJ. 1996. Monsoons and the dynamics of deserts. *Q. J. R. Meteorol. Soc.* 122:1385–404

Rossignol-Strick M. 1983. African monsoons, an immediate climate response to orbital insolation. *Nature* 304:46–49

Rowley DB, Currie BS. 2006. Palaeo-altimetry of the late Eocene to Miocene Lunpola Basin, central Tibet. *Nature* 439:677–81

Ruddiman W, Prell W, Raymo M. 1989. Late Cenozoic uplift in southern Asia and the American West: rationale for general circulation modeling experiments. *J. Geophys. Res.* 94(D15):18379–91

Russell JM, Johnson TC. 2005. A high-resolution geochemical record from Lake Edward, Uganda Congo and the timing and causes of tropical African drought during the late Holocene. *Quat. Sci. Rev.* 24:1375–89

Sachs JP, Sachse D, Smittenberg RH, Zhang Z, Battisti DS, et al. 2009. Southward movement of the Pacific intertropical convergence zone AD 1400–1850. *Nat. Geosci.* 2:519–25

Sankar-Rao M. 1966. Equations for global monsoons and toroidal circulations in the σ-coordinate system. *Pure Appl. Geophys.* 65:196–215

Sankar-Rao M. 1970. On global monsoons—further results. *Tellus* 22:648–54

Sarnthein M, Tetzlaff G, Koopmann B, Wolter K, Pflaumann U. 1981. Glacial and interglacial wind regimes over the eastern subtropical Atlantic and North-West Africa. *Nature* 293:193–96

Schneider EK. 1987. A simplified model of the modified Hadley circulation. *J. Atmos. Sci.* 44:3311–28

Schneider EK, Lindzen RS. 1977. Axially symmetric steady state models of the basic state of instability and climate studies. Part I: Linearized calculations. *J. Atmos. Sci.* 34:253–79

Schulz H, von Rad U, Erlenkeuser H. 1998. Correlation between Arabian Sea and Greenland climate oscillations of the past 110,000 years. *Nature* 393:54–57

Sepulchre P, Ramstein G, Fluteau F, Schuster M, Tiercelin JJ, et al. 2006. Tectonic uplift and Eastern Africa aridification. *Science* 313:1419–23

Shi Z, Liu X, Liu Y, Sha Y, Xu T. 2014. Impact of Mongolian Plateau versus Tibetan Plateau on the westerly jet over North Pacific Ocean. *Clim. Dyn.* In press. doi: 10.1007/s00382-014-2217-2

Shi Z, Liu X, An Z, Yi B, Yang P, et al. 2011. Simulated variations of eolian dust from inner Asian deserts at the mid-Pliocene, last glacial maximum, and present day: contributions from the regional tectonic uplift and global climate change. *Clim. Dyn.* 37:2289–301

Shindell DT, Schmidt GA, Mann ME, Rind D, Waple A. 2001. Solar forcing of regional climate change during the Maunder Minimum. *Science* 294:2149–52

Sikka D, Gadgil S. 1980. On the maximum cloud zone and the ITCZ over Indian, longitudes during the southwest monsoon. *Mon. Weather Rev.* 108:1840–53

Sima A, Paul A, Schulz M. 2004. The Younger Dryas—an intrinsic feature of late Pleistocene climate change at millennial timescales. *Earth Planet. Sci. Lett.* 222:741–50

Sinha A, Cannariato KG, Stott LD, Li HC, You CF, et al. 2005. Variability of Southwest Indian summer monsoon precipitation during the Bølling-Allerød. *Geology* 33:813–16

Sinha A, Stott L, Berkelhammer M, Cheng H, Edwards RL, et al. 2011. A global context for megadroughts in monsoon Asia during the past millennium. *Quat. Sci. Rev.* 30:47–62

Sirocko F, Sarnthein M, Erlenkeuser H, Lange H, Arnold M, et al. 1993. Century-scale events in monsoonal climate over the past 24,000 years. *Nature* 364:322–24

Stager JC, Ryves D, Cumming BF, Meeker LD, Beer J. 2005. Solar variability and the levels of Lake Victoria, East Africa, during the last millenium. *J. Paleolimnol.* 33:243–51

Stríkis NM, Cruz FW, Cheng H, Karmann I, Edwards RL, et al. 2011. Abrupt variations in South American monsoon rainfall during the Holocene based on a speleothem record from central-eastern Brazil. *Geology* 39:1075–78

Sun DH, An ZS, Shaw J, Bloemendal J, Sun YB. 1998. Magnetostratigraphy and palaeoclimatic significance of late Tertiary aeolian sequences in the Chinese Loess Plateau. *Geophys. J. Int.* 134:207–12

Sun XJ, Wang PX. 2005. How old is the Asian monsoon system? Palaeobotanical records from China. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 222:181–222

Sun YB, An ZS. 2004. An improved comparison of Chinese loess with deep-sea δ^{18}O record over the interval 1.6–2.6 Ma. *Geophys. Res. Lett.* 31:L13210

Sun YB, An ZS, Clemens SC, Bloemendal J, Vandenberghe J. 2010. Seven million years of wind and precipitation variability on the Chinese Loess Plateau. *Earth Planet. Sci. Lett.* 297:525–35

Sun YB, Chen J, Clemens SC, Liu QS, Ji JF, et al. 2006a. East Asian monsoon variability over the last seven glacial cycles recorded by a loess sequence from the northwestern Chinese Loess Plateau. *Geochem. Geophys. Geosyst.* 7:Q12Q02

Sun YB, Clemens SC, An ZS, Yu ZW. 2006b. Astronomical timescale and palaeoclimatic implication of stacked 3.6-Myr monsoon records from the Chinese Loess Plateau. *Quat. Sci. Rev.* 25:33–48

Sun YB, Clemens SC, Morrill C, Lin X, Wang X, et al. 2012. Influence of Atlantic meridional overturning circulation on the East Asian winter monsoon. *Nat. Geosci.* 5:46–49

Talbot M, Delibrias G. 1977. Holocene variations in the level of Lake Bosumtwi, Ghana. *Nature* 268:722–24

Talbot M, Filippi ML, Jensen NB, Tiercelin JJ. 2007. An abrupt change in the African monsoon at the end of the Younger Dryas. *Geochem. Geophys. Geosyst.* 8:Q03005

Tan L, Cai Y, An Z, Yi L, Zhang H, Qin S. 2011. Climate patterns in north central China during the last 1800 yr and their possible driving force. *Clim. Past* 7:685–92

Tan L, Cai Y, Cheng H, An Z, Edwards RL. 2009. Summer monsoon precipitation variations in central China over the past 750 years derived from a high-resolution absolute-dated stalagmite. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 280:432–39

Tan L, Cai Y, Yi L, An Z, Ai L. 2008. Precipitation variations of Longxi, northeast margin of Tibetan Plateau since AD 960 and their relationship with solar activity. *Clim. Past* 4:19–28

Tang H, Micheels A, Eronen JT, Ahrens B, Fortelius M. 2013. Asynchronous responses of East Asian and Indian summer monsoons to mountain uplift shown by regional climate modelling experiments. *Clim. Dyn.* 40:1531–49

Tang MC, Reiter ER. 1984. Plateau monsoons of the Northern Hemisphere: a comparison between North America and Tibet. *Mon. Weather Rev.* 112:617–37

Tao SY, Chen LX. 1987. A review of recent research on the East Asian summer monsoon in China. In *Monsoon Meteorology*, ed. CP Chang, TN Krishnamurti, pp. 60–92. Oxford, UK: Oxford Univ. Press

Thompson LG, Mosley-Thompson E, Dansgaard W, Grootes PM. 1986. The Little Ice Age as recorded in the stratigraphy of the tropical Quelccaya ice cap. *Science* 234:361–64

Tiedemann R, Sarnthein M, Shackleton NJ. 1994. Astronomical timescale for the Pliocene Atlantic δ^{18}O and dust flux records of Ocean Drilling Program site 659. *Paleoceanography* 9:619–38

Tierney J, Oppo D, Rosenthal Y, Russell J, Linsley B. 2010. Coordinated hydrological regimes in the Indo-Pacific region during the past two millennia. *Paleoceanography* 25:PA1102

Tierney JE, Smerdon JE, Anchukaitis KJ, Seager R. 2013. Multidecadal variability in East African hydroclimate controlled by the Indian Ocean. *Nature* 493:389–92

Tipple BJ, Pagani M. 2007. The early origins of terrestrial C_4 photosynthesis. *Annu. Rev. Earth Planet. Sci.* 35:435–61

Tokinaga H, Xie SP, Deser C, Kosaka Y, Okumura YM. 2012. Slowdown of the Walker circulation driven by tropical Indo-Pacific warming. *Nature* 491:439–43

Tomas RA, Holton JR, Webster PJ. 1999. The influence of cross-equatorial pressure gradient on the location of near-equatorial convection. *Q. J. R. Meteorol. Soc.* 125:1107–27

Tomas RA, Webster PJ. 1997. The role of inertial instability in determining the location and strength of near-equatorial convection. *Q. J. R. Meteorol. Soc.* 123:1445–82

Torrence C, Webster P. 1999. Interdecadal changes in the ENSO-monsoon system. *J. Clim.* 12:2679–90

Trauth MH, Larrasoaña JC, Mudelsee M. 2009. Trends, rhythms and events in Plio-Pleistocene African climate. *Quat. Sci. Rev.* 28:399–411

Trenberth KE, Stepaniak DP, Caron JM. 2000. The global monsoon as seen through the divergent atmospheric circulation. *J. Clim.* 13:3969–93

Tuenter E, Weber SL, Hilgen FJ, Lourens LJ, Ganopolski A. 2005. Simulation of climate phase lags in response to precession and obliquity forcing and the role of vegetation. *Clim. Dyn.* 24:279–95

Verschuren D, Laird KR, Cumming BF. 2000. Rainfall and drought in equatorial east Africa during the past 1,100 years. *Nature* 403:410–14

Vieira LEA, Solanki SK, Krivova NA, Usoskin I. 2011. Evolution of the solar irradiance during the Holocene. *Astron. Astrophys.* 531:A6

Voelker AHL. 2002. Global distribution of centennial-scale records for Marine Isotope Stage (MIS) 3: a database. *Quat. Sci. Rev.* 21:1185–212

Voyeikov AI. 1879. Regional monsoon climate over East Asia: Amur region, Baikal, Manchuria, Eastern Mongolia, China, Japan, etc. *Proc. Imp. Russ. Geogr. Soc.* 15:321–410

Wagner JDM, Cole JE, Beck JW, Patchett PJ, Henderson GM, et al. 2010. Moisture variability in the southwestern United States linked to abrupt glacial climate change. *Nat. Geosci.* 3:110–13

Wang B, Ding QH. 2006. Changes in global monsoon precipitation over the past 56 years. *Geophys. Res. Lett.* 33:L06711

Wang B, Ding QH. 2008. Global monsoon: dominant mode of annual variation in the tropics. *Dyn. Atmos. Oceans* 44:165–83

Wang B, Liu J, Kim HJ, Webster PJ, Yim SY. 2012. Recent change of the GM precipitation (1979–2008). *Clim. Dyn.* 39:1123–35

Wang B, Wu R, Lau KM. 2001a. Interannual variability of the Asian summer monsoon: contrasts between the Indian and the western North Pacific–East Asian monsoons. *J. Clim.* 14:4073–90

Wang B, Yang J, Zhou TJ. 2008a. Interdecadal changes in the major modes of Asian-Australian monsoon variability: strengthening relationship with ENSO since the late 1970s. *J. Clim.* 21:1771–89

Wang C, Zhao X, Liu Z, Lippert PC, Graham SA, et al. 2008b. Constraints on the early uplift history of the Tibetan Plateau. *Proc. Natl. Acad. Sci. USA* 105:4987–92

Wang L, Sarnthein M, Erlenkeuser H, Grimalt J, Grootes P, et al. 1999a. East Asian monsoon climate during the late Pleistocene: high-resolution sediment records from the South China Sea. *Mar. Geol.* 156:245–84

Wang LJ, Sarnthein M, Grootes PM, Erlenkeuser H. 1999b. Millennial reoccurrence of century-scale abrupt events of East Asian monsoon: a possible heat conveyor for the global deglaciation. *Paleoceanography* 14:725–31

Wang PX. 2009. Global monsoon in a geological perspective. *Chin. Sci. Bull.* 54:1113–36

Wang PX, Clemens S, Beaufort L, Braconnot P, Ganssen G, et al. 2005a. Evolution and variability of the Asian monsoon system: state of the art and outstanding issues. *Quat. Sci. Rev.* 24:595–629

Wang PX, Jian ZM, Zhao QH, Li QY, Wang RJ, et al. 2003. Evolution of the South China Sea and monsoon history revealed in deep-sea records. *Chin. Sci. Bull.* 48:2549–61

Wang X, Auler AS, Edwards RL, Cheng H, Cristalli PS, et al. 2004. Wet periods in northeastern Brazil over the past 210 kyr linked to distant climate anomalies. *Nature* 432:740–43

Wang X, Auler AS, Edwards RL, Cheng H, Ito E, et al. 2006. Interhemispheric anti-phasing of rainfall during the last glacial period. *Quat. Sci. Rev.* 25:3391–403

Wang YJ, Cheng H, Edwards RL, An ZS, Wu JY, et al. 2001b. A high-resolution absolute-dated Late Pleistocene monsoon record from Hulu Cave, China. *Science* 294:2345–48

Wang YJ, Cheng H, Edwards RL, He Y, Kong XG, et al. 2005b. The Holocene Asian monsoon: links to solar changes and North Atlantic climate. *Science* 308:854–57

Wang YJ, Cheng H, Edwards RL, Kong XG, Shao XH, et al. 2008c. Millennial- and orbital-scale changes in the East Asian monsoon over the past 224,000 years. *Nature* 28:1090–93

Wang Z, Mysak LA. 2006. Glacial abrupt climate changes and Dansgaard–Oeschger oscillations in a coupled climate model. *Paleoceanography* 21:PA2001

Wasson RJ, Bayliss P. 2010. River flow and climate in the "Top End" of Australia from the last 1000 years, and the Asian-Australian monsoon. In *Kakadu National Park Landscape Symposia Series 2007–2009. Symposium 4: Climate Change*, ed. S Winderlich, pp. 15–36. Intern. Rep. 567. Darwin, Aust.: Supervising Scientist

WCRP (World Clim. Res. Programme). 2009. *The World Climate Research Programme Implementation Plan 2010–2015*. Geneva: WCRP

Webster PJ, Fasullo J. 2003. Monsoon: dynamical theory. In *Encyclopedia of Atmospheric Sciences*, ed. J Holton, JA Curry, pp. 1370–85. San Diego, CA: Academic

Webster PJ, Magaña VO, Palmer TN, Shukla J, Tomas RA, et al. 1998. Monsoons: processes, predictability, and the prospects for prediction. *J. Geophys. Res.* 103(C7):14451–510

Wehausen R, Brumsack HJ. 2000. Chemical cycles in Pliocene sapropel-bearing and sapropel-barren eastern Mediterranean sediments. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 158:325–52

Wehausen R, Brumsack HJ. 2002. Astronomical forcing of the East Asian monsoon mirrored by the composition of Pliocene South China Sea sediments. *Earth Planet. Sci. Lett.* 201:621–36

Wei G, Li XH, Liu Y, Shao L, Liang X. 2006. Geochemical record of chemical weathering and monsoon climate change since the early Miocene in the South China Sea. *Paleoceanography* 21:PA4214

Weldeab S, Lea DW, Schneider RR, Andersen N. 2007a. 155,000 years of West African monsoon and ocean thermal evolution. *Science* 316:1303–7

Weldeab S, Lea DW, Schneider RR, Andersen N. 2007b. Centennial scale climate instabilities in a wet early Holocene West African monsoon. *Geophys. Res. Lett.* 34:L24702

Wheeler MC, McBride JL. 2012. Australasian monsoon. In *Intraseasonal Variability in the Atmosphere-Ocean Climate System*, ed. WK Lau, DE Waliser, pp. 147–97. Berlin: Springer-Verlag

Wolff C, Haug GH, Timmermann A, Damsta JSS, Brauer A, et al. 2011. Reduced interannual rainfall variability in East Africa during the last ice age. *Science* 333:743–47

Wolff EW, Chappellaz J, Blunier T, Rasmussen SO, Svensson A. 2010. Millennial-scale variability during the last glacial: the ice core record. *Quat. Sci. Rev.* 29:2828–38

Wu GX, Li W, Guo H, Liu H, Xue J, et al. 1997. The sensible heat driven air pump over the Tibetan Plateau and the Asian summer monsoon. In *Collection in Memory of Dr. Zhao Jiuzhang*, ed. D Ye, pp. 116–26. Beijing: Science

Wu GX, Liu Y, Dong B, Liang X, Duan A, et al. 2012a. Revisiting Asian monsoon formation and change associated with Tibetan Plateau forcing. I: Formation. *Clim. Dyn.* 39:1169–81

Wu GX, Liu Y, He B, Bao Q, Duan A, et al. 2012b. Thermal controls on the Asian summer monsoon. *Sci. Rep.* 2:404

Wu GX, Liu Y, Liu X, Duan A, Liang X. 2005. How the heating over the Tibetan Plateau affects the Asian climate in summer. *Chin. J. Atmos. Sci.* 29:47–56

Wu GX, Liu Y, Wang T, Wan R, Liu X, et al. 2007. The influence of mechanical and thermal forcing by the Tibetan Plateau on Asian climate. *J. Hydrometeorol.* 8:770–89

Wu GX, Pan B, Gao H, Guan Q, Xia D. 2006. Climatic signals in the Chinese loess record for the Last Glacial: the influence of northern high latitudes and the tropical Pacific. *Quat. Int.* 154–55:128–35

Wu GX, Zhang Y. 1998. Tibetan Plateau forcing and the monsoon onset over South Asia and the South China Sea. *Mon. Weather Rev.* 126:913–27

Xiao J, An Z. 1999. Three large shifts in East Asian monsoon circulation indicated by loess-paleosol sequences in China and late Cenozoic deposits in Japan. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 154:179–89

Xiao J, Porter SC, An Z, Kumai H, Yoshikawa S. 1995. Grain size of quartz as an indicator of winter monsoon strength on the Loess Plateau of central China during the last 130,000 yr. *Quat. Res.* 43:22–29

Yan H, Sun L, Oppo DW, Wang Y, Liu Z, et al. 2011. South China Sea hydrological changes and Pacific Walker circulation variations over the last millennium. *Nat. Commun.* 2:293

Yasunari T. 1991. The monsoon year—a new concept of the climate year in the tropics. *Bull. Am. Meteorol. Soc.* 72:1331–38

Yeh TC. 1952. The seasonal variation of the influence of Tibetan Plateau on the general circulation. *Acta Meteorol. Sin.* 23:33–47 (in Chinese)

Yeh TC, Luo S, Chu P. 1957. The wind structure and heat balance in the lower troposphere over Tibetan Plateau and its surrounding. *Acta Meteorol. Sin.* 28:108–21 (in Chinese)

Yu SY, Colman SM, Lowell TV, Milne GA, Fisher TG, et al. 2010. Freshwater outburst from Lake Superior as a trigger for the cold event 9300 years ago. *Science* 328:1262–66

Yuan DX, Cheng H, Edwards RL, Dykoski CA, Kelly MJ, et al. 2004. Timing, duration, and transitions of the Last Interglacial Asian Monsoon. *Science* 304:575–78

Yuan DY, Ge WP, Chen ZW, Li CY, Wang ZC, et al. 2013. The growth of northeastern Tibet and its relevance to large-scale continental geodynamics: a review of recent studies. *Tectonics* 32:1358–70

Zachos J, Pagani M, Sloan L, Thomas E, Billups K. 2001. Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* 292:686–93

Zarriess M, Mackensen A. 2010. The tropical rainbelt and productivity changes off northwest Africa: a 31,000-year high-resolution record. *Mar. Micropaleontol.* 76:76–91

Zeng QC. 2005. About King Shun's poem "Southerly wind." *Clim. Environ. Res.* 10:283–84 (in Chinese)

Zeng QC, Li JP. 2002. Interactions between the Northern and Southern Hemispheric atmospheres and the essence of monsoon. *Chin. J. Atmos. Sci.* 26:207–26

Zeng Y, Chen J, Zhu Z, Li J, Wang J, Wan G. 2011. The wet Little Ice Age recorded by sediments in Huguangyan Lake, tropical South China. *Quat. Int.* 263:55–62

Zhang P, Cheng H, Edwards R, Chen F, Wang Y, et al. 2008. A test of climate, sun, and culture relationships from an 1810-year Chinese cave record. *Science* 322:940–42

Zhang R, Delworth TL. 2005. Simulated tropical response to a substantial weakening of the Atlantic thermohaline circulation. *J. Clim.* 18:1853–60

Zhang R, Kang SM, Held IM. 2010. Sensitivity of climate change induced by the weakening of the Atlantic meridional overturning circulation to cloud feedback. *J. Clim.* 23:378–89

Zhang R, Liu XD. 2010. The effects of tectonic uplift on the evolution of Asian summer monsoon climate since Pliocene. *Chin. J. Geophys.* 53:2817–28

Zhang XY, Arimoto R, An ZS. 1997. Dust emission from Chinese desert sources linked to variation in atmospheric circulation. *J. Geophys. Res.* 102(D23):28041–47

Zhao P, Zhou ZJ, Liu JP. 2007a. Variability of Tibetan spring snow and its associations with the hemispheric extratropical circulation and East Asian summer monsoon rainfall: an observational investigation. *J. Clim.* 20:3942–55

Zhao P, Zhu Y, Zhang R. 2007b. An Asian-Pacific teleconnection in summer tropospheric temperature and associated Asian climate variability. *Clim. Dyn.* 29:293–303

Zheng HB, Powell CM, An Z, Zhou J, Dong G. 2000. Pliocene uplift of the northern Tibetan Plateau. *Geology* 28:715–18

Zhou J, Lau KM. 1998. Does a monsoon climate exist over South America? *J. Clim.* 11:1020–40

Zhou J, Lau KM. 2001. Principal modes of interannual and decadal variability of summer rainfall over South America. *Int. J. Climatol.* 21:1623–44

Zhou T. 1984. *Chinese Natural Geography. Paleogeography 1.* Beijing: Science (in Chinese)

Zhou T, Gong D, Li J, Li B. 2009a. Detecting and understanding the multi-decadal variability of the East Asian summer monsoon—recent progress and state of affairs. *Meteorol. Z.* 18:455–67

Zhou T, Yu R, Zhang J, Drange H, Cassou C, et al. 2009b. Why the Western Pacific subtropical high has extended westward since the late 1970s. *J. Clim.* 22:2199–215

Zhou WJ, Head MJ, An ZS, Deckker PD, Liu ZY, et al. 2001. Terrestrial evidence for a spatial structure of tropical-polar interconnections during the Younger Dryas episode. *Earth Planet. Sci. Lett.* 191:231–39

Ziegler M, Lourens LJ, Tuenter E, Hilgen F, Reichart GJ, et al. 2010a. Precession phasing offset between Indian summer monsoon and Arabian Sea productivity linked to changes in Atlantic overturning circulation. *Paleoceanography* 25:PA3213

Ziegler M, Tuenter E, Lourens LJ. 2010b. The precession phase of the boreal summer monsoon as viewed from the eastern Mediterranean (ODP site 968). *Quat. Sci. Rev.* 29:1481–90