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Evolution and variability of the Asian monsoon system: state of the art and outstanding issues

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Abstract

The Asian monsoon is comprised of the Indian and East Asian subsystems. These two components are linked to one another in varying degrees by regions of strong sensible heating (Indo-Asian landmass) and strong latent heat export (the Western Pacific Warm Pool and the southern subtropical Indian Ocean). Variability within the Indian and East Asian subsystems, interactions among them, and the extent to which they interact with other climate phenomena (e.g., ENSO) are current topics of modern and paleoclimate research. This work provides an overview of past and current paleomonsoon research on tectonic to interannual time scales with a primary focus on marine sediment records. While past and current work has contributed greatly to our understanding of paleomonsoon variability at all time scales, additional efforts are required to make further progress in two critical areas. (1) Additional efforts are needed in terms of proxy development and evaluation, requiring concerted efforts at long-term sediment trap deployments in key monsoon-influenced regions as well as development of adequate and widely available core-top databases. These are necessary to assess the impact of modern oceanographic and seafloor processes on potential monsoon proxies. (2) Additional effort is also needed in acquiring a sufficient geographic distribution of downcore records to assess linkages among the subsystems and their role in the context of extratropical climate change. These records should have high sedimentation rates (including varved sections). This requires substantial survey support to identify the most appropriate coring and drilling targets.

1. Introduction

The Asian monsoon system, reaching from the western Arabian Sea through East Asia and North Australia, is a dynamic component of the modern climate system; changes in this convectively active region can result in severe draught or flood over large, densely populated regions (Webster et al., 1998). The inherent seasonality of monsoon circulation results in cool, dry winters and warm, wet summers over the continents. These seasonal changes in atmospheric circulation and precipitation also affect the ocean, leading to strong seasonality in current strength and direction, sea-surface temperature (SST) and salinity patterns, as is observed in the Indian Ocean and the South China Sea (SCS). In specific regions, such as the Northwest Arabian Sea, these dynamics lead to well-defined seasonal upwelling regimes in the open ocean and near-shore environments.

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The Asian monsoon is composed of two subsystems, the Indian (or South Asian) monsoon and the East Asian monsoon, roughly divided at ~105°E. These two subsystems are linked in that they both respond to the strength of the continental high- and low-pressure cells, which grow and decay seasonally over the Asian landmass (Fig. 1). However, they also have significant differences dictated by the contrasting sea–land distributions. The Indian system is characterized by land in the north and ocean in the south whereas the East Asian system has land in the north and south, a maritime continent in the west, and open ocean to the east. These geographic boundary conditions lead to significant differences in the relative strengths of the summer- and winter-monsoon regimes as well as to differences in their sensitivity to internal feedback mechanisms (B. Wang et al., 2003). The reader is referred to recent syntheses by Dodson et al. (2004), Mitra et al. (2002) and Fu et al. (2002) for climate and environmental overviews of the Asian monsoon region.

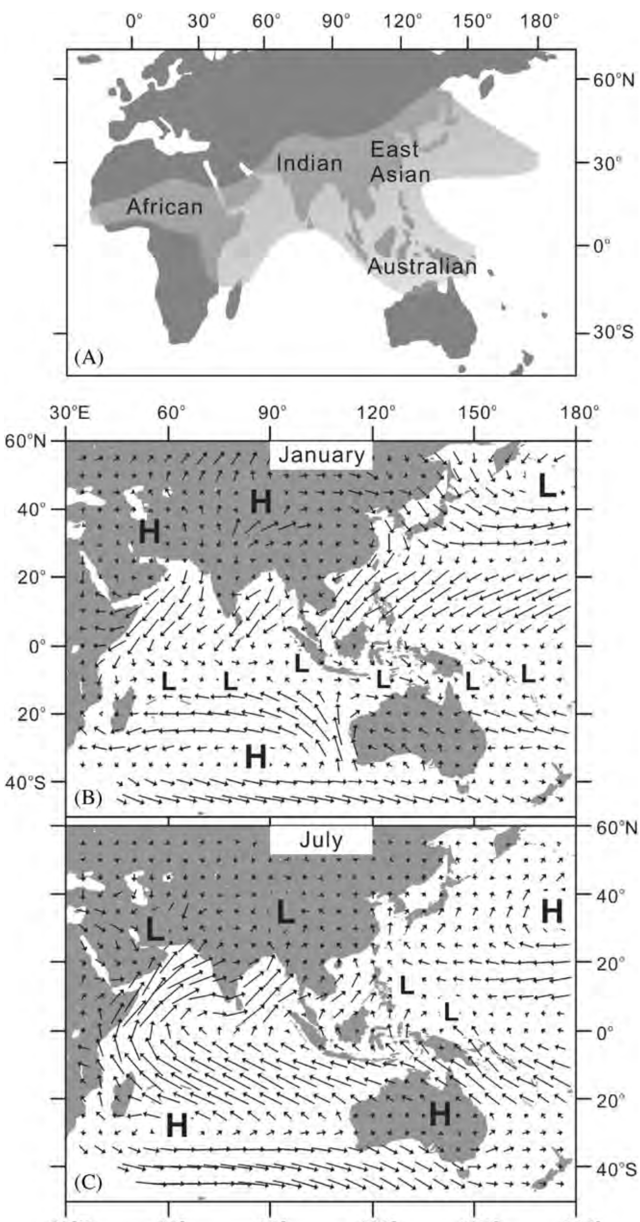

Fig. 1. Modern Asian monsoon system: (A) distribution of modern monssonal regions in Asia, Africa and Australia (modified from Black, 2002); (B) pressure and surface wind pattern in winter and (C) in summer (redrawn from P. Wang et al., 2000).

During the last three decades, scientific interest in the Asian monsoon system has increased significantly, resulting in a series of large-scale international field experiments designed to improve our understanding of the modern monsoon systems, including their variability and predictibility over a variety of time scales. Such experiments include the INDEX (Indian Monsoon Expedition, 1975–76) and the MONEX (International Winter/Summer Monsoon Experiment, 1978–79) in the Indian Ocean (Krishnamurti, 1985), the AMEX (Australian Monsoon Experiment, 1986–87) for the Australian monsoon, and the South China Sea Monsoon Experiment (SCSMEX, 1998–2002) in the SCS (Lau et al., 2000; Ding and Liu, 2001). In addition, long-term observations and process studies such as the Tropical Ocean-Global Atmosphere (TOGA, 1985–1995) and Global Ocean–Atmosphere–Land system (GOALS, 1995–) programs have resulted in significant advancements in understanding basic processes of the Asian monsoon system (Webster et al., 1998). Several international Arabian Sea expeditions between 1992 and 1997 have also contributed tremendously to oceanography and climatology of the monsoon-influenced regions. Such expeditions included cruises organized by the US (e.g. Smith et al., 1998; Smith., 2001), the Netherlands (e.g., van Hinte et al., 1995; Van Weering et al., 1997), the UK (ARABESQUE, Burkill, 1999) , Germany (e.g., Pfannkuch and Lochte, 2000) and others.

Paleomonsoon studies initiated in the late 1970s and early 1980s with African lake level projects (e.g., Street and Grove, 1979) and upwelling records from the Arabian Sea (e.g., Hutson and Prell, 1980; Prell et al., 1980). East Asian monsoon studies commenced later with studies of the Loess Plateau in China (e.g., An et al., 1990; Ding et al., 1992; Porter et al., 1992), followed by studies of sediment archives in the SCS (e.g., Wang and Wang, 1990; L. Wang et al., 1999a). Within the last two decades, a number of drilling and coring cruises have been devoted to paleomonsoon studies. These include ODP Leg 117 to the Arabian Sea and Oman Margin in 1987 (Prell, Niitsuma et al., 1989, 1991), the SONNE 95 cruise "Monitor Monsoon" to the SCS in 1994 (Sarnthein et al., 1994), the SONNE 90 cruise "PAKOMIN (Pakistan Oxygen Minimum)" to the northern Indian Ocean in 1993 (von Rad et al., 1995), and ODP Leg 184 to the SCS (P. Wang et al., 2000). In addition, paleomonsoon studies have advanced through continuing studies of the Chinese Loess Plateau, as reviewed by a number of authors (e.g., Liu and Ding, 1998; An, 2000; Porter, 2001).

Simultaneously, general circulation model (GCM) studies of monsoon climates have progressed significantly and contributed greatly to our understanding of how the monsoon system responds to various large-scale forcings, including tectonics, CO_2, and orbital variations in incident solar radiation (e.g., Kutzbach et al., 1993; Wright et al., 1993). The role of more regional internal feedbacks have also been investigated, including those associated with clouds (e.g. Liao et al., 1994), soil moisture (e.g. Kutzbach and Guetter, 1986), surface albedo (e.g. Bonfils et al., 2001), as well as vegetation and lakes (e.g. Kutzbach et al., 1996). Most of the early studies have used atmospheric general circulation models (AGCMs) although, in the last decade, a large number of coupled models have been successfully developed which integrate atmosphere, ocean, and land surface processes (Claussen and Gayler, 1997; Kutzbach and Liu, 1997; Texier et al., 1997; Hewitt and Mitchell, 1998; Braconnot et al., 1999, 2000; Otto-Bliesner, 1999; de Noblet-Ducoudré et al., 2000; Doherty et al., 2000). Several groups have also fostered intensive model data comparisons (e.g., COHMAP, 1988; Joussaume and Taylor, 1995; PMIP, 2000). Most of these comparisons considered the atmospheric part of the system and model results over land. However, now that ocean–atmosphere GCMs (OAGCMs) are being applied to paleoclimate studies, integration of model results and marine proxy records will advance. Coupled model simulations now offer the possibility of investigating how interannual to multidecadal variability has evolved in the past, in particular the possible role of tropical variability in climate fluctuations through changes in ENSO and related teleconnections (Cane et al., 1997; Clement et al., 1999).

Given these significant developments and advances, Working Group 113 was jointly established in 1998 by the Scientific Committee on Ocean Research (SCOR) and International Marine Past Global Changes Study (IMAGES) project to assess the current status and outstanding issues in this dynamic area of paleoclimate study. The *S*COR-IMAGES *E*volution of *A*sian *MONS*oon (SEAMONS) Working Group conducted three workshops and one symposium: the planning workshop in Shanghai, 1998; the workshop on seasonal and interannual monsoon dynamics in Amsterdam, 1999; the symposium on "Asian monsoons and global linkages on Milankovitch and sub-Milankovitch timescales" in Beijing, 2001 (Clemens et al., 2003); and the final workshop for synthesis and monsoon tectonic evolution in Aix-en-Provence, France, 2002. As the final report of the SEAMONS Working Group, this paper will discuss the current status of monsoon proxies, the temporal variability of the Asian monsoon system from decadal to tectonic time scales, and put forward recommendations for future research efforts.

2. Proxies

Assessing paleoclimate variability using the geological record necessarily depends on the development and reliability of climate proxies. A summary of the more prevalent geological archives and their proxy data which record past changes in the Asian monsoon is presented in Table 1. This proxy overview also provides an indication of some of the potential deficiencies in interpretation of individual proxies. In general, monsoon proxies can be divided into two groups according to the primary aspects of the monsoon that they address: proxies related to monsoon winds (direction, strength and persistence), and those associated with monsoon-induced precipitation.

One of the best studied monsoon proxies is census counts of the planktonic foraminifera *Globigerina bulloides*. The high abundance of this subpolar species in the low-latitude coastal ocean is a clear indicator of monsoon-driven upwelling as evidenced by sediment trap time series and plankton-tow data (Curry et al., 1992), as well as by geographic distribution in core-top sediments (Prell, 1984a, b). As a result, this micropaleontological proxy has been widely used in upwelling regions of the Arabian Sea for paleomonsoon studies ranging from tectonic long-term evolution (Kroon et al., 1991) to high-resolution millennial scale variability (Anderson et al., 2002). Modern observations show that different phases of upwelling in the Indian Ocean can be monitored using different species, *Neogloboquadrina dutertrei* for the initial phase and *G. bulloides* for the final phase of upwelling (Kroon and Ganssen, 1989). By contrast, *G. falconensis* is indicative of non-upwelling conditions (Peeters and Brummer, 2002; H. Schulz et al., 2002a). *G. bulloides* is not as dominant in the significantly weaker upwelling regions of the South and East China Seas. Here, *N. dutertrei* has been applied as an upwelling indicator (Jian et al., 2001).

The integrity of each monsoon proxy depends on the extent to which it responds only to monsoon forcing or, if influenced also by non-monsoon processes, the extent to which this unwanted signal can be identified and removed. Examples include the recent debate on the use of tree-ring δD as a monsoon proxy (Feng et al., 1999, 2002; Zhou, 2002), and the difficulty in differentiating the monsoon signal from the sea level signal when interpreting the deep-sea pollen record (Sun et al., 2003). For these reasons a multi-proxy approach is often employed, an example is the summer monsoon factor for the northern Arabian Sea based on factor analyses of five proxies: lithogenic grain size, Ba accumulation rate, $\delta^{15}N$, abundance of *G. bulloides* and opal mass

Table 1
Synthesis of proxy data indicating changes in Asian monsoon system

Features and processes	Proxies	Archives	Example	Time range	References	Controversial issues and pitfalls
Wind system						
Wind directions (surface and mid-tropospheric wind)	Lithogenic tracers	Hemipelagic sediments	South China Sea Arabian Sea	Neogene Quaternary	Liu et al. (2003)	1. Two-level wind track of Shemal dust source on top of summer monsoon jet
	Clay minerals				Sirocko and Sarnthein (1989)	
	Trace elements				Sirocko et al. (1991, 1993, 2000)	
	Carbonate minerals (dolomite)					
	Specific pollen types and assemblages	Hemipelagic sediments	Arabian Sea, North Pacific, South China Sea	Neogene	Van Campo et al. (1982)	2. Unknown role of sea level change and emerged shelves
					Heusser and Morley (1985)	
					Prell and Van Campo (1986)	
					Sun et al. (1999)	
	Loess-type sediments (>6 μm)		South China Sea	Late Quaternary	L. Wang et al. (1999a)	
	Eolian dust quartz (winter)		North Pacific and western marginal seas	Neogene	Rea et al. (1998)	Signal of grain size distribution <10 μm is pure artifact of lab. analysis
			Sea of Japan	Quaternary	Dersch and Stein (1994)	
					Irino and Tada (2002)	
	$\delta^{18}O$ as SST proxy: wind-driven surface water influx from Red Sea	Corals	Indian Ocean	Quaternary	Tudhope et al. (1996)	
	Downwind grain size gradient	Loess and red clay	Chinese Loess Plateau	Neogene	Ding et al. (1998b)	
	Trace minerals in dust deposits	Lake deposits	Lake Biwa, Japan	Quaternary	Xiao et al. (1995)	
	Wind-induced coastal upwelling:	Hemipelagic sediments	South China Sea,	Neogene	L. Wang et al. (1999a)	Comparatively weak signals in South China Sea
	1. Summer upwelling S.W. of Vietnam and off Somali-Oman	Arabian Sea			Jian et al. (2001)	
	2. Winter upwelling N.W. of Luzon and W. of India				Huang et al. (2002)	
Wind strength	Lithogenic grain sizes (>6 μm)	Hemipelagic sediments	Arabian Sea and central Indian Ocean	Quaternary, perhaps Neogene	Sirocko and Sarnthein (1989)	Quantitative proxy with ±0.4 m/s
	Lithogenic median grain size		Arabian Sea		Sirocko et al. (1991)	1. Two dust tracks in Shemal and monsoon jet below
					Clemens and Prell (1991b)	2.Vegetation in source area
	Silt modal grain sizes (>6 μm)	North Pacific and western marginal seas	Neogene		L. Wang et al. (1999a)	Variations in weathering of source region
	Ti/Al ratio		Arabian Sea	Quaternary	Shimmield et al. (1990)	
					Reichart et al. (1997)	
					Sirocko et al. (2000)	
	Siliciclastic grain sizes	Lake deposits	Lake Biwa, Japan	Quaternary	Xiao et al. (1995)	
	Grain size	Loess and red clay	Loess Plateau, China	Plio-Pleistocene	Ding et al. (2001c)	

	Median (Md)				Lu et al. (1999)	
	$>30\,\mu m$				Porter and An (1995)	
	$>40\,\mu m$				Ding et al. (1992)	
	Ratio, $<2\,\mu m$/ $>10\,\mu m$ %				Vandenberghe et al. (1997)	
	Ratio, 44–16 µm/16–5 µm					
Processes linked with wind strength	Thermocline depth	Hemipelagic sediments	South China Sea	Quaternary	Jian et al. (2000a)	
1. Structure of surface ocean	Planktonic foraminifera and nannofossil indices				L. Wang (2000)	
					Liu et al. (2002)	
	Meridional SST gradient				Wei et al. (2003)	
	Difference in planktonic $\delta^{18}O$					
2. Upwelling productivity	Micropaleontological proxies					
	1. Planktonic foraminifera		Arabian Sea	Neogene	Kroon et al. (1991)	1. Nutrient content of upwelled water
	Abundance of *G. bulloides*				Naidu and Malmgren (1996)	2. $CaCO_3$ and opal dissolution/preservation
					Anderson et al. (2002)	
					Peeters et al. (2002)	
					H. Schulz et al. (2002a)	
	N. dutertrei		South China Sea	Neogene	Jian et al. (2003)	
	Abundance of *G. falconensis* as inverse signal		Arabian Sea	Quaternary	H. Schulz et al. (2002a)	
	2. Benthic foraminifera indicative of high organic-carbon flux		Indian Ocean, South China Sea	Neogene	den Dulk et al. (1998, 2000)	
					Jian et al. (1999)	
	3. Radiolaria					
	Upwelling index		Indian Ocean	Quaternary	Nigrini and Caulet (1992)	
					Grazzini et al. (1995)	
					Vénec-Peyré et al. (1995)	
					Gupta (1999)	
			South China Sea	Quaternary	R. Wang and Abelmann (2002)	
				Neogene	M.H. Chen et al. (2003)	
	4. Dinoflagellate					
	Upwelling species		Indian Ocean	Quaternary	Zonneveld and Brummer (2000)	
	5. Nannofossils		Indian Ocean		Beaufort et al. (1997)	
					Andruleit et al. (2000)	
			South China Sea		Liu et al. (2002)	
	Geochemical proxies					
	1. Organic carbon % and flux		Arabian Sea	Neogene	Clemens et al. (1996)	
					H. Schulz et al. (1998)	
	2. Opal % and flux		Arabian Sea		Clemens et al. (1996)	
			South China Sea		Lin et al. (1999)	
	3. Ba flux		Indian Ocean		Clemens et al. (1991)	
	4. Ba/Al ratio		Arabian Sea,		Shimmield et al. (1990)	
			South China Sea		Wehausen and Brumsack (2002)	
			South China Sea		Lin et al. (1999)	

Table 1 (*continued*)

Features and processes	Proxies	Archives	Example	Time range	References	Controversial issues and pitfalls
	5. Cd/Ca ratio in foraminiferal tests					
	6. $\delta^{15}N$		Arabian Sea, South China Sea		Altabet et al. (1995, 1999)	
			South China Sea		Kienast et al. (2002)	
	7. Pigment/chlorin		South China Sea		Higginson et al. (2003)	
	Aragonite preservation index related to intensity of oxygen-minimum zone		Arabian Sea		Reichart et al. (1997)	
	9. $\delta^{13}C$ in near-surface dwelling planktic foraminifera as inverse nutrient signal		South China Sea		L. Wang et al. (1999a)	
					L. Wang (2000)	
					Jian et al. (2003)	
Monsoon-driven precipitation						
Continental runoff	SSS estimation	Hemipelagic sediments	Indian Ocean	Quaternary	Sarkar et al. (2000)	
	Planktonic $\delta^{18}O$				Kudrass et al. (2001)	
					Doose-Rolinski et al. (2001)	
			South China Sea		L. Wang et al. (1999a, b)	
	Planktonic foraminifera		Bay of Bengal	Quaternary	Cullen (1981)	
	Water column stratification and laminated deposits		Sea of Japan		Oba et al. (1991); Tada et al. (1999)	
	Varve thickness in laminated deposits		Indus Fan	Quaternary	Von Rad et al. (1999a, b, 2002a)	
					Berger and von Rad (2002)	
	$\delta^{18}O$	Corals	Indian Ocean	Quaternary	Charles et al. (1997, 2003)	
			West Pacific marginal seas			
Precipitation rate	$\delta^{18}O$ of lacustrine authigenous carbonates	Lake deposits	West China	Quaternary	Wei and Gasse (1999)	
	Lake level		Australia		Bowler et al. (2001)	
			West China		Zhang et al. (2001)	
	Lake salinity		Tibet and elsewhere		Gasse et al. (1996)	
	Diatoms, ostracoda, $\delta^{18}O$, $\delta^{13}O$				Gasse et al. (1997)	
	$\delta^{13}C$ (C3/C4 plants)	Paleosol carbonate	Pakistan, China, Australia	Neogene	Qin et al. (1999)	CO_2 effect
		Egg shell			Johnson et al. (1999)	
	$\delta^{18}O$ (summer vs. winter rainfall)	Stalagmite	Oman	Quaternary	Neff et al. (2001)	
			China		Fleitmann et al. (2003)	
					Y. Wang et al. (2001)	
					Yuan et al. (2004)	
	Mollusks	Loess	China	Quaternary	Rousseau et al. (2000)	

	Pollen (vegetation changes)	Lake deposits and peat	East and South Asia,	Quaternary	Huang et al. (1997)	
			Australia		Phadtare (2000)	
					Kershaw et al. (2002)	
					Maxwell and Liu (2002)	
		Hemipelagic sediments	Maritime continent and off Australia;	Quaternary	Kershaw et al. (2002)	Sea-level changes
					Sun and Li (1999); Sun et al. (2003)	
			South China Sea			
	Charcoal	Hemipelagic sediments	South China Sea	Quaternary	Luo et al. (2001)	
			Banda Sea		van der Kaars et al. (2000)	
			Sulu Sea		Beaufort et al. (2003)	
			E. Indian Ocean		X. Wang et al. (1999)	
	Eolian dust flux	Oceanic sediments	Northwest Pacific	Quaternary	Hovan et al. (1991)	
	Tree ring density, width and isotopes	Tree ring	China	Quaternary	Hughes et al. (1994)	
					Liu et al. (1996)	
	Snowline	Low-latitude mountains	East Asia	Quaternary	Ono and Naruse (1997)	
Weathering and pedogenesis	Clay minerals	Hemipelagic sediments	South China Sea	Quaternary	Trentesaux et al. (in press)	Non-coeval weathering and transport/deposition
			Arabian Sea		Sirocko and Sarnthein (1989)	
					Sirocko et al. (1991)	
	Ti/Al		Arabian Sea	Quaternary	Shimmield et al. (1990)	
	K/Si		South China Sea	Neogene	Sirocko et al. (2000)	
					Wehausen and Brumsack (2002)	
	Magnetic grain size	Indian Ocean, South China Sea	Quaternary		Colin et al. (1998)	
	(ARM/SIRM ratio)				Kissel et al. (2003)	
	Rb/Sr and other element ratios	Loess–paleosol, red clay	China	Neogene	Gallet et al. (1996)	
					J. Chen et al. (1999)	
					Ding et al. (2001b)	
	Chemical index of weathering				Guo et al. (1996)	
	FeD/FeT (ratio of CBD-extractable Fe_2O_3 and total Fe_2O_3)				Guo et al. (2000)	
	Magnetic susceptibility				Kukla et al. (1988)	
					Zhou et al. (1990)	
					Maher and Thompson (1995)	
					Evans and Heller (2001)	

accumulation rate (MAR) (Fig. 2A; Clemens and Prell, 2003). While the multi-proxy approach is clearly useful in that it provides a means of accounting for problems associated with multiple mechanisms influencing individual proxies, it does not replace the need for detailed understanding of the mechanisms influencing individual proxies; the key to reliable paleoclimate reconstruction is a firm understanding of the various processes contributing to variance in any specific climate proxy. One means of accomplishing this is through long-term sediment trap studies. Table 2 lists sediment trap efforts within the Asian monsoon region for which data are available from the literature (for locations see Fig. 3). Careful analyses and applications of the sediment trap data will significantly improve our knowledge of monsoon proxies. However, one must keep in mind that sediment trap results do not include the many processes, which take place during and after deposition which can strongly impact the integrity of a climate proxy signal. In addition, the majority of trap deployments are of short duration (2 years or less) and limited to one or two trap locations. Long-duration efforts are

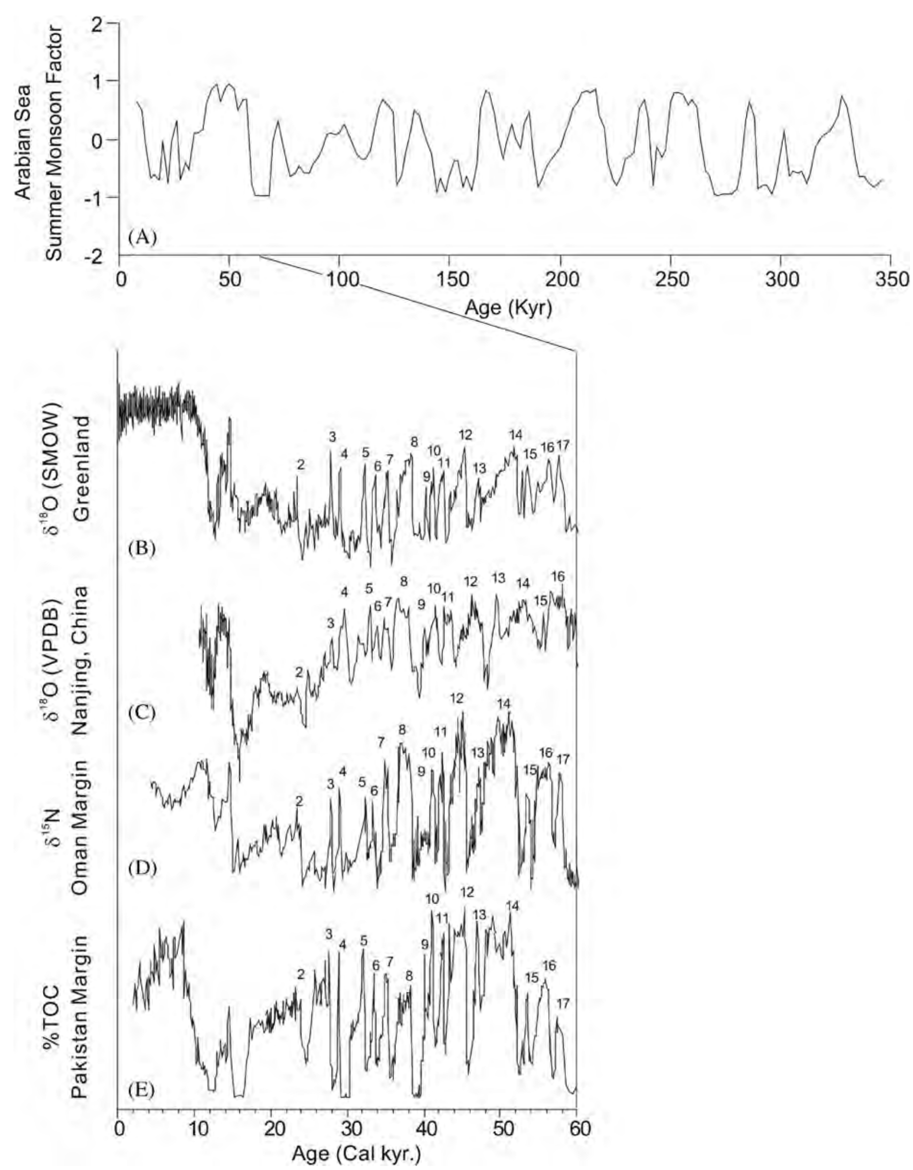

Fig. 2. Orbital and millennial-scale climate records. (A) Arabian Sea summer monsoon factor, a proxy for summer monsoon strength based on factor analysis of five independent records related to summer monsoon circulation (Clemens and Prell, 2003). (B) Greenland Ice Sheet Project 2 (GISP2) $\delta^{18}O$ record, a proxy for temperature change over Greenland (Dansgaard et al., 1993). (C) Hulu Cave $\delta^{18}O$ (near Nanjing, China), a proxy for Asian summer-monsoon strength as it impacts the oxygen isotopic composition of stalagmite calcite (Y. Wang et al., 2001). (D) $\delta^{15}N$, a proxy for the strength of the Indian summer-monsoon as it impacts the productivity-driven denitrification of the Oman Margin (Altabet et al., 2002). (E) Total organic carbon, a proxy for the strength of the Indian summer-monsoon as it impacts the strength of the oxygen minimum zone on the Pakistan Margin (Schulz et al., 1998). Within the error of estimated ages, the numbered interstadial (warm) events are coincident; strengthened Asian summer monsoons during interstadial events in the Northern Atlantic region. All records are plotted on their own published chronologies.

Table 2
Sediment traps deployed in the Asian monsoon region (see Fig. 1 for locations)

Sea	No.	Site	Latitude	Longitude	Water depth (m)	Trap depths (m)	Deploy. period[a]	References
Arabian Sea	T1	WAST	16°23′N	60°32′E	4020	3024	05.1986–10.1995	Nair et al. (1989); Haake et al. (1993); Rixen et al. (2000)
	T2	CAST	14°30′N	64°44′E	3900	2919	05.1986–02.1989	Haake et al. (1993)
	T3	EAST	15°32′N	58°51′E	3770	2787	05.1986–10.1990	
	T4	MS-1	17°41′N	68°45′E	1147	808, 998	11.1994–12.1995	Honjo et al. (1999)
	T5	MS-2	17°24′N	58°48′E	3642	828, 903, 1974, 3141		
	T6	MS-3	17°12′N	59°36′E	3465	764, 858, 1857, 2871		
	T7	MS-4	15°20′N	61°30′E	3974	821, 2229, 3478		
	T8	MS-5	15°38′N	68°34′E	4411	800, 2363, 3915		
	T9	MST-8	10°45′N	51°57′E	1533	1265	06.1992–02.1993	Koning et al. (1997); Broerse et al. (2000); Conan and Brummer (2000); Zonneveld and Brummer (2000)
	T10	MST-9	10°43′N	53°34′E	4047	1032		
	T27	EPT	24°46′N	65°47′E	1099	500	10.1993–2.1994 5.1995–2.1996	Andruleit et al. (2000)
	T28	WPT	24°36′N	65°35′E	2004	500, 1500	10.1993–2.1994	H. Schulz et al. (2002b); von Rad et al. (2002a)
Bay of Bengal	T11	NBBT-N	17°27′N	89°36′E	2263	967, 1498, 2029	10.1987–11.1989	Unger et al. (2003);
	T12	NBBT-S	15°14′N	89°10′E	2738	1131, 1717	01.1990–12.1992	Guptha et al. (1997)
	T13	CBBT	13°09′N	84°22′E	3259	906, 2282	10.1987–10.1991, 11.1993–08.1996	
	T14	CBBT-06	11°02′N	84°26′E	3462	1588, 2527	02.1992–12.1992	
	T15	SBBT	05°24′N	86°46′E	3950	886, 2968	07.1995–12.1996	
West Pacific	T16	KNOT	43°58′N	155°03′E	5370m	2857m	12.1997–12.1998	Kuroyanagi et al. (2002); Honda et al. (2002)
	T17	M1	04°03′N	135°00′E	4762	970, 2940	01.1999–11.1999	Gupta and Kawahata (2002)
	T18	M3	00°08′N	145°02′E	3680	1020, 2060	05.1999–11.1999	
	T19	N1	02°59′N	135°01′E	4413	1589, 3902	04.1991–04.1992	Kawahata et al. (2000, 2002)
	T20	N2	04°07′N	136°16′E	4888	1769, 4574		
East China Sea	T21	T12	24°53′N	122°12′E		569, 768, 967	3–8.1996 10.1996–4.1997	Chung et al. (2003)
	T22	T13	24°55′N	122°19′E		946, 1144, 1343		
	T23	SST-1	29°23′N	128°15′E	1070	600, 800, 1020	3.1993–2.1994	Yamada et al. (1994); Yamada and Aono (2003); Oguri et al. (2003); Tanaka (2003); Katayama and Watanabe (2003)
	T24	SST-2	28°08′N	127°04′E	1090	600, 800, 1040		
South China Sea	T25	NSCS	18°28′N	116°01′E	3750	1000, 3350	9.1987–10.1988	J.F. Chen et al. (1998, 1999)
	T26	CSCS	14°36′N	115°06′E	4350	1200, 2240, 3770	12.1990–4.1995	Wiesner et al. (1994, 1996)

[a]Deployment period only shows the total range, interrupted intervals less than a half year are not indicated.

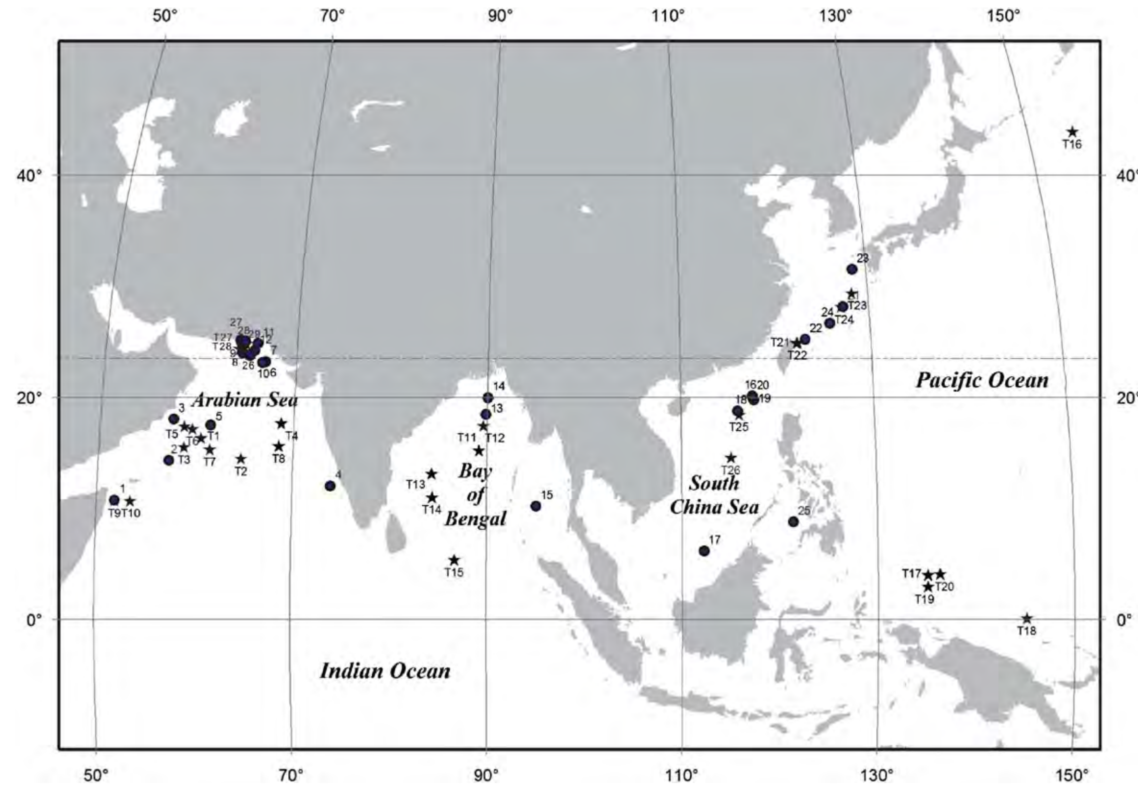

Fig. 3. Location map of the Asian monsoon region showing sites of sediment traps (see Table 2) and sites of deep-sea sediment sequence with a time resolution equal to or better than 500 years (see Table 3, partly from Voelker, 2002).

required to record the entire range of climate variability associated, for example, with El Niño. Short duration efforts of a year or two cannot record the entire range of variability necessary to evaluate proxy response under the full range of modern climate conditions. Multiple traps are required to record gradients, which are often prevalent and important in regions of large variability such as the monsoon influenced regions. The JGOFS Arabian Sea effort was exceptional in that it deployed 5 traps beneath and outside the low-level Somali jet in an effort to record the spatial signal associated with the coastal and divergent upwelling cells driven by this southwest monsoon jet (Honjo et al., 1999). The effort was rewarded with the finding that adjacent traps recorded significantly different signals over the span of weeks, the variability being associated with a strong coastal-derived filament influencing adjacent traps differently. Unfortunately, this was only a year-long deployment. Much more can be learned with long-term deployments such as WAST in the Arabian Sea (Table 2).

High-quality core-top data sets are another way to establish and evaluate monsoon proxies. An example is the distribution of planktonic foraminiferal species in the northern Indian Ocean based on analyses of 251 core-top samples (Cullen and Prell, 1984). This survey not only illustrated the link between % *G. bulloides* and monsoon-driven upwelling, but also brought to light the alteration of planktonic foraminiferal assemblages by carbonate dissolution. Similar work has been done for planktonic foraminifers (Pflaumann and Jian, 1999) and pollen (Sun et al., 1999) in the SCS, but the number of sites is relatively low. In general, there are too few core-top data sets published for the Asian monsoon region, and continued work on systematic core-top analyses to further develop monsoon proxies should be a high priority.

In the following sections we provide overviews of monsoon variability based on marine sediment records from the Asian monsoon regions. These overviews are presented according to time scales of variability.

3. Sub-orbital variability

3.1. New high-resolution sediment archives show new range of short-term periodicities

Paleomonsoon research initiated with efforts to understand links between changes in monsoon intensity and large-scale boundary conditions such as orbital forcing and changes in global ice-volume. These remain active areas of research. However, ice cores from Greenland have since revealed records of climate change on far shorter time scales as well (Dansgaard et al., 1993; Grootes and Stuiver, 1997; Johnsen et al., 2001) (Figs. 2 and 4). These high-resolution records revealed modes of pronounced climate variability, with abrupt amplitude shifts that occur on time scales from millennia to decades and shorter (M. Schulz et al., 1999; Kudrass et al., 2001; M. Schulz and Paul, 2002).

Subsequent to these findings, a number of marine sediment archives have been recovered that are varved and thus provide a paleomonsoon record with annual resolution. Examples include those along the continental margins of Pakistan and western India (von Rad et al., 1995, 1999a, b; H. Schulz et al., 1998; Doose-Rolinski et al., 2001; Lueckge et al., 2001; Berger and von Rad, 2002). Moreover, cores from coral heads have produced short records of monsoon variability with monthly resolution (Quinn et al., 1996). Elsewhere, marine archives have been cored that have sedimentation rates sufficient for resolving long-term monsoon history at least on decades and shorter time spans, such as reported from the northern and southern margins of the SCS (L. Wang et al., 1999a; Buehring, 2001; Higginson et al., 2003). Published paleoceanographic sites in the Asian monsoon region with time resolution of 500 years or better are listed in Table 3 (for location see Fig. 3).

These new marine records can now be compared with a set of recently established ultrahigh-resolution companion records from terrestrial archives, as obtained from East African and Tibetan lake sediments (Gasse

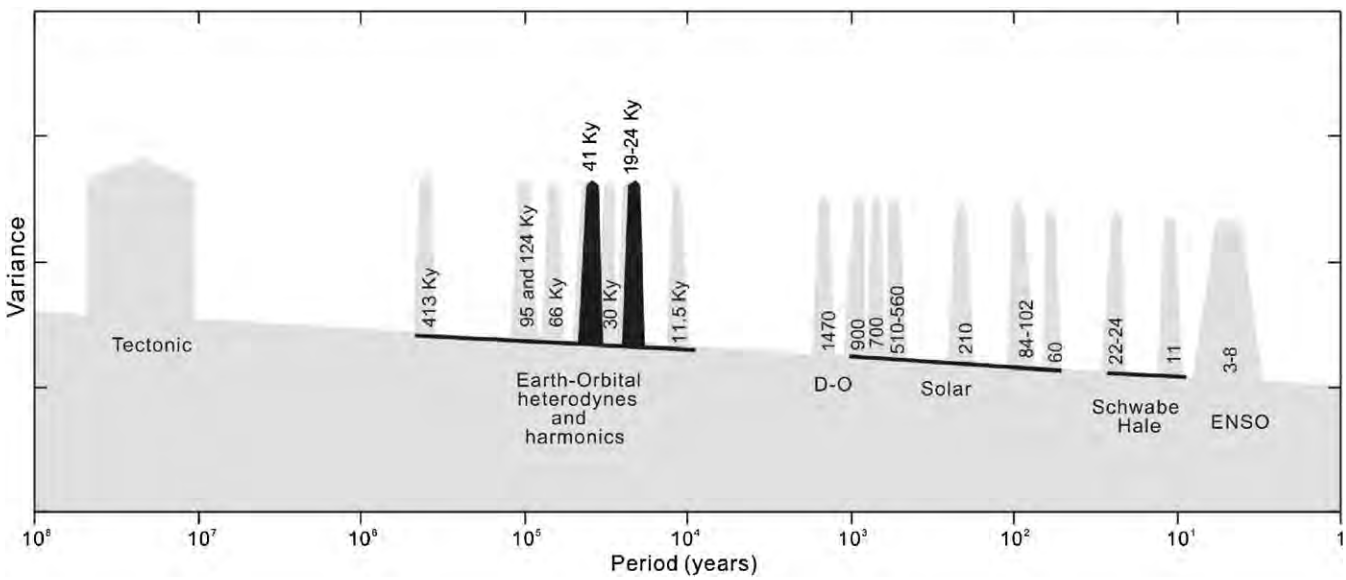

Fig. 4. Conceptual spectrum of monsoon variability on the annual to tectonic time scale. The periods of individual spectral peaks are labeled. Relative concentrations of variance at these periods are unknown. The two black peaks at the 41- and 23-ky periods indicate the Earth-orbital periods, which account for nearly all variability in incoming solar radiation. References address Indian, Asian, and Australian monsoon variability at these periods based on analysis of the marine geological record: *3–8 years (ENSO)*: Buehring (2001), Klein et al. (1997), Peng et al. (2003), Tudhope et al. (1996). *11 years, 22–24 years*: Buehring (2001). *60 years*: Agnihotri et al. (2002). *84–102 years*: Agnihotri et al. (2002), Buehring (2001), von Rad et al. (1999b), L. Wang et al. (1999a, b). *210–250 years*: Agnihotri et al. (2002), Buehring (2001), von Rad et al. (1999b), Yu et al. (2004). *510–560 years*: Buehring (2001), Buehring et al. (in press). *700 years*: Sarkar et al. (2000). *General Decadal to Centennial*: Jung et al. (2001, 2002a, b.), Klein et al. (1997). *900 years:* Buehring (2001), Buehring et al. (in press). *1470 years (or GISP-like)*: Buehring et al., in press), M. Chen et al. (1999), Colin et al. (1998), de Garidel-Thoron et al. (2001), Jian et al. (2000b), Kudrass et al. (2001), Leuschner and Sirocko (2000), L. Wang et al. (1999a, b), Berger and von Rad (2002). *4k (multiples of 1470 years)*: Buehring et al. (in press). *11.5 ky*: Beaufort et al. (2003). *19–23 ky*: Almogi-Labin et al. (2000), Altabet et al. (1995, 1999), Anderson and Prell (1993), Bassinot et al. (1994), Beaufort (1996), Beaufort et al. (1997, 1999, 2001), Bloemendal and de Menocal (1989), Budziak et al. (2000), Buehring (2001), Buehring et al. (in press), Clemens and Prell (1990, 1991b, 2003), Clemens et al. (1991, 1996), Colin et al. (1998), de Garidel-Thoron et al. (2001), Kershaw et al. (2003), Morley and Heusser (1997), Murray and Prell (1991, 1992), Prell (1984a, b), Prell and Van Campo (1986), Prell et al. (1992), Reichart et al. (1997, 1998), Shimmield (1992), Shimmield et al. (1990), Weedon and Shimmield (1991). *~30 ky:* Beaufort et al. (2001), Budziak et al. (2000), Clemens and Prell (1991a, b), Jian et al. (2001), Kershaw et al. (2003), Prell et al. (1992). *41 ky:* Almogi-Labin et al. (2000), Altabet et al. (1995, 1999), Beaufort et al. (2001), Bloemendal and de Menocal (1989), Budziak et al. (2000), Buehring et al. (in press), Clemens and Prell (1990, 1991b, 2003), de Garidel-Thoron et al. (2001), Murray and Prell (1991, 1992), Prell (1984a, b), Prell and Van Campo (1986), Jian et al. (2001), Kershaw et al. (2003), Reichart et al. (1997, 1998), Shimmield (1992), Shimmield et al. (1990), Weedon and Shimmield (1991), L. Wang et al. (1999b). *~66 ky*: Budziak et al. (2000), Clemens and Prell (1991a, b), Jian et al. (2001), Prell et al. (1992). *100 ky*: Almogi-Labin et al. (2000), Altabet et al. (1995, 1999), Anderson and Prell (1993), Bloemendal and de Menocal (1989), Beaufort et al. (1997, 1999), Budziak et al. (2000), Buehring (2001), Buehring et al. (in press), Clemens and Prell (1990, 2003), Clemens et al. (1991, 1996), de Garidel-Thoron et al. (2001), Jian et al. (2001), Kershaw et al. (2003), Morley and Heusser (1997), Murray and Prell (1991, 1992), Prell et al. (1992), Shimmield (1992), Shimmield et al. (1990), Weedon and Shimmield (1991). *413 ky*: Beaufort et al. (1997, 1999). *Tectonic:* An et al. (2001), Kutzbach et al. (1989), Molnar et al. (1993), Prell and Kutzbach (1997), Ruddiman et al. (1989. In addition, Berger and von Rad (2002) found the following periods: 12.4, 14–15, 17, 18.6, 25–26, 29–31, 39, 44, 51–54, 56, 95, 125 and 250 years.

Table 3
Paleoceanographic sites of deep-sea sediment sequence with a time resolution of ⩽500 year in the Asian monsoon region

No.	Location	Core	Latitude	Longitude	Water depth (m)	Age (ka)	Resolution (yr)	References
1	Arabian Sea	905	10°46′N	51°57′E	1586	10-Jul	20	Jung et al. (2002a, b)
2		74KL	14°19′N	57°21′E	3212	0-24	300	Sirocko et al. (1993, 2000)
3		ODP723A	18°03′N	57°37′E	808	0–19	250	Naidu and Malmgren (1996)
4		3268G5	~12°	~74°E	600	0–10	250	Sarkar et al. (2000)
5		SO42-70KL	17°30′N	61°30′E	3810	0–46	~400	Leuschner and Sirocko (2000)
6		SO90-111KL	23°10′N	66°48′E	775	0–62	100	Schulz et al. (1998)
7		SO90-136KL	23°12′N	66°50′E	568	0–65	100	
8		SO90-93KL	23°59′N	64°22′E	1802	0–110	250	
9		SO90-88KL	23°59′N	64°22′E	1783	0–110	250	
10		SO90-137KL	23°07.3′N	66°29.8′E	573	0–31	300	von Rad et al. (1999a, b)
11		SO90-56KA	24°50′N	65°55′N	695	0–5	1	Berger and von Rad (2002)
12		NIOP478	24°12.7′N	65°39.7′E	565	0–225	500	Reichart et al. (1998)
13	Bay of Bengal	MD77-180	18°28′N	89°51′E	1986	0–280	200	Colin et al. (1998)
14		SO93-126KL	19°58′N	90°02′E	3422	0–80	50–200	Kudrass et al. (2001)
15	Andaman Sea	MD77-169	10°13′N	95°03′E	2360	0–160	200	Colin et al. (1998)
16	South China Sea	17940-2	20°07′N	117°23′E	1727	0–41	30–100	L. Wang et al. (1999a, b); Pflaumann and Jian (1999)
17		17964-2/3	06°10′N	112°13′E	1556	0–33	~400	Jian et al. (1999);
18		SO50-31KL	18°45′N	115°52′E	3360	0–110	~500	Huang et al. (1997); Chen and Huang (1998)
19		SO17938-2	19°47′N	117°32′E	2840	0–30	300	M.T. Chen et al. (1999)
20		ODP 1144	22°03′N	117°25′E	2037	0–600	<100	Bühring (2001); Bühring et al. (in press)
21	East China Sea	DGKS9603	28°09′N	127°16′E	1100	0–45	~200	Li et al. (2001)
22		255	25°12′N	123°07′E	1575	0–20	185	Jian et al. (2002)
23		B-3GC	31°29′N	128°31′E	555	0–10	130	
24		170	26°38′N	125°48′E	1470	0–20	250	Li et al. (1997)
25	Sulu Sea	ODP 769A	8°47′N	121°22′E	3644	0–150	~400	Linsley (1996)
26	SO130-289KL	23°07′N	23°07′N	66°30′E	571	0–76	100	von Rad et al. (2002b)
27	SO130-261KL	24°46′N	24°46′N	65°49′E	873	0–35	30	von Rad et al. (2003)
28	SO90-33KL	25°54′N	23°07′N	66°30′E	571	0–76	100	von Rad et al. (2002a)
29	SO90-63KA	24°37′N	24°46′N	65°49′E	873	0–35	30	Staubwasser et al. (2002)

et al., 1996), Tibetan ice cores (Thompson et al., 2000), Chinese and Oman speleothems (Y. Wang et al., 2001; Fleitmann et al., 2003; Yuan et al., 2004), and from Tibetan tree rings (see Bao et al., 2003, for references). Among the millennial-scale climate records generated to date, those from monsoonal regions show extremely strong similarities to those from the Greenland ice cores, indicating the coupled nature of high- and low-latitude abrupt climate change (Figs. 2B–E).

A number of conceptual models are under development in efforts to constrain the forcing mechanisms underlying short-term changes in monsoon climate. Most hypothesize that the short-term changes in monsoon intensity are linked to internal oscillations in thermohaline circulation and as well as atmospheric energy and moisture transfer. In some cases, a link to millennial- and centennial-scale variations of solar activity has been made on the basis of variations in ^{10}Be flux and ^{14}C production (Bond et al., 2001; Sarnthein et al., 2002; M. Schulz and Paul, 2002). Exploration of the tidal influences on monsoon variability, like the solar cycles, has also been discussed (e.g. Fairbridge, 1986); and Berger and von Rad (2002) evoked lunar tidal action. The extent to which sub-orbital scale variability is driven externally by short-term changes on solar radiation or internally by feedbacks among climate system components is crucial to understanding monsoon climate change.

Further progress in understanding sub-orbital variability in monsoon climate relies on addressing a range of specific issues within specific frequency bands as established in records from the primary monsoon subsystems. Analysis of these records should address the following objectives: (1) evaluate the different levels of overlapping paleomonsoon periodicities [e.g., Dansgaard–Oeschger (DO) cycles, solar cycles, lunar tidal cycles and ENSO cycles] with regard to changes in both moisture transport and wind strength and direction; (2) identify and evaluate systematic differences in the short-period monsoon spectrum between cold and warm regimes; and (3) evaluate the phase relationships of climate signals in the monsoon region as compared to high-latitude climate components (although this approach still remains at the edge of dating accuracy). This information will enhance our understanding of the role monsoon circulation plays in abrupt climate change as a potential driving mechanism, as an amplification mechanism, and/or as a response to external high- or low-latitude processes.

3.2. Dansgaard– Oeschger cycles during the last glaciation

Monsoon variations on DO time scales (~1500 years) were reported from the varved sediment sections off Pakistan (H. Schulz et al., 1998), from bioturbated high-resolution sediment sections in the Arabian Sea (Sirocko et al., 1993, 1996; Altabet et al., 1995, 1999), Bengal Fan (Kudrass et al., 2001), the SCS (L. Wang et al., 1999a; Buehring et al., in press; Higginson et al., 2003), the Sulu Sea (de Garidel-Thoron et al., 2001; Oppo et al., 2001), and the Sea of Japan (Tada et al., 1999). As an overriding feature these records reveal increased summer monsoons during interstadials, and increased winter monsoons during stadials. These marine-based results are consistent with the exceptionally well-dated monsoon records of Hulu cave (Y. Wang et al., 2001) as well as results from loess sections in China (Porter and An, 1995; An, 2000). However, the age models associated with these various archives are insufficient to assess the fine details of phasing among the monsoon records themselves nor the details of monsoon phasing relative to high-latitude climate change at the level necessary to determine which system initiates these abrupt transitions. This is especially relevant during MIS 3, when DO cycles were most prominent; marine ^{14}C reservoir ages and the incremental time scale of the Greenland ice cores are not known well enough to resolve potential differences in timing which would allow inferences about the mechanism of global energy transfer at this time scale (Sarnthein et al., 2001, 2002). Likewise, the accuracy of U/Th dates is insufficient to resolve ages with a precision better than 1% within MIS 3.

DO cycles are not restricted to MIS 3. Monsoon proxies from Arabian Sea sediments show, for example, that the transition from interglacial MIS 5 to glacial MIS4 occurred in discrete steps, characterized by a number of warming rebounds. This record, chronologically constrained by the well-dated Toba ash, reflects strong millennial-scale monsoonal fluctuations (H. Schulz et al., 2002b).

3.3. Millennial-to-decadal-scale cycles in the Holocene

During the Holocene DO-scale variability evident during the last glaciation is no longer dominant (M. Schulz et al., 1999). Instead, the Holocene interval is characterized by variability with a broad range of periodicities near 890–950, 550, 200, 145, 80–105, 20–24 and 11 years (Fig. 4). These scales of variability may be linked to changes in solar activity, which produce very weak variations in incident solar energy (Haigh, 1996; Beer et al., 2000; Labitzke, 2001). Since monsoon variance at this time scale appears non-stationary (Buehring, 2001), the most commonly used techniques of time series analyses cannot be used for detecting global phase relationships, even in cases where the age control is based on annual-layer counting.

Both laminated Holocene sediment sections off Pakistan (Berger and von Rad, 2002) and coral rings from the equatorial Indian Ocean (Charles et al., 1997) provide high-resolution monsoon records. Most remarkable are significant, but non-stationary periodicities in the range of 3–6 years, periodicities that likely represent a d response to ENSO dynamics and its moisture contribution to the Southeast Asian monsoon system (Beaufort et al., 2001, 2003). Similar periodicities were recently found (and reproduced in triplicate) in non-laminated high-resolution records from the northern SCS by means of color records with a sampling resolution of 0.3 mm per color pixel (Buehring, 2001); this resolution, however, implies a number of questions about an unexpectedly low influence of bioturbational mixing.

3.4. Modern interannual to decadal variability

Our interpretation of monsoon variation at millennial to decadal time scales begins with our knowledge of the mechanisms responsible for present day interannual to decadal variability. We therefore summarize here some of the factors and teleconnections identified in the modern monsoon system that can serve to guide interpretation of proxy data. The monsoon has been shown to vary at different time scales from intraseasonal, with monsoon active and break phases, to interannual and multidecadal. Similarities have been found between the pattern of interannual variability and intraseasonal variability (Webster et al., 1998), which suggests that the monsoon oscillates between several basic states under the influence of remote forcing induced by SST or land surface conditions.

The snow cover over Eurasia can strongly affect the Asian monsoon (Shukla and Mintz, 1982). In particular, abundant Eurasian winter snow cover tends to be followed by an anomalously southern position of the April 500 mb ridge over India, which appears as a precursor of deficient summer monsoon rain (Bhanukumar, 1989). This relationship is evident in model studies (Zwiers, 1993) that show the link between the energy used to melt excessive snow cover, surface temperature and sensible heat over the Tibetan Plateau (Vernekar et al., 1995). Snow cover and soil moisture thus appear to be important factors at this time scale, since the sensible heat flux in the elevated heat source over Tibet is important in regulating the strength of the summer monsoon (Yanai and Li, 1994).

The relationship between the monsoon and ENSO has also received much attention. Soman and Slingo (1997), for example, have analyzed the sensitivity of the summer monsoon to aspects of SSTs anomalies associated with ENSO. The results suggest that the changes in the Walker circulation associated with SST anomalies in the East Pacific, with implied additional subsidence over the eastern hemisphere, is the dominant mechanism whereby the Asian monsoon is weakened during El-Niño years. A relationship with warmer than normal SSTs in the west Pacific was also established. From paleo-records, Himalayan ice cores provided evidence for a devastating Indian drought from 1790 to 1796, concurrent with the very strong ENSO event of 1790 to 1793 followed by a moderate ENSO event of 1794–1797, suggesting an association between ENSO and failure of the Asian monsoon (Thompson et al., 2000). Some modeling studies investigated the relative roles of internal atmospheric dynamics, land surface evaporation, and SST forcings on the coupling between the Asian monsoon and the Southern Oscillation (Lau and Bua, 1998). They concluded that the long-term memory and the planetary scale patterns of the Asian monsoon–Southern oscillation system are primarily forced by large-scale anomalies of SST. Land–atmosphere interactions have a minor influence on the large-scale, but over East-Asia a strong coupling exists between the local water and energy cycle which strongly modulates the local response (Lau and Bua, 1998). These relationships between ENSO and the Indian monsoon are not constant with time and several indices suggest a weakening over the last decade (Kumar et al., 1999).

Additional attention has also given to the role of the Indian ocean and its influence on monsoon circulation in the surrounding areas (Webster et al., 1998). The Indian ocean dipole appears as an oscillation across the Indian ocean manifest as an anticorrelation between SST anomalies along the equator between the eastern and the western part of the basin. This oscillation is linked to large-scale displacement over the ocean of deep convection according to the location of the warmer waters. Fluctuation of East African and Arabian Sea precipitation are related to this dipole (Saji et al., 1999). More recently Gadgil et al. (2003) demonstrated that the monsoon precipitation over the Indian sector is also affected by fluctuations of the Indian dipole and not only by ENSO variability. Also the mid-latitude processes affect the monsoon and result in changes in the Asian monsoon/ENSO relationship (Charles et al., 2003).

3.5. Summary

Recent studies indicate a strong coupling between monsoon circulation and Northern Hemisphere (NH) climate change. However, our understanding of the monsoon variability on decadal to millennial time scales remains insufficient. It is unclear whether the monsoon responds through teleconnections to high-latitude mechanisms, or plays a primary role in initiating and/or amplifying abrupt changes initiated in the tropics via ENSO-like mechanisms. In either case, the monsoon system does not operate in isolation, and a concerted effort is needed to assess the teleconnections between the various climate regimes and the stability of their links through time (Zahn, 2003).

Resolving these questions requires analysis of long (40–60 m length), undisturbed, continuous, and large-diameter cores from high deposition rate and varved sediments within monsoon-influenced regions. These archives are critical to understanding monsoon variability at sub-orbital time scales and the extent to which tropical monsoons drive extratropical climate change or respond to it.

The dating accuracy of marine sediment archives requires substantial improvements for the last 100 ka. One method which does not require technological advances is the application of lamina counting in varved sequences which will enable correlation of paleomonsoon records with the paleoclimatic records obtained from the Greenland ice cores and other well-dated terrestrial sections.

4. Orbital-scale variability

4.1. Monsoon variability: a response to insolation forcing and internal feedbacks

Paleoclimate analysis at the orbital time scale enjoys a distinct advantage relative to other time scales of investigation in that the forcing function can be accurately calculated for many tens of millions of years into the past (Laskar et al., 1993; Laskar, 1999). In theory, if one knows the system input (insolation) and can reliably measure the system output (climate), then the internal workings of the system can be derived. Critical to understanding climate system physics is the ability to know accurately the amplitude and phase response of climatic subsystems (atmosphere, hydrosphere, cryosphere, biosphere, and lithosphere) relative to the insolation forcing and relative to one another (Imbrie et al., 1992, 1993). Combined, amplitude and phase yield information on the extent to which a given subsystem is forcing (or forced by) another and the extent to which a climate response is amplified or diminished relative to the forcing.

Paleoclimate studies over the past decades have clearly demonstrated that proxy records of climate change (e.g. $\delta^{18}O$) often bear little resemblance to the primary external forcing (e.g. solar insolation). This is clear evidence that the system output we measure using proxy records is the result of a large-scale internal redistribution of insolation energy among climate subsystems. Analysis of Pleistocene monsoon variability has largely concentrated on the relationship between monsoon circulation and two primary internal and external forcing mechanisms, insolation and global ice volume. The overview below shows that monsoon variability is also the product of energy redistribution among climate subsystems; it is not a direct, linear response to changing insolation or global ice volume.

4.2. *Orbital-scale monsoon variability*

Analysis of orbital-scale monsoon variability preserved in marine sediments spans the past 20 years, having been initiated with studies on the Indian monsoon system (Arabian Sea) and recently expanding eastward to the Asian system (SCS) and southward to the Indo-Australian system (Indonesia and Northern Australia). This section provides a chronological overview of these efforts with emphasis on the necessity of a multi-proxy approach and the evaluation of temporal leads and lags among insolation forcing, internal feedback processes, and the monsoon response. The multi-proxy approach allows differentiation of monsoon variance from that driven by processes not necessarily related to monsoon circulation (e.g. dissolution, preservation, sediment reworking). The analysis of phase provides critical information regarding the monsoon response to external insolation forcing and changing internal boundary conditions such as high-latitude ice volume. For example, if changes in monsoon strength take place before changes in ice volume (within a specific frequency band) then clearly monsoon variance is not driven by changes in high-latitude ice volume. In contrast, if changes in monsoon strength take place in phase with, or shortly after, changes in ice-volume, then a link between high-latitude climate change and monsoon circulation is indicated and mechanisms by which this might take place need to be put forth and tested. It is important to note that the phase relationships (and implied physics) observed in one frequency band may be entirely different from those in another band. In light of this, analysis by visual inspection alone is insufficient, the climate spectrum is far too complex, time series analytical techniques are required (Clemens and Prell, 2003). Within this framework, we do not discuss in detail the rich and extensive Loess record from central Asia as chronostratigraphic issues limit land–sea correlations at the level necessary to evaluate phase relationships (Zhou and Shackleton, 1999; Spassova et al., 2001). We view this as a critical impediment in linking the terrestrial and marine records of the monsoon although we note that progress is being made in this regard (Heslop et al., 2000). The reader will also note that the narrative is biased toward the summer-monsoon as revealed by Arabian Sea proxy records. The first proxy records of summer- and winter-monsoon variability of sufficient length for orbital-scale analysis from the East Asian and Indo-Australian systems are currently under development and will add greatly to the body of knowledge in the coming years.

4.2.1. *Indian monsoon*

Prell (1984a, b) initiated orbital-scale studies of the Indian summer monsoon within the context of time series analysis using a *G. bulloides* proxy from the Owen Ridge, northern Arabian Sea. This 164 ky record was sufficient in length to assess variance in the precession band but not in the obliquity band. Within the precession band, summer-monsoon maxima were found to be in phase with $\delta^{18}O$ minima and to lag NH summer insolation maxima (perihelion at June 21) by ~5.5 ka (−90°). The lag was attributed to the influence of terrestrial ice sheets on albedo and sensible heating over Asia (sensible heating maxima during times of ice-volume minima). Prell and Van Campo (1986) presented a 140 ky long pollen record from core MD76-135 which was coherent and in phase with the *G. bulloides* record, linking the marine and terrestrial monsoon response. In an effort to better understand the links among the summer monsoon, insolation, and terrestrial ice volume, Prell and Kutzbach (1987) conducted a series of GCM sensitivity tests which indicated that the monsoon strengthened with increasing NH summer insolation and weakened with increasing glacial boundary conditions. These early findings and experiments resulted in a paradigm whereby monsoon variability came to be described almost wholly within the context of glacial–interglacial variability with a strong emphasis on the precession component. Monsoon variability is now known to be considerably more complex.

Ocean Drilling Program (Leg 117) in the Arabian Sea stimulated the generation and down-core application of several new summer-monsoon proxies applied to the Owen ridge and Oman margin records, including lithogenic grain size and MAR (Clemens and Prell, 1990; Krissek and Clemens, 1992), Ti/Al and excess Ba

MAR (Shimmield et al., 1990), biogenic opal and organic carbon MAR (Murray and Prell, 1991), as well as *G. bulloides* shell flux (Anderson and Prell, 1993). The question then became, which of the proxies were faithful recorders of summer-monsoon variability given the possibility that each was influenced in some manner by non-monsoon processes. Clemens et al. (1991) addressed this issue using a spectral approach to evaluate which proxies shared variance and similar phase responses at the primary orbital periods. A core group of four proxies (lithogenic grain size, opal MAR, excess Ba MAR, and *G. bulloides*) shared variance and the same phase at the precession and obliquity bands over the past 350 ky In the context of NH precession-driven insolation (June 21 perihelion) and ice volume, these analyses indicated a –122° phase lag relative to summer insolation maxima and a −43° phase lag relative to ice minima (Fig. 5). The obliquity band analysis indicated an in-phase relationship with NH summer insolation maxima and a lead of over 70° relative to ice-volume minima (Fig. 5). These phase results could not be interpreted within the context of NH summer insolation and simple glacial–interglacial forcing, especially in light of the fact that summer-monsoon maxima lead ice minima at the obliquity band. These phase relationships were interpreted to indicate that the timing of strong monsoons in the precession band are driven by latent heat transport from the southern subtropical Indian Ocean and by changes in ice-volume as it impacts albedo and sensible heating of the Asian Plateau. The interpretation of obliquity band variance was that NH summer insolation and latent heat export from the southern subtropical Indian Ocean are dominant driving mechanisms (Clemens et al., 1991).

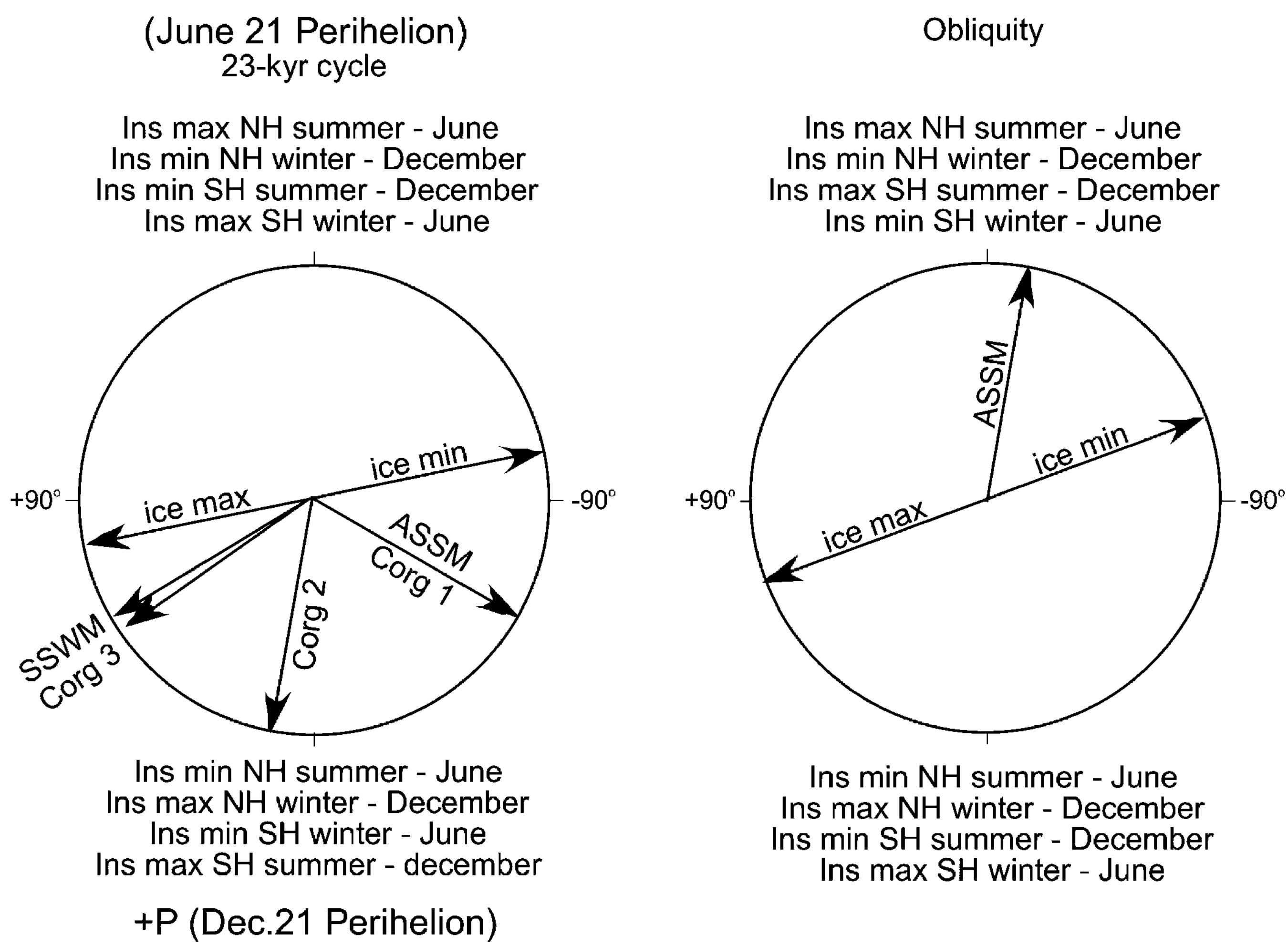

Fig. 5. Phase wheels for precession and obliquity. Phase wheels display coherence and phase relationships between climate proxies and $\delta^{18}O$. Zero phase is set at June 21 perihelion (−P, minimum precession) for the precession wheel and at maximum obliquity; both correspond to maximum NH summer insolation. Vectors represent the timing of proxy maxima relative to the zero point. Positive sign (counterclockwise) represents a lead and negative sign (clockwise) represents a lag. Text at 0° and 180° indicate seasonal insolation relationships associated with each orbital configuration for the SH and NH. Labeled vectors indicate the timing of maximum climate response within each 23,000 year precession and 41,000 year obliquity cycle as discussed in the text. Arabian Sea summer monsoon (ASSM), Sulu Sea Winter Monsoon (SSWM), Organic Carbon Groups 1, 2, and 3 (Corg 1, 2, 3). See text for explanation.

The large precession-based phase lag has since been replicated in organic carbon (Corg), *G. bulloides* and $\delta^{15}N$ from the Murray Ridge, northern Arabian Sea (Reichart et al., 1997; Reichart et al., 1998), in $\delta^{15}N$ from the Owen Ridge and Oman Margin, northern Arabian Sea (Altabet et al., 1995; Altabet et al., 1999) as well as in a pollen record of *Cryptomeria japonica* from the north Pacific off Japan (Morley and Heusser, 1997). The Morley and Heusser work provided the first indication that the East Asian summer monsoon has the same phase response as the Indian summer monsoon. However, their age model is based on *Cycladophora davisiana* and, as such, the phase result needs to be replicated in a core with a $\delta^{18}O$-based age model. While the precession phase lag itself is not debated, its interpretation as a function of latent heat export from the southern subtropical Indian Ocean is debated (Reichart et al., 1998; Clemens and Prell, 2003). Reichart et al. (1998) call upon a prolonged summer monsoon season driven by increased insolation at the end of the summer to account for the phase lag. Anderson and Prell (1993) favor a more direct influence of insolation and ice-volume given that their record of *G. bulloides* shell flux from the Oman margin is dominated by 100- and 23-ka variance, with almost no variance at the obliquity band. Emeis et al. (1995) present a SST record (alkenone unsaturation index) from the Oman margin as a summer-monsoon indicator. Based on visual analysis (no spectral are presented) they also call upon a strong ice-volume forcing (strengthened monsoons during interglacial periods).

Clemens et al. (1991) extended the lithogenic grain size, *G. bulloides*, and opal flux records of summer-monsoon strength to 3.5 Ma. They documented that the phase of strong monsoons has systematically drifted over the past 2.6 Ma. During the initiation and growth of the NH ice sheets, the phase of strong monsoons moves away from the phase of maximum ice-volume, systematically shifting by ~83° and ~124° at the precession and obliquity bands, respectively. While these data clearly indicate non-stationarity in the phase of the monsoon system relative to developing NH ice volume, resolution of the extent to which the monsoon may be non-stationary in phase relative to insolation forcing awaits a means of developing age models that are independent of astronomical tuning.

In addition to the summer-monsoon proxies which share variance and are in phase with one another at the precession and obliquity bands, a number of other proxies were developed but rejected as summer monsoon indicators. The flux of lithogenic material (bulk, mineralogic, elemental) was found to be coherent and in phase with maximum ice volume at all orbital periods and thus interpreted not as a monsoon signal but as an aridity and/or sea-level signal depending on location (Clemens and Prell, 1990; Shimmield et al., 1990; Krissek and Clemens, 1991; Krissek and Clemens, 1992; Shimmield, 1992).

The interpretation of the Corg signal as a winter- and/or summer-monsoon indicator is debated given the potential influence of sedimentation rate on preservation. A number of publications have addressed these issues relative to the monsoon (Murray and Prell, 1991, 1992; Shimmield, 1992; Reichart et al., 1998; van der Weijden et al., 1999; Budziak et al., 2000). Budziak et al. (2000) provide an excellent summary, dividing the Corg records into three phase groups within the precession band (obliquity-band variance is not addressed). Of the five Arabian Sea cores for which Corg has been reported, only the record from the Murray Ridge (Reichart et al., 1998), northern Arabian Sea, is consistent with the –121° precession phase found in *G. bulloides*, LGS, $\delta^{15}N$, Opal MAR, and excess Ba MAR (Group 1). The other three northern Arabian Sea records cluster about a phase of –190° (Group 2) while the Corg record from the southern Arabian Sea (Rostek et al., 1997) has a phase of −235° (Group 3) (Fig. 5). The phase of the Corg response appears to vary depending on location. Murray and Prell (1991, 1992) reject the Group 2 response as a direct monsoon proxy, suggesting that it is driven largely by preservation as a function of sedimentation rate. Shimmield (1992) acknowledges the influence of remineralization, preservation, upwelling productivity, and sediment redistribution on the Corg accumulation record which puts it at odds with other summer-monsoon proxies but at the same time recognizes that it does yield a coherent geographic pattern across the Oman margin. Budziak et al. (2000) observe that the Group 2 (−190°) and 3 (−235°) phase clusters lie between the summer-monsoon indices of Group 1 (−121°) and maximum glacial ice volume (−278°). They interpret these Corg phase groups as the combined influence of summer-monsoon productivity and productivity driven by enhanced winter monsoon winds (maximized during glacial intervals). They suggest that the Corg signal in the southern Arabian Sea (with a phase very near max ice-volume) is dominantly influenced by winter monsoon circulation.

Among the coccolithophore, *Florisphaera profunda* has the unique ecological characteristic of living deep in the photic zone. Although the correlation of its relative abundance with monsoon-driven upwelling is complex (Andruleit et al., 2000; Broerse et al., 2000), it has been used as a reverse proxy of primary production related to changes in wind-stress. When applied to central Indian ocean (Beaufort, 1996) and to southern Arabian Sea (Rostek et al., 1997) it shows a great correlation with Corg. The precession component is well expressed in these records and has a phase of −240° (Group 3) (Beaufort, 1996; Beaufort et al., 1997; Rostek et al., 1997). Beaufort et al. (2001) notice similar phase in other records from the Sulu Sea and the Pacific Ocean and therefore concluded that Group 3 phase may be related to an ENSO influence. Almogi-Labin et al. (2000) use benthic foraminifera and *G. bulloides* as productivity indicators in a core from the Gulf of Aden, northern Arabian Sea. Visual inspection of these 530 ky long records (no phase analyses are reported) leads them to conclude that their proxies are responding to enhanced winter monsoon productivity indicating stronger winter monsoons during glacial intervals.

4.2.2. East Asian monsoon

Studies of East Asian monsoon variability in marine records were initiated with SST records indicating that Last Glacial Maximum seasonal SST contrast in the northern SCS was 3–4 °C higher relative to Holocene values (Wang and Wang, 1990; P. Wang et al., 1995). This enhanced seasonality was interpreted as evidence for intensification of the winter monsoon, a conclusion later confirmed by Huang et al. (1997) and L. Wang et al. (1999a).

L. Wang et al. (1999a) compared the distribution of planktonic $\delta^{13}C$ in the surface sediments of the SCS with upwelling, developing planktonic $\delta^{13}C$ as an indicator of East Asian summer-monsoon strength in the SCS. Foraminifer- and alkenone-based SST records were developed as indicators of winter-monsoon strength. Visual inspection of these 250 ky long records (no spectra or phase results were reported) indicated intensified winter- and weakened summer-monsoons during glacial periods; interglacial periods were interpreted to have weakened winter- and strengthened summer-monsoons.

Jian et al. (2001) developed a multi-proxy set of upwelling indicators from areas of known summer- and winter-monsoon upwelling within the SCS; proxies included benthic foraminifer flux, Corg flux, *Uvigerina peregrina* %, and estimated thermocline depth based on planktonic foraminifera. Spectra of these 220 ky long records indicate that winter-monsoon upwelling varies at 100-, 41-, and 23-ka periods, whereas the summer-monsoon upwelling varies at the precession period but also contains a rich set of heterodynes often associated with the primary orbital periods. Visual inspection of these records (no phase is reported) suggests strengthened summer monsoons during interglacial periods and strengthened winter monsoons during glacial periods. In the southwestern SCS, opal content and radiolarian abundance were used by R. Wang and Abelmann (2002) to indicate upwelling variations over the last million years, again with enhanced summer monsoons during interglacials, in particular since 0.6–0.7 Ma.

De Garidel-Thoron et al. (2001) and Beaufort et al. (2003) developed *F. profunda* in the Sulu Sea as a proxy for winter-monsoon primary productivity. They document that the winter monsoon has strong concentrations of variance at 100- and 41-ka with considerably smaller variance in the 23-ka precession band. They present the first phase estimates of East Asian winter monsoon strength indicating a 20° lead relative to ice maxima at the precession band (Fig. 4) and an 18° lag relative to ice maximum at the 100-ka eccentricity band; these small phase differences are consistent with glacial boundary conditions being a primary factor in driving winter monsoon strength.

Pollen and spore records have been used for paleomonsoon studies in the northwest Pacific as well as the SCS. As discussed above, Morley and Heusser (1997) interpreted *C. japonica* as a proxy for summer-monsoon strength in Japan indicating a similar phase response as is found in the Arabian Sea. Working on pollen from surface sediments and cores from the SCS, Sun et al. (1999) interpret the influx of tree pollen as a proxy for winter-monsoon strength and the percentage of fern spores as a proxy for summer-monsoon strength. The results again indicate strengthened winter monsoons during glacials and enhanced summer monsoons in interglacials.

Changes in grain size of terrigenous sediments, in clay mineral composition, and in element ratios have also been used as East Asian monsoon proxies. In the northern SCS, L. Wang et al. (1999a) used silt median size of siliciclastics to distinguish eolian versus fluvial sediments; the fluctuations in their relative abundance were interpreted to indicate predominance of winter- versus summer-monsoon strength. In general, these proxies indicate enhanced winter monsoons during glacials and enhanced summer monsoons during interglacials. Liu et al. (2003) estimated the relative intensity of the winter and summer monsoons over the past 2 Ma based on smectite/(illite + chlorite) ratio. They found large amplitude fluctuations between 1.2 and 0.4 Ma. Wehausen and Brumsack (2002) employed K/Si ratios from the northern SCS to reconstruct the history of chemical weathering as a function of summer monsoon precipitation; they found clear precession forcing of the summer monsoon over the time interval 3.2–2.5 Ma.

4.2.3. Indonesian– Australian monsoon

Four pollen records from marine cores across the northern tropical region of Australia provide evidence of vegetation change, relative to marine $\delta^{18}O$, over the last 100,000–300,000 years (X. Wang et al., 1999; Moss and Kershaw, 2000; van der Kaars et al., 2000; van der Kaars and De Deckker, 2002). All four cores are in Southern Hemisphere (SH) regions predominantly influenced by the Australian summer monsoon. Time series analysis on major pollen components as well as charcoal, indicate statistically significant variance at Earth-orbital periods. However, the spectral patterns of the various pollen components, while expressing internal consistency between records, show a great deal of variation between components (Kershaw et al., 2003).

Charcoal and Rainforest gymnosperms, indicative of marginal rainforest environments, show a strong 30 ka period and good correspondence with ENSO variability modeled by Clement et al. (1999). Similar frequency results were obtained from a charcoal record in the Sulu Sea (Beaufort et al., 2003). On the assumption that phases of high ENSO variability would promote burning that would in turn impact fire-sensitive gymnosperms, it is considered that these component signatures are primarily a response to ENSO. Rainforest angiosperms, indicative of complex rainforest, produced spectra visually similar to that of the marine $\delta^{18}O$ record suggesting a response to glacial boundary forcing. However, analysis of phase relationships demonstrates a close correspondence with summer-monsoon proxies from the Arabian Sea indicating an interhemispheric connection between the two summer monsoon systems. By contrast, pollen of the dominant sclerophyll vegetation, *Eucalyptus*, displayed a precession-dominated spectrum that was in phase with solar radiation received in the summer hemisphere tropics and is considered to reflect monsoon rainfall generated in southern latitudes without necessarily any connection with NH forcing.

4.3. General circulation modeling

Modeling experiments have been used to evaluate how the monsoon reacts to seasonal changes in insolation, and to study how internal feedback mechanisms (e.g. clouds, vegetation, ocean circulation) act to amplify or dampen the initial solar forcing. Time slices most commonly selected include the early Holocene (11 or 9 ka), the middle Holocene (6 ka) and the Last Interglacial (125 or 126 ka) to study monsoon change under increased summer insolation, the Last Glacial Maximum period (23–19 ka) during which the hydrological cycle is strongly reduced due to the presence of the large ice sheets in the NH and consequent global cooling, and the last glacial inception (115 ka), where, compared to present day, a larger eccentricity leads to reduced summer insolation, colder summers and reduced monsoon activity. One study also emphasizes the possibility of a strong monsoon under a glacial climate 175 ka ago (Masson et al., 2000). Within the Paleoclimate Modeling Intercomparison Projects (Joussaume and Taylor, 1995; PMIP, 2000), several models have performed exactly the same experiments for 6 and 21 ka. This ensemble of simulations provides the best AGCM estimate of these two climates, and has been compared to available data sets over land (Harrison et al., 1998). All these studies strongly emphasize the direct link between increased NH summer insolation and strengthened Indian and East Asian monsoons (Prell and Kutzbach, 1987; Wright et al., 1993; de Noblet et al., 1996; Dong et al., 1996). This stresses the dominant role of the land–sea contrast in the inland penetration of the monsoon flow in Asia (Joussaume et al., 1999). Comparing the characteristics of monsoon rain in AGCM simulation of 6, 126, 115 and 0 ka, de Noblet et al. (1996) have shown the northwest/southeast redistribution of precipitation between the Indian and East Asian monsoon as well as the changes in the distribution of rainy events between high and low precipitation rates.

Several simulations clearly show the impact of land surface conditions on the monsoon, including analyses of the role of surface albedo (Street-Perrott et al., 1990; Bonfils et al., 2001), soil moisture (Kutzbach and Guetter, 1986; Kutzbach et al., 1996) and vegetation (Claussen and Gayler, 1997; Texier et al., 1997; Doherty et al., 2000). The latter studies indicate that the change in vegetation represents a strong feedback on the monsoon through both surface albedo and the capacity of vegetation to recycle water more efficiently than bare soil. Ocean feedback also plays an important role. It alters the characteristics of the Asian monsoon by introducing a late summer SST warming that acts to delay the retreat of the monsoon flow in late summer (Hewitt and Mitchell, 1998; Braconnot et al., 2000). The combination of vegetation and ocean feedbacks appears non-linear and enhances external insolation forcing (Braconnot et al., 1999). For most of these simulations, the focus was put on the African monsoon although there is relevance to the Indo-Asian systems as well.

Recent analyses extend the results of PMIP intercomparison to coupled atmosphere simulations of the Asian monsoon (Braconnot et al., 2005). Fig. 6 shows a small cooling during the middle Holocene in terms of the annual mean response of surface temperature over this region. The structure of this cooling is a response of the ocean to reduced insolation in the tropics during winter. Since the cooling is applied over a longer period than the increase summer insolation, the summer warming does not balance winter conditions (Fig. 7). Over land, the cooling in monsoon areas arises also from increased evaporation cooling in monsoon regions and increased cloud cover that reduces the amount of solar radiation reaching the surface. The seasonal cycle of the incoming insolation at the top of the atmosphere is about 5% larger in the NH. Because of this, the monsoon flow penetrates further inland during summer, which explains why the summer continental warming is associated with increased precipitation (Fig. 7). Over the ocean precipitation is decreased, except for two models that locally enhance precipitation over warmer than present surface waters. During winter the flux is intensified from the continent to the ocean, which contributes to enhance precipitation over the ocean. Here also, in some of the coupled models, local changes over the ocean counteract the large-scale effect of the land–sea contrast.

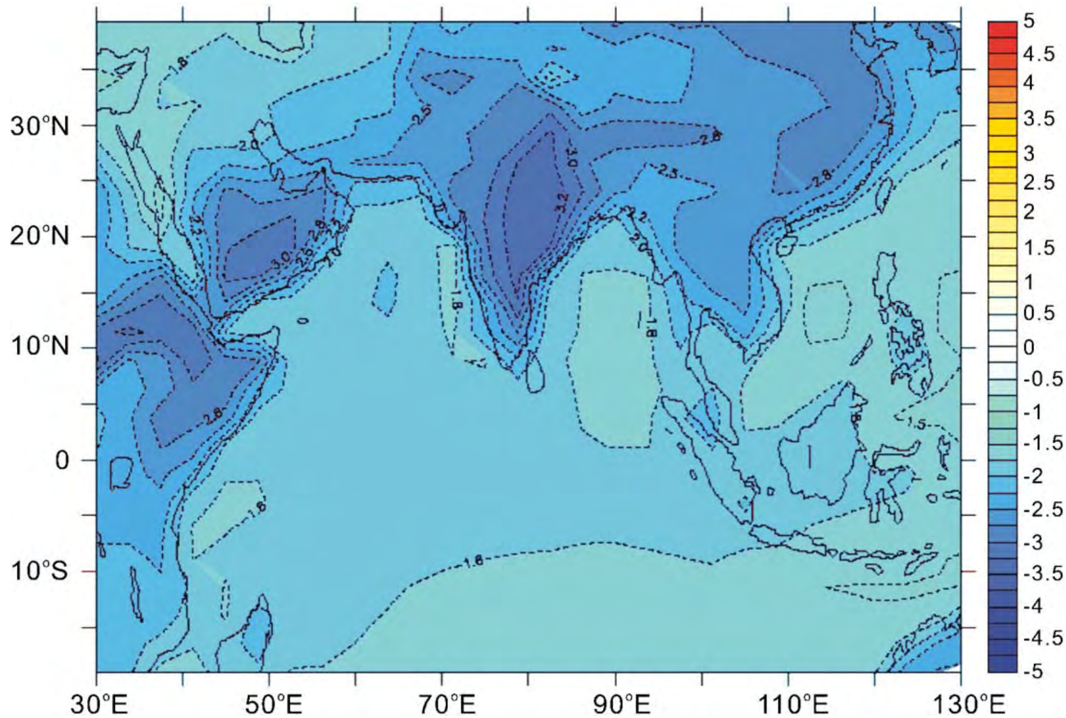

Fig. 6. Annual mean difference in surface temperature (in °C) between the mid-Holocene and the modern climates, as simulated using version 4 of the Institut Pierre Simon Laplace (IPSL-CM4) ocean–atmosphere coupled model without flux correction at the air–sea interface. Isolines are plotted every 0.15 °C.

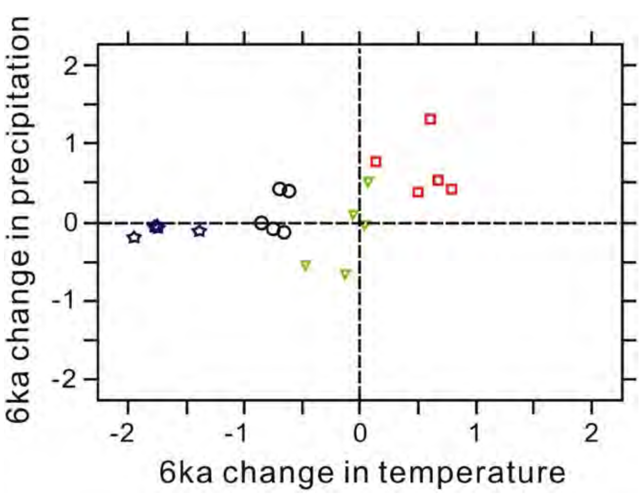

Fig. 7. Difference between mid-Holocene and modern climates for temperature and precipitation averaged over the Indian ocean sector (40°E–110°E; 20°S–40°N) as simulated by 5 different fully coupled ocean–atmosphere models (see Wohlfart et al., 2004, for a description of the simulations). Each dot represents one model result. Stars stand for winter averaged over land points, the dark circles for winter averaged over ocean points, the squares for summer over land and the triangles for summer over the ocean (adapted from Braconnot, 2004).

It is also interesting to note that the shape of the seasonal cycle of SST is quite different from that of the present day, as shown by the IPSL-CM4 coupled ocean–atmosphere model. Over the northwestern part of the basin, for example, the SST reaches a maximum in spring prior to the monsoon onset, then decreases, followed by a secondary maximum in autumn. For the mid-Holocene, the autumn maximum is dominant. In the eastern part of the basin, the SST remains stable all year long at 6 ka whereas it shows variations of about 1 °C for present day between spring and autumn. The timing and relative maxima of the Indian dipole that represents the SST meridional gradient across the Indian Ocean is also modified. In autumn the reinforcement of this dipole is associated with a slowdown of the monsoon retreat from the continent to the ocean, characterized by the maintenance of the ITCZ in the northwestern part of the basin corresponding to a late response of the ocean to summer warming and to a local enhancement of the warming.

The signature of the above changes in the mean seasonal cycle arise from the timing of the insolation forcing relative to the solstices and the equinoxes as recently shown by Braconnot and Marti (2003). Indeed, using a set of coupled ocean–atmosphere simulations they show that, depending on the timing of the seasonal cycle of insolation in the NH, two periods with enhanced monsoons can be characterized by different seasonal response of the Indian ocean, because of a different response of the hydrological cycle over the basin (precipitation, evaporation and river runoff). In particular, when the monsoon penetrates far into the

Himalayan foothills at the beginning of the monsoon season the river runoff in the Bay of Bengal can be twice as large as when the monsoon increase occurs later in the year. This fresh water input stratifies the surface ocean, reducing the mixed layer depth and thereby the thermal inertia of the surface ocean. In this case the surface ocean responds nearly in phase with the seasonal insolation forcing whereas when the mixed layer depth is deeper a one to two month delay strongly alters the shape of the SST seasonal cycle. These effects must be considered in model-data comparisons and in interpretation of proxy records.

Several authors have also investigated the link between changes in El-Niño variability and orbital forcing. Using a simplified (Zebiak and Cane, 1987) coupled ocean–atmosphere model, Clement et al. (1999) analyzed te evolution of El-Niño occurrence in response to changes in orbital parameters. The results suggest that ENSO variability changes in response to orbital variation, but does not vanish. During the middle Holocene, for example, ENSO events continue to occur roughly every 4 years, but there are fewer strong events and the mean amplitude of strong events is less than in the modern climate. Otto-Bliesner (1999) using a fully coupled OAGCM, found no change in the Niño 3 index for the mid-Holocene. However the teleconnection pattern was more confined to the Pacific Ocean than in the control climate. Liu et al. (2000) also investigated changes in ENSO in simulations of the Early Holocene and suggested that El Niño in the Early Holocene is reduced at least in part by the remote influence of the intensified Asian monsoon on the Pacific trades. Analyses of the different coupled simulations of the mid-Holocene, show that the simulated interannual variability is slightly reduced during mid-Holocene. In particular the variability of the Indian dipole is reduced in autumn as if it was damped by the larger dipole like structure present each year as a response to the seasonal insolation forcing. Similar analysis are now extended to Last Glacial climate and will be further analyzed during the second phase of the Paleoclimate intercomparison Project (Harrison et al, 2002; Braconnot et al., 2005).

4.4. Summary

The monsoon spectrum contains a rich set of frequencies related to primary orbital forcing as well as internal feedbacks within the climate system. Unraveling the forcing and response relationships involved requires the application of time series analytical techniques capable of resolving amplitude and phase relationships among climate proxies and insolation forcing. Analysis based on visual inspection is insufficient and can be misleading.

As clearly indicated by GCM experiments, monsoon variability is sensitive to any process that alters interhemispheric pressure gradients (winds) or the availability and transport of moisture. Thus, monsoon variability should not be considered only in the context of glacial–interglacial variability as is commonly the case, but also in the broader context of tropical and subtropical variability involving ocean–atmosphere interaction (SST and moisture flux).

Adequate spatial coverage of sufficiently long cores is necessary to unravel potential interaction among the Indian, East Asian and Indonesian–Australian summer- and winter-monsoon systems as well as the extent to which these systems interact with proximal equatorial regions outside the direct monsoon influence.

The monsoon system has a very large influence on the composition of underlying marine sediments. However, equally strong non-monsoon influences such as differential preservation and dissolution, sea level change, and physical sediment dynamics can also influence individual proxies to varying extents. Only through a multiproxy approach can the monsoon variance be isolated and analyzed.

5. Tectonic forcing and the long-term evolution of monsoon

5.1. Tectonic forcing

While pioneering works on the paleomonsoon concentrated on the Late Quaternary record, recent studies have traced monsoon history into early geological periods such as the late Triassic (e.g., Olsen, 1986; Kutzbach and Gallimore, 1989; Dubial et al., 1991). First principles suggest that monsoon circulation must exist through geological history whenever the tropics are occupied by land and sea, but it varies greatly with tectonically induced geographic and topographic changes. Such changes occur on the order of 10^6–10^7 years and represent the longest time scale changes in the monsoon spectrum of variability (Fig. 4). Three tectonic factors have been proposed as exercising control over the evolution of Asian monsoon circulation: plateau uplift, sea–land distribution, and oceanic gateways.

Himalayan–Tibetan *plateau uplift* is the most widely discussed tectonic factor responsible for altering monsoon circulation. GCM experiments indicate that strong monsoons (similar to modern) can be induced by solar forcing only when the elevation of Tibet–Himalaya is at least half that of today (Prell and Kutzbach, 1992). The main problem in testing this uplift–climate relationship is the poor constraints of the Tibetan Plateau elevation history (e.g., Copeland, 1997). Nevertheless, it is now generally accepted that Himalayan–Tibetan plateau uplift has been a step-wise process. A number of studies around the end of 1980s investigated the climate consequences of uplift (e.g., Kutzbach et al., 1989;

Ruddimann et al., 1989; Ruddiman and Kutzbach, 1989), and it was found that uplift may be responsible for both late Cenozoic cooling and significant strengthening of the Asian monsoon system. Geological data and computer modeling were used to support the hypothesis of intensified uplift of the Tibetan Plateau around 8 Ma causing enhanced aridity in the Asian interior and onset of the Indian and east Asian monsoons (Fig. 8A, Prell and Kutzbach, 1997; An et al., 2001). Although the age of the Asian monsoon system is challenged by the new discovery of Miocene loess (Guo et al., 2002), intensification of Asian aridity and monsoons at about 3 Ma ago is convincingly documented both on land and in the ocean, a development probably not related to uplift but the permanent glaciation of Greenland.

The question concerning the role of *sea–land distribution* in monsoon evolution was raised by Ramstein et al. (1997) on the basis of AGCM simulation results. The Paratethys, a epicontinental sea stretching over Eurasia 30 Ma ago, progressively receded during the Miocene, resulting in continentalization of the Asian interior and enhancement of monsoon circulation. They found that the retreat of the Paratethys played as important a role as uplift of the Himalayan/Tibetan Plateau in Asian monsoon development. As the shrinkage of the Paratethys occurred from the Oligocene to the Middle-to-Late Miocene, its monsoon effects should be manifest between 30 and 10 Ma, significantly earlier than the 8 Ma hypothesis. If we go further back to the Eocene, when Europe was separated from Asia by an epicontinental sea, GCM simulation results show a zonal distribution of high precipitation in the tropics and much more arid conditions in east China (P. Wang, in press). This pre-monsoonal climate pattern corresponds well with the paleoenvironmental records (P. Chen et al., 2000). All these findings underscore the role of sea–land patterns in the evolution of a monsoon climate.

The third aspect is the role of *oceanic gateways*. An oceanic circulation model revealed that 'closure' of the Indonesian seaway 3–4 Ma ago could be responsible for east African aridification (Cane and Molnar, 2001). The authors found that the northward shift of the Australian Plate may have switched the source of the Indonesian Throughflow from warm south Pacific to relatively cold North Pacific waters. This would have decreased SSTs in the Indian Ocean, leading to reduced precipitation in east Africa and to decreased strength of the Indian summer monsoon in general. Similar changes in oceanic circulation took place in the Western Pacific, in particular in the marginal seas, where the numerous

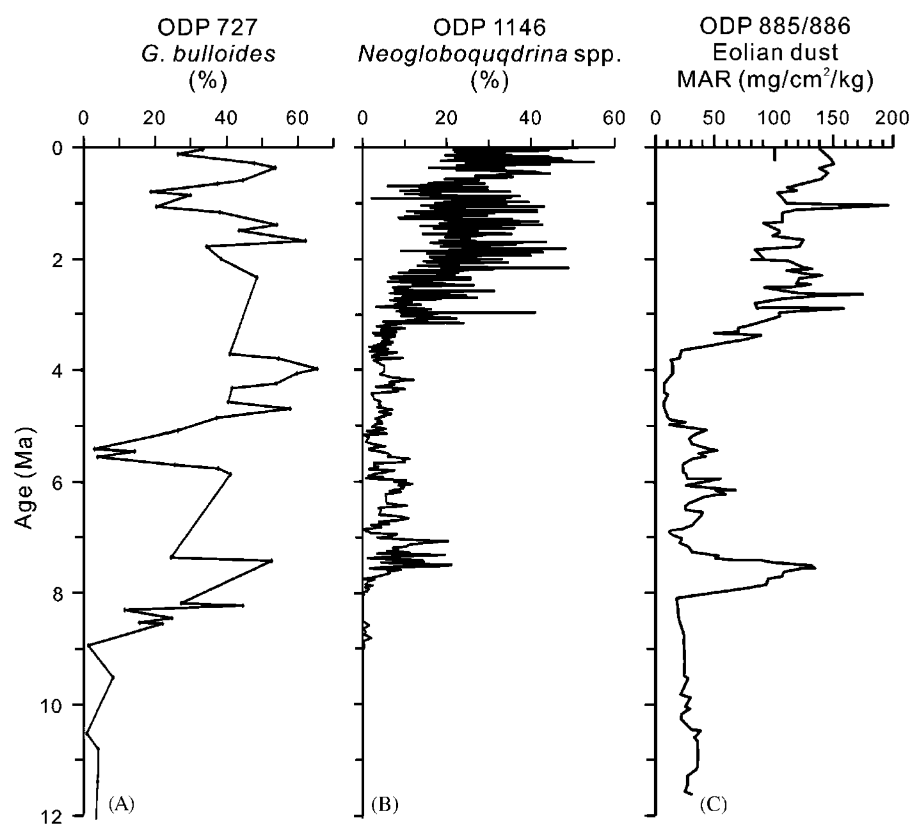

Fig. 8. Marine records of Asian monsoon and aridity evolution over the past 12 Ma. (A) *G. bulloides* % (>150 μm), ODP site 722, Arabian Sea (Prell et al., 1992); (B) *N. dutertrei* %, ODP Site 1146, SCS (J. Wang, 2001; P. Wang et al., 2003b); (C) dust flux (mg/cm²/ka), ODP sites 885/886, central Pacific (Rea et al., 1998).

gateways are highly sensitive to tectonic and eustatic changes. Since the Western Pacific Warm Pool exerts great influence on monsoon circulations (Webster et al., 1998), the role of the Western Pacific and its boundary currents should not be ignored. In fact, the potential roles of changes in the West Pacific Warm Pool and the Indoneasian Throughflow have been implicated recently in late Quaternary climate trends within the monsoon-influenced Maritime Continent region (Kershaw et al., 2002).

Although it is unanimously agreed that the long-term evolution of the Asian monsoons depends on tectonic changes, opinions diverge on which of the tectonic factors exercises the major control. The uplift of the Tibetan Plateau is the most widely discussed factor, but it does not mean that the sea–land distribution is inferior in importance, or other topographical features can be ignored. The development of some geographic features outside Asia may also affect the Asian monsoon system, such as the uplift of the East African Plateau or the closure of the Panama gateway (Haug et al., 1999). At the present stage it is believed that all three factors are significant, but their relative roles remain unclear. To single out the role played by each of the factors, we need much better constraints on the timing of the tectonic and climate events. Of crucial importance is a number of long sequences of paleomonsoon records from various areas coupled with high-quality chronologies.

5.2. Long-term evolution

5.2.1. Late Miocene hypothesis

Long-term monsoon records have been developed from DSDP/ODP cruises to the Indian Ocean, the Mediterranean Sea, and the SCS (Leg 22, 24, 115, 116, 117, 121, 184), as well as from studies of the Chinese Loess Plateau. On the basis of the oldest recorded occurrence of diatoms at DSDP Site 238 as indicator of monsoon-driven productivity, Burckle (1989) suggested that the initiation of the modern monsoon system over the Indian subcontinent was about 10–11 Ma. Similar records with the initiation of opal accumulation near 11.5 Ma were found at ODP Site 722, Arabian Sea, although the monsoon upwelling indicator, *G. bulloides*, significantly increased only about 8.5 Ma (Kroon et al., 1991; Prell et al., 1992). This latter date is very close to the rapid ecological transition from C3-dominated to C4 dominated vegetation about 7.4–7.0 Ma as revealed by the $\delta^{13}C$ data of pedogenic carbonates from northern Pakistan in the Himalayan foreland, interpreted then as the origination or intensification of the Asian monsoon system (Quade et al., 1989; see below Section 5.2.2). In addition, the dating of extensive faults suggests significant Tibetan Plateau uplift/extension about 8 Ma (Harrison et al., 1992; Molnar et al., 1993). Based on these finding and GCM simulations, it was hypothesized that the Tibetan Plateau may have reached at least half its present height at ~8 Ma, causing intensification of the Asian monsoon (Prell and Kutzbach, 1992; Prell et al., 1992). Since then, a large number of new results have contributed to the understanding of the long-term history of the Asian monsoon system. Because the recent land-based paleomonsoon reviews are usually focused on the Quaternary only, our following discussion will cover both the terrestrial and marine records.

5.2.2. Terrestrial records

Toward the end of 1980s, intensive paleomonsoon study of the loess–paleosol sequence in China began. The origin of the loess/paleosol sequences at about 2.6 Ma was ascribed to the development of a system of alternate predominance of the winter and summer East Asian monsoons. Later, Chinese scientists found that the Red Clay underlying the loess sequence was also of wind-blown origin, and the history of aeolian deposits of the Loess Plateau was extended to 7–8 Ma (Sun et al., 1997, 1998; Ding et al., 1998a, 2001a; An et al., 2001). However, it remains debatable whether the fine-grained aeolian deposits were transported by the westerlies or the winter monsoon (Ding et al., 1998b), and whether the lower part of the red clay is composed of water-reworked deposits related to alluvial and slope processes (Guo et al., 2001). Since the uplift of the Himalayan Mountains and the Tibetan Plateau can lead to enhanced aridity in the Asian interior and to intensification of the Asian monsoon system (Kutzbach et al., 1993), the nearly coincident beginning of upwelling in the Indian Ocean and dust accumulation in central China about 8 Ma was explained by the onset of the Indian and east Asian monsoons, which in turn implies a significant increase in altitude of the Plateau (An et al., 2001).

The Chinese dust history has been further extended by the recent discovery of a Miocene loess sequence from Qinan, western Loess Plateau (Guo et al., 2002). A total of 231 interbedded loess–paleosol layers represents a nearly continuous history of aeolian dust accumulation from 22 to 6.2 Ma, indicating that large source areas of aeolian dust and energetic winter monsoon winds existed since early Miocene, at least 14 Ma earlier than previously thought. The discovery was explained by the authors as the ‘‘onset of Asian desertification’’, resulting from tectonic changes and global cooling.

On the other hand, an early Miocene change in climate was indicated by other studies. In East China a transition from arid to humid climatic conditions around the Oligocene/Miocene boundary was discovered more than 20 years ago. Summarizing the data from oil exploration and stratigraphic studies in China, it was found that a broad belt of aridity stretched across China from west to east in the Paleogene, particularly the Paleocene, while in the Neogene the arid zone was restricted to Northwest China, suggesting a transition from a planetary to monsoonal system in atmospheric circulation (P. Wang, 1984; Zhou, 1984). In other words, it was the East Asian summer monsoon that brought moisture from the ocean to East China, and the reorganization of the climate system around the Oligocene/Miocene boundary provides evidence for enhancement if not establishment of the East Asian summer monsoon (P. Wang, 1990; Liu, 1997; Sun and Wang, MS). The loess–paleosol sequence at Qinan supports the Oligocene/Miocene transition in East Asian climate, with loess as evidence for enhanced aridity in the dust source areas and energetic winter monsoon winds required for dust transport, and with paleosols indicating increased moisture supply by summer monsoon winds. New results of pollen analyses from Linxia, another site in the western Loess Plateau near Qinan, clearly show a transition from sparse-woods/steppe to forest at about 22 Ma, where the Oligocene climate was definitely drier than that of the Miocene (Ma et al., 1998). In fact, the arid climate in Northwest China can be traced back at least to the Late Cretaceous, but the accumulation of a thick sequence of loess–paleosol deposits requires not only dust but also moisture. The paleosol layers at Qinan show significantly more intensive weathering than those in the Pleistocene loess–paleosol sequence (Guo, personal communication).

The early occurrence of the Asian monsoon was reported also from northern Thailand where the middle Miocene mammalian faunas about 16–14 Ma suggest a monsoon-like wet climate (Ducrocq et al., 1994). This fits well with the finding that the southern Tibetan Plateau reached its present altitude about 15 Ma ago (Spicer et al., 2003), if the uplift is taken as the main factor in driving monsoon development.

A trustworthy reconstruction of the monsoon history depends on the proxies adopted. An example is the late Miocene development of C4 vegetation exhibited by $\delta^{13}C$ shift in pedogenic carbonates. When first found in Pakistan, it was interpreted as a signal of origination of the Indian monsoon (Quade et al, 1989). Later, with its findings in various continents, it was re-considered as a global vegetation change probably caused by decreased CO_2 concentration (Cerling, 1997; Cerling et al., 1997) and remains a debatable issue. This example illustrates the complexity in use of proxies.

5.2.3. Marine records

The uplift-induced monsoon intensification at 8 Ma was further challenged by sedimentological data from ODP Leg 116, Bay of Bengal. Although the dramatic increase of $\delta^{13}C$ of total organic carbon in Bengal Fan sediments at ca 7 Ma supports the development of the monsoon in the Himalayan foreland at that time (France-Lanord and Derry, 1994), the sediment accumulation records are in conflict with the 8 Ma uplift model. The accumulation rates of the Bengal Fan as recorded in ODP Sites 717 and 718, as well as in DSDP site 218 and 222, were high from 17 to 7 Ma, but decreased from 7 to 1 Ma with the clay mineral assemblages indicating reduced physical erosion and strengthened chemical weathering (Burbank et al., 1993; Derry and France-Lanord, 1996, 1997). Analyses of Nb and Sr isotopes also show that early Miocene sediments of the Bengal Fan have the same source as the younger one (Galy et al., 1996), again at odds with the late Miocene uplift hypothesis.

The same question was raised from the Indus Fan studies. Analyses of DSDP/ODP sediment cores suggest that the Indus River and Fan system has been draining western Tibet since ~50 Ma, and the erosion appears to peak in the middle Miocene (Clift, 2001, 2002), implying a much earlier uplift of the Tibetan Plateau than the 8 Ma hypothesis proposed.

No remarkable change in sediment accumulation was found in the ODP Leg 184 to the SCS around 8 Ma (P. Wang et al., 2000). On the other hand, the frequency of the planktic foraminifer *N. dutertrei*, an indicator of the East Asian monsoon and enhanced productivity in the SCS (Jian et al., 2000a, b), increased abruptly at Site 1146 at 7.6 Myr and again, at 3.2 Ma (Fig. 8B; J. Wang, 2001; P. Wang et al., 2003b). This corresponds well to the Indian monsoon records (Prell et al., 1992). Another interesting finding from the SCS is the early appearance of C4 plants in the East Asian vegetation at least since ~20 Ma. The isotopic analyses of black carbon from deep-sea sediments at ODP Site 1148 revealed 5 positive excursions in the 30 Ma record, and the earliest one is from the early Miocene, significantly predating the 7–8 Ma age of sudden expansion of C4 plants in Pakistan. The authors ascribed the new finding to initiation of the East Asian monsoon (Jia et al., 2003), although this can not be conclusive evidence as the C4 plants expansion may be caused by factors other than monsoon (see Section 5.2.2).

Thus, the Asian monsoon system has a longer history than previously thought, but the system also displays great variability both in space and in time. The Miocene sequence of Qinan does not show any obvious long-term intensification of loess deposition, but two intervals are distinguished by higher dust accumulation: 15–13 and 8–7 Ma (Guo et al., 2002). These might represent periods of enhanced aridity in the source areas, an interpretation supported by a new pollen diagram from Yumen, northeast of Tibet, which records an arid–semiarid climate since 13.8 Ma with an interval of increased humidity between 11.2 and 8.6 Ma (Ma et al., 2004), broadly correlated with the Qinan record. The increased aridity around 8–7 Ma was recorded also in the North Pacific as a peak in dust accumulation rate (Rea et al., 1998), but this interval of higher dust accumulation was not sustained, unlike the major increase since about 3.5 Ma (Fig. 8C).

5.3. Summary

The new records from the SCS and Loess Plateau indicate similar stages in the development of the East and South Asian monsoons, with an enhanced winter monsoon over East Asia being the major difference (P. Wang et al., 2003b). However, the number and geographic coverage of monsoon records decreases with increasing age; our knowledge of pre-Quaternary monsoon history remains relatively deficient. Of paramount importance is the production of more long-range high-quality marine and continental records, including hemipelagic and lacustrine sediments of Miocene and Paleogene age. Only long-range records can provide the opportunity to test the numerous hypotheses regarding monsoon evolution and the relative roles played by uplift, sea–land distribution and oceanic gateways.

6. Recommendations

6.1. Process observations and monsoon proxies

With growing interest of the scientific community and a growing number of publications dealing with paleomonsoon evolution, care is necessary to minimize misuse of monsoon proxies. Because most natural processes are driven by more than one factor, it is crucial to distinguish monsoon and non-monsoon variability within proxies; to this end long-term modern observations are useful.

During the past decade, the Indian Ocean, particularly the Arabian Sea, has been a major focus of JGOFS studies, and many expeditions studied interseasonal and interannual variations of the Indian monsoons. Compared to those in the Indian Ocean, long-term observations are rare in the Western Pacific where the East Asian monsoon prevails. For example, the journal "Deep-Sea Research II" published 12 special issues resulting from expeditions to the monsoonal Indian Ocean, mostly to the Arabian Sea (see "Introduction"), but only one for the Western Pacific marginal seas. There are only two sediment trap studies reported from the SCS, as compared to 12 from the Arabian Sea (Table 2; Fig. 1). Particular attention should be given to development of winter monsoon proxies in East Asia and the West Pacific.

To improve our understanding of the monsoon records, we need internationally coordinated efforts to obtain core-top data sets calibrated to a broad set of sediment trap time series. Coeval long-term series of sediment traps will be crucially important from basins relevant to monsoon studies such as the Bay of Bengal, Pakistan margin, Somalian margin, South and East China Sea, Sulu Sea and others. These should be accompanied by coeval satellite meteorological observations.

6.2. High-resolution long-term records

We regard phase relationships as the key to undertand the origin of variations in monsoon circulation. This requires high-resolution records with accurate chronostratigraphies. Thus, we need long, undisturbed, continuous, and large-diameter cores (40–60 m length) from varved sediment archives. Implementation of IODP leg(s) to the Northern Arabian Sea is a high priority in this regard. Additional varved sediment archives can perhaps be expected along the western margin of India or along the margin of the Gulf of Bengal.

Dating precision in marine sediment archives requires substantial improvements for the last 100 ka, especially by employing annual-layer counts over long time series to arrive at a first correlation of paleomonsoon records with the paleoclimatic records obtained from the ice cores of Greenland, a correlation that needs to be far better substantiated than at present. New long-term coral ring records may play a key role in these correlation efforts as will ash layers and geomagnetic intensity events for the long-distance correlation of stratigraphic events.

6.3. Monsoon, ENSO and tropical– extratropical linkage

Understanding the role of the Asian monsoon in the context of climate change at orbital through sub-orbital times scales requires a broad geographic distribution of records covering the Australian-, East Asian-, and Indian-monsoon regions. Only with adequate coverage can differing regional responses be identified and understood in the context of global teleconnections and their underlying physics. For example, the relationship between the Asian monsoons, ENSO and Indian Ocean Dipole has received considerable attention lately. Modeling experiments and meteorological data indicate reasonably complicated, region-dependant links between the monsoon and ENSO (Lau and Bua 1998; B. Wang et al., 2003). During the ENSO initiation phase (June–August), for example, Arabian Sea summer-monsoon winds are weakened while those in the SCS are much strengthened. In contrast, during the mature phase of ENSO (December–February) SCS winter-monsoon winds are significantly weakened while Arabian Sea winter-monsoon winds are only slightly weakened (B. Wang et al., 2003). Mechanistically, some observations suggest that an intensified winter monsoon in East Asia leads to El Niño (Li and Mu, 2000). Similarly, a 150-year coral record in the tropical Indian Ocean suggests that a 12-year monsoon cyclicity may modulate the quasi-biennial cycles of ENSO (Charles et al., 1997).

On the longer time scale, precession cycles related to an ENSO-like pattern apparently influenced the monsoon as far as the equatorial Indian Ocean and the Somali coast (Beaufort et al., 2001). Modeling by Clement et al. (1999, 2001) indicates that ENSO shutdowns may lead to a semi-precession cyclicity (11 ka) in the monsoon-influenced sediment record. Testing these region-dependent types of modeling results and verifying mechanistic interpretations of individual climate records require adequate spatial coverage of long, continuous, large-diameter cores from high sedimentation rate or varved sediment archives. These goals, in turn, require resources such as multi-beam echosounding and high-resolution seismic survey in order to identify optimal coring and drilling targets.

6.4. Integrated monsoon transect

Precise land–sea correlation of paleomonsoon records is badly needed for improving our understanding of the underlying mechanisms. Moreover, we need a great number of long records to provide a sufficient spatial and temporal resolution of monsoon signals in the ocean and on land to trace the intensity and speed of change. Transects of long cores (such as IMAGES cores) targeting areas with sedimentation rates greater than 10 cm/ka spanning the last 500 ka will provide the temporal and spatial gradients necessary to evaluate threshold changes and the processes of past monsoon variability in general.

On the tectonic time scale, long records over 20–40 Ma in strategic areas with hemipelagic deposits are needed to evaluate processes related to monsoon variability such as precipitation, continental weathering, and to balance the discharge of eolian and fluvial sediments from the continent to the ocean. This contribution will likely stem from IODP drilling.

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