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EVOLUTION OF THE INDIAN MONSOON SINCE LATE MIOCENE INTENSIFICATION - MARINE AND LAND PROXY RECORDS

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ABSTRACT

A large part of the Indian Ocean and its surrounding continents are influenced by seasonally reversing (summer and winter) monsoon winds. Variability in the summer monsoon impacts different components of the Earth system, both regionally and globally. The Indian monsoon constitutes a critical resource for the region's largely agrarian economies, as almost two thirds of India's food production depend on summer rains, so are the rivers that cater to the domestic needs of the region. Evolution of the Indian monsoon has been closely related to the uplift of the Himalaya-Tibetan plateau which is believed to have intensified the monsoon during 10-8 Ma. However, there are debates about the relation of the 10-8 Ma event with the elevation changes in the Himalaya-Tibetan plateau. The elevated heat source of the Himalaya-Tibetan complex is of vital importance for the establishment and maintenance of the Indian summer monsoon circulation.

The strength of the monsoon is controlled by a number of forcing factors operating over a variety of time scales. While long term changes in the Indian monsoon have been linked to the phased uplift of the Himalaya-Tibetan plateau superimposed by orbital changes, small scale, rapid changes as documented in late Quaternary and Holocene proxy records from marine sequences, cave deposits, peat deposits, runoff in the Bay of Bengal, and fluvial sediments have been related to boundary conditions including Himalayan-Tibetan snow, North Atlantic variability, Eurasian temperatures, tropical sea surface temperatures, solar activity, vegetation changes, and linkages with the ENSO, Indian Ocean Dipole (IOD) or North Atlantic Oscillations.

Keywords: Indian Monsoon, late Miocene, proxy records, Arabian sea, eastern Indian Oceen, Pakistan Siwaliks and Nepal Himalaya

INTRODUCTION

The Indian monsoon, also known as the South Asian monsoon, described in all its glory and fury over the Indian subcontinent, is a unique climatic feature marked by regularly occurring seasonal reversals in wind direction. During the summer (June-September), monsoonal winds are southwesterly whereas during the winter season (December-February) winds are northeasterly. The summer or southwest (SW) monsoon plays an important role in global hydrological and carbon cycles, affecting climate and societies over large parts of Asia. The summer monsoon rains are critical for food production, water supply and economic well-being of the Asian societies. Almost two thirds of India's food production depend on summer rains, so are the rivers that cater to the domestic needs of the region. Thus the Indian monsoon constitutes a critical resource for the region's largely agrarian economies.

Evolution of the Indian monsoon has been closely related to the uplift of the Himalaya-Tibetan plateau. The elevated heat source of the Himalaya and the Tibetan plateau is of vital importance for the establishment and maintenance of the Indian summer monsoon circulation through mechanical and thermal factors. However, there are different propositions about the timing of attainment of the critical elevation by the Himalaya-Tibetan plateau complex to drive the Indian monsoon, ranging from 35 to 7.5 Ma (table 1). While the northern Indian Ocean marine records indicate a major strengthening of the Indian monsoon between 10 and 8 Ma (Kroon *et al.*, 1991; Gupta and Srinivasan, 1992; Gupta *et al.*, 2004), the continental vegetation captured this shift somewhat later between 8 and 7 Ma (Quade *et al.*, 1989; Quade *et al.*, 1995). Recent study from China suggests a wet phase in the early Miocene and beginning of an arid phase (weakening of the summer monsoon) during 13-11 Ma (Jiang *et al.*, 2008). A recent model study, on

Table 1: Evidence and timing of the Himalayan uplift and monsoon intensification (modified from Gupta *et al.*, 2004)

Source	Type of evidence	Event	Timing (Ma)
Rowley and Currie, 2006	Oxygen isotope	Tibetan Plateau	35
Ramstein *et al.* (1997)	Modeling	Monsoons and Paratethys retreat	~30
Guo *et al.* (2002)	China loess deposits	Monsoon climate	22
Wang (1990)	Sediments in China	Monsoons	20
Clift and Gaedicke (2002)	Indus Fan sediments	Erosion and weathering	~16
Clift *et al.* (2002)	South China Sea smectite mineral	Precipitation and monsoons	~15.5
Spicer *et al.* (2003)	Fossil flora	Himalayan elevation and monsoons	>15
Coleman and Hodges (1995)	Tectonics	Himalayan elevation	>14
Blisniuk *et al.* (2001)	Tectonics	Himalayan uplift and monsoons	>14
Chen *et al.* (2003)	Oceanic microfossils	Monsoons and upwelling	12–11
Dettman *et al.* (2001)	Isotopes and land	Monsoons	~10.7
An *et al.* (2001)	Land and marine sediments	Uplift and onset of monsoons	9–8
Kroon *et al.* (1991)	Oceanic microfossils	Monsoons and upwelling	8.6
Filipelli (1997)	Weathering and sediments	Monsoons	~8
Quade *et al.* (1989)	Isotopes and flora	Monsoons	8–7.6

* 28th M. R. Sahni Memorial Lecture deliverd on 8th March, 2010.

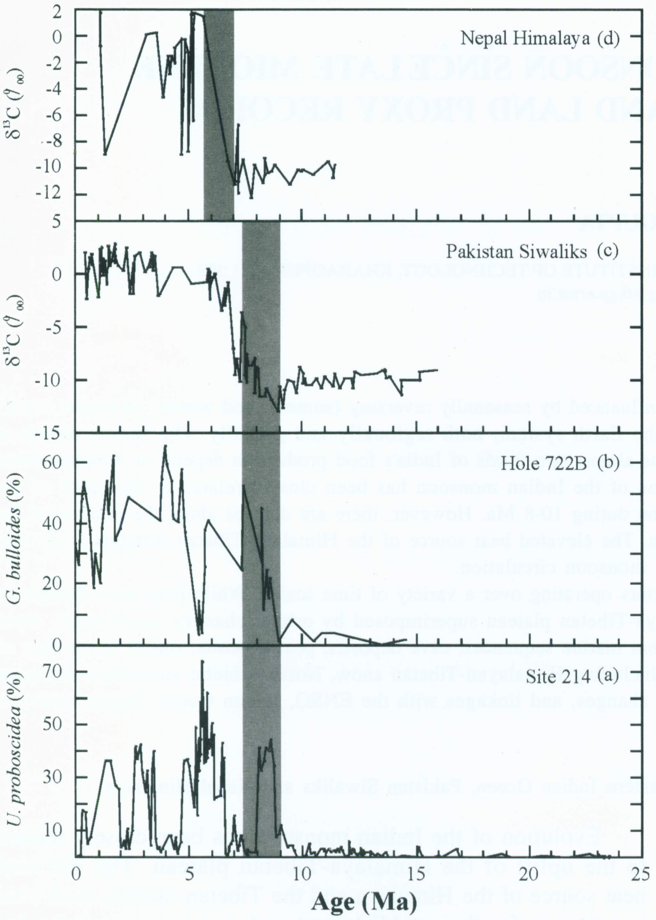

Fig. 1. Proxy record of monsoon from the northern Indian Ocean, Pakistan Siwaliks and Nepal Himalaya, showing a major increase during 10-8 Ma. (a) Percent distribution of benthic foraminifer *Uvigerina proboscidea* at DSDP Site 214 (Gupta and Srinivasan, 1992), (b) Percent distribution of planktic foraminifer *Globigerina bulloides* at ODP Site 722 (Kroon *et al.*, 1991), (c) $\delta^{13}C$ values of paleosols from the Pakistan Siwaliks (Quade and Cerling, 1995), and (d) $\delta^{13}C$ values of soil carbonate from the Nepal Himalaya (Quade et al., 1995).

the other hand, suggests a major increase of the Tibetan plateau elevation ~ 35 Ma which may have intensified the Indian monsoon (Rowley and Currie, 2006).Thus to resolve these issues, a coordinated effort is required to analyze and compare high resolution records from marine cores from high sedimentation areas of the northern Indian Ocean as well as continental records of continuity.

While long term changes in the Indian monsoon have been linked to the phased uplift of the Himalaya-Tibetan plateau (Zhisheng *et al.*, 2001), small scale, rapid changes as documented in late Quaternary and Holocene proxy records from marine sequences (Schulz *et al.*, 1998; Gupta *et al.*, 2003), cave deposits (Fleitmann *et al.*, 2003), peat deposits (Hong *et al.*, 2003), runoff in the Bay of Bengal (Kudrass *et al.*, 2001), and fluvial sediments (Sharma *et al.*, 2004) have been related to boundary conditions including Himalayan-Tibetan snow, North Atlantic variability, Eurasian temperatures and solar activity. Extreme monsoon events had a potentially dramatic effect on the fluvial systems (Goodbred and Kuehl, 2000), the terrestrial and marine fauna and flora, and human populations in Asia during the Holocene (Gupta *et al.*, 2003; Gupta, 2004). To understand impact of monsoon variability on marine and terrestrial fauna and flora, this study reviews published reports on long-term monsoon variability since the late Miocene and short-term changes during the Quaternary and the Holocene.

MATERIALS AND METHODS

This study combines published marine data from Ocean Drilling Program (ODP) Site 722 (Arabian Sea) and Deep Sea Drilling Project (DSDP) Site 214 (eastern Indian Ocean) with continental records from the Pakistan Siwaliks and Nepal Himalaya to analyze late Miocene intensification of the Indian monsoon (fig. 1). The Plio-Pleistocene record is from Site 758 (figs. 2-4), whereas late Quaternary and Holocene records are

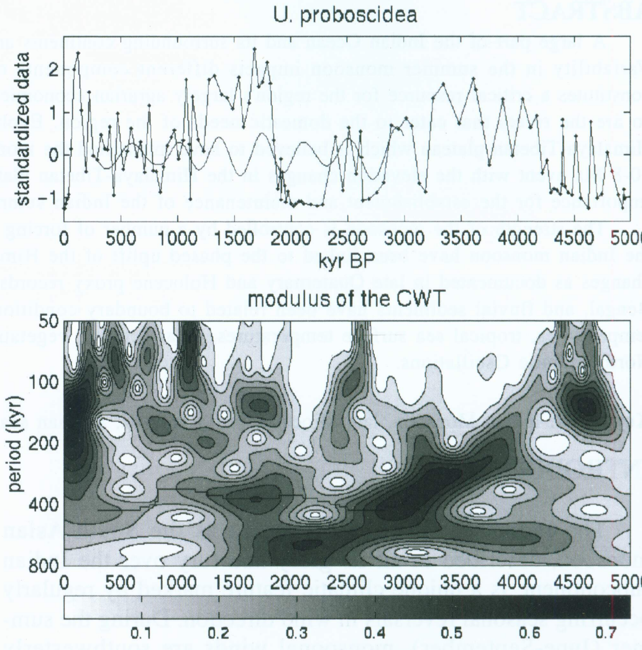

Fig. 2. *Uvigerina proboscidea* and its ~400-kyr component (top). CWT modulus and ~400-kyr ridge (thick line) of the *U. proboscidea* signal (bottom) at ODP Site 758 (reproduced from Gupta and Melice, 2003).

from the Arabian Sea ODP Site 723 (northwestern Arabian Sea; Gupta, 2008) and core SO90-111KL (off Pakistan, northeastern Arabian Sea; Schulz *et al.*, 1998) combined with land records from Oman, China, Tibetan Plateau and North Atlantic (figs. 5-6). An attempt has also bee made to understand decadal scale changes in the summer monsoon during the past millennium using both marine and land proxy records (figs. 7-8).

DISCUSSION

Continuous and well-preserved marine and land records from the northern Indian Ocean, Oman and China suggest dramatic changes in the behaviour of the Indian monsoon since the late Miocene, which drove significant changes in the fauna and flora including marine biota, continental vegetation and vertebrate animals. These changes had major impacts on development of the south Asian human societies.

Late Miocene (10-8 Ma) monsoon intensification

Both marine and continental records from the northern Indian Ocean and Indian subcontinent show a major increase in monsoon intensity during the late Miocene (fig 1). While this shift in monsoon intensity is seen at ~9 Ma (ranging from 10 to 8 Ma) in the marine records, the continental records from Pakistan captured it at ~8 Ma and those from the Nepal Himalaya at ~7 Ma. There is sufficient evidence from the ma-

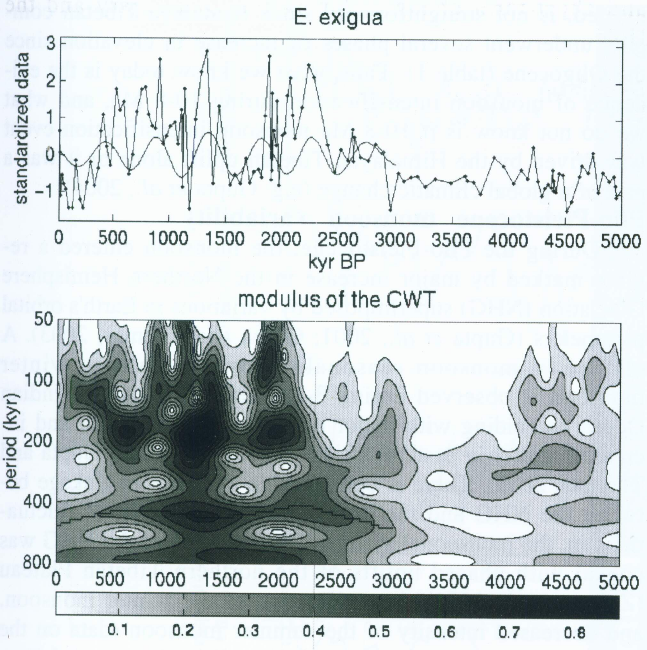

Fig. 3. *Epistominella exigua* and its ~400-kyr component (top). CWT modulus and ~400-kyr ridge (thick line) of the *E. exigua* signal (bottom) at ODP Site 758 (reproduced from Gupta and Melice, 2003).

rine records to suggest strengthening of monsoonal winds during the latest Miocene that caused widespread upwelling and high productivity in different parts of the northern Indian Ocean. The first record of monsoonal upwelling comes from cold-water diatoms from the Indian Ocean at about 7 Ma (Burckle, 1989). Likewise, a dramatic increase in an upwelling-indicator planktic foraminifer *Globigerina bulloides* from Ara-

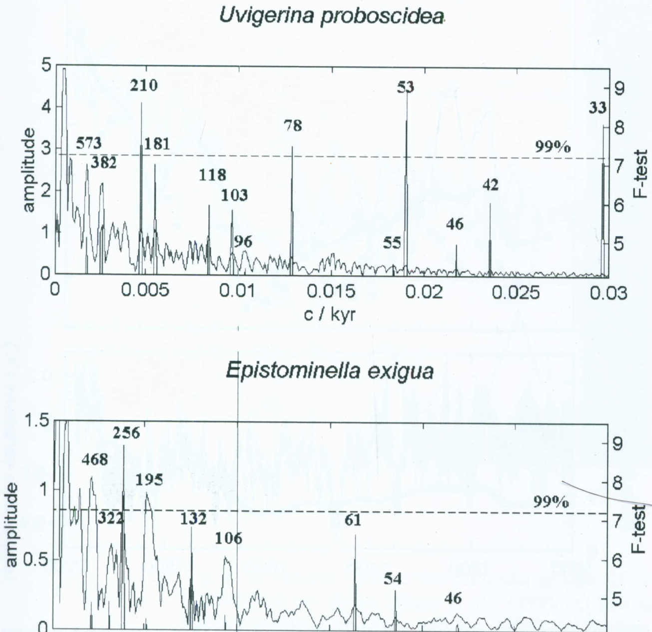

Fig. 4. Multi-taper harmonic analysis amplitude (continuous curve) and F-test values (peaks) of the *U. proboscidea* and *E. exigua* signals from ODP Site 758, eastern Indian Ocean (Gupta and Mélice, 2003). Only the F-test values above the 95% level are displayed. The 99% significance level for the F-test is also displayed. The two records show different orbital cycles.

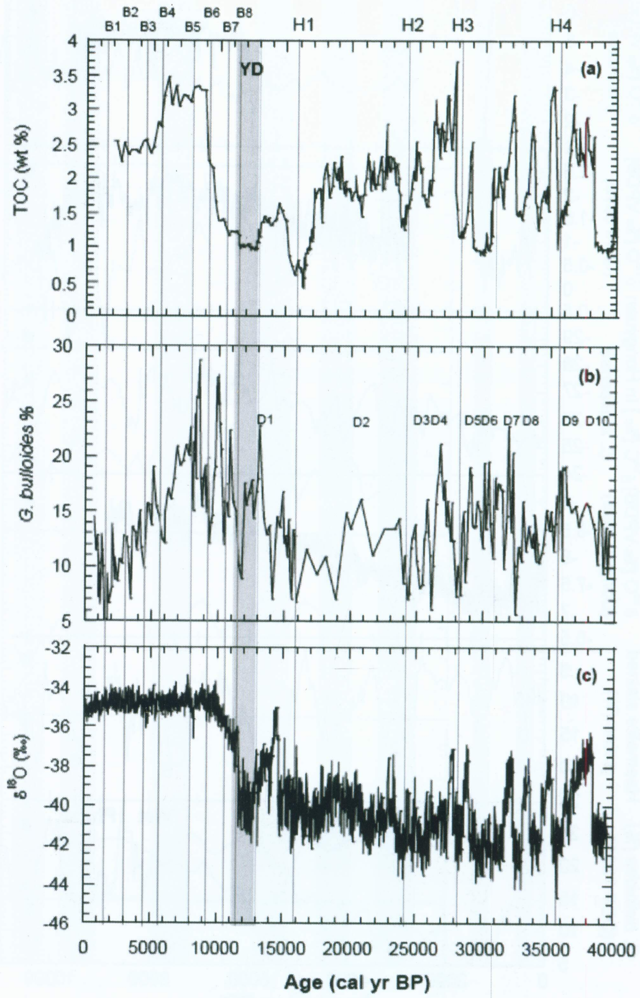

Fig. 5. Summer monsoon proxy record from the Arabian Sea (reproduced from Gupta, 2008). **(a)** Marine Total Organic Carbon (TOC) record of SO90-111KL off Pakistan, northeastern Arabian Sea (Schulz *et al.*, 1998), **(b)** *G. bulloides* percentages in Hole 723A, **(c)** GISP2 ^{18}O record (Grootes *et al.*, 1993). B1-8 = Bond events; H1-4 = Heinrich events; D1-10 = D-O events.

bian Sea Site 722 and in a benthic foraminifer *Uvigerina proboscidea* from eastern Indian Ocean Site 214 at ~9 Ma mark an intense upwelling caused by strong summer monsoon winds (Kroon *et al.*, 1991; Gupta and Srinivasan, 1992). The monsoon-linked terrigeneous flux to the ocean increased during 9 to 7 Ma (Rea, 1992).

The palaeovegetational records from the Pakistan Siwaliks and Nepal Himalaya show a marked shift from C3 to C4 type vegetation between 8 and 7 Ma (Quade *et al.*, 1995; Behrensmeyer *et al.*, 2007). The expansion of C4 grasses during this time is explained in light of the advantages that C4 grasses enjoy over C3 grasses in warm and humid monsoonal climates (Quade *et al.*, 1995). Quade *et al.* (1989) initially suggested that the abrupt appearance of C4 grasses in Pakistan may have been preceded by a period of rapid uplift of the Himalaya and Tibetan plateau which may have intensified the monsoon. Though the evidence of the uplift of the Himalaya was immediately captured by the marine proxies, it took a million years for vegetation change in Pakistan and a little less than two million years for vegetation in the Nepal Himalaya to record this change. The cause of this late response of conti-

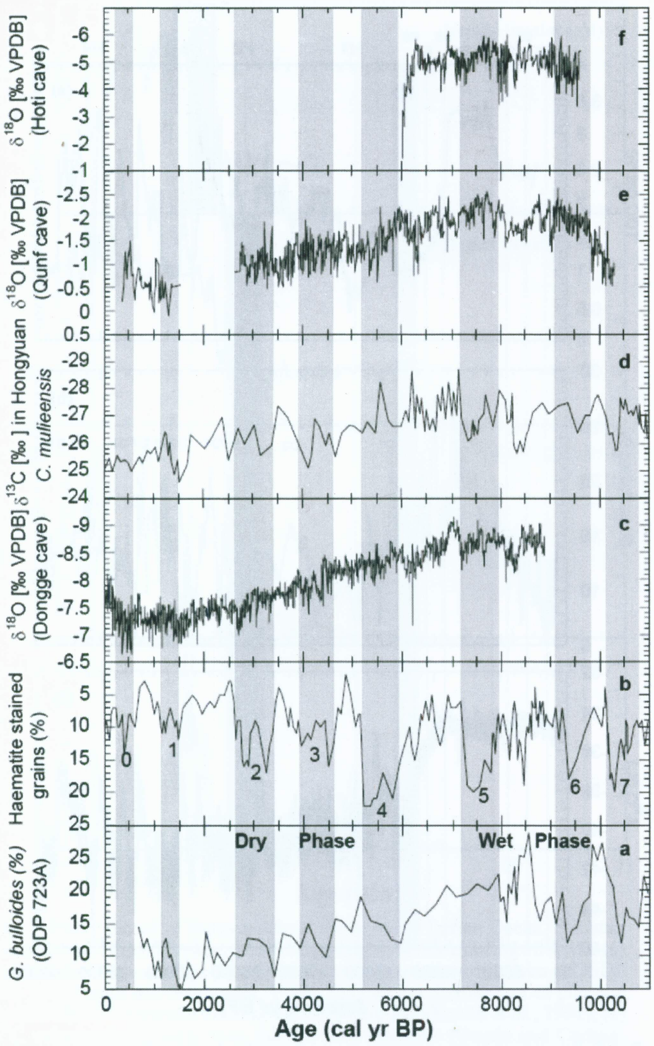

Fig. 6. Visual correlation between proxy records (reproduced from Gupta, 2008). **(a)** *G. bulloides* percent from ODP Hole 723A, Arabian Sea (Gupta *et al.*, 2003), **(b)** Percent haematite in core MC52 in the North Atlantic (Bond *et al.*, 2001), **(c)** oxygen isotope record from Dongge cave, southern China (Wang *et al.*, 2005), **(d)** Hongyuan peat deposit (Hong *et al.*, 2003), **(e)** Qunf cave ^{18}O (Fleitmann *et al.*, 2003), and **(f)** Hoti cave ^{18}O (Neff *et al.*, 2001). Grey bars highlight the Bond events 0-7 (Bond *et al.*, 2001).

nental vegetation to monsoonal intensification is not known, however, the marine records and evidence from paleosols and from palaeobotanical remains collectively provide firm support for major climate/vegetation changes driven by monsoon intensification during the latest Miocene. This shift in monsoon strength has been related to rapid uplift of the Himalaya and Tibetan plateau owing to near synchroneity of these disparate events (Quade *et al.*, 1989; Kroon *et al.*, 1991; Prell and Kutzbach, 1997). Zhisheng *et al.* (2001) related evolution of the Indian monsoon to phased uplift of the Himalaya-Tibetan plateau, whereas Gupta *et al.* (2004) suggested an alternative hypothesis linking late Miocene Indian monsoon shift to Himalayan-Tibetan uplift combined with Antarctic glaciation. The model experiments have shown that summer monsoon response is highly sensitive to the surface elevation of the Himalayan-Tibetan plateau and that half of the present - day elevation of this complex is required to produce the strong southwesterly winds over the Arabian Sea (Prell and Kutzbach, 1997). However, the link between monsoonal intensification and Himalayan-Tibetan uplift during 10-8 Ma, if such a link existed, is not straightforward since Himalaya-Tibetan complex underwent several phases of increase in elevation since the Oligocene (table 1). Thus, what we know today is the evidence of monsoon intensification during 10-8 Ma, and what we do not know is if 10-8 Ma monsoon intensification event was driven by the Himalayan-Tibetan uplift alone or it was a part of a global climatic change (e.g. Gupta *et al.*, 2004).

Plio-Pleistocene monsoon variability

During the Plio-Pleistocene, the monsoon entered a regime marked by major increase in the Northern Hemisphere Glaciation (NHG) superimposed by variations in Earth's orbital parameters (Gupta *et al.*, 2001; Gupta and Thomas, 2003). A change in monsoon seasonality towards intense winter monsoon is observed during 3-2.5 Ma in the eastern Indian Ocean coinciding with major expansion of the NHG, and increased aridity in eastern Africa (deMenocal, 1995; Gupta and Thomas, 2003). There is considerable evidence of linkage between the NHG and the development of, as well as fluctuations in, the monsoonal system. The initiation of the NHG was coeval with phased uplifts of the northern Tibetan Plateau (Zheng *et al.*, 2000), intensification of the winter monsoon, and decreased intensity of the summer monsoon (data on the Chinese loess deposits: Liu and Ding 1993; climate modeling: Prell and Kutzbach, 1997; Zhisheng *et al.*, 2001). An increased strength of and increasing seasonality in the winter monsoon at about 2.8-2.5 Ma was also seen in records from the Chinese loess deposits (Liu and Ding, 1993; Zhisheng *et al.*, 2001), where Monsoon system I (summer monsoon only) was replaced by Monsoon system II (summer and winter monsoons). The monsoon underwent more rapid changes during 3.6-2.6 Ma (Zhisheng *et al.*, 2001). Gupta and Thomas (2003) suggested that fluctuations in the strength of the Indian monsoon during the middle Pliocene might be linked to shifts in the

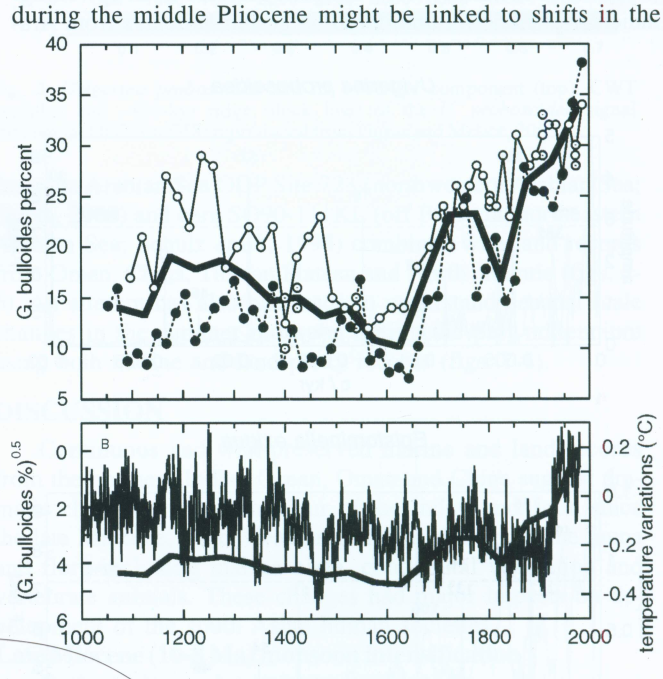

Fig. 7 (A) Time series of *G. bulloides* abundance (percent) from box cores RC2730 (solid circles) and RC2735 (open circles), and the composite record produced by averaging samples within 50-year intervals from 2000 to 1000 AD. (B) Time series of northern hemisphere temperature variations from Mann *et al.* (1999) superimposed on the index linearly related to monsoon wind speed, the square root of the difference in composite *G. bulloides* abundance with respect to the 1975 average (reproduced from Anderson *et al.*, 2002).

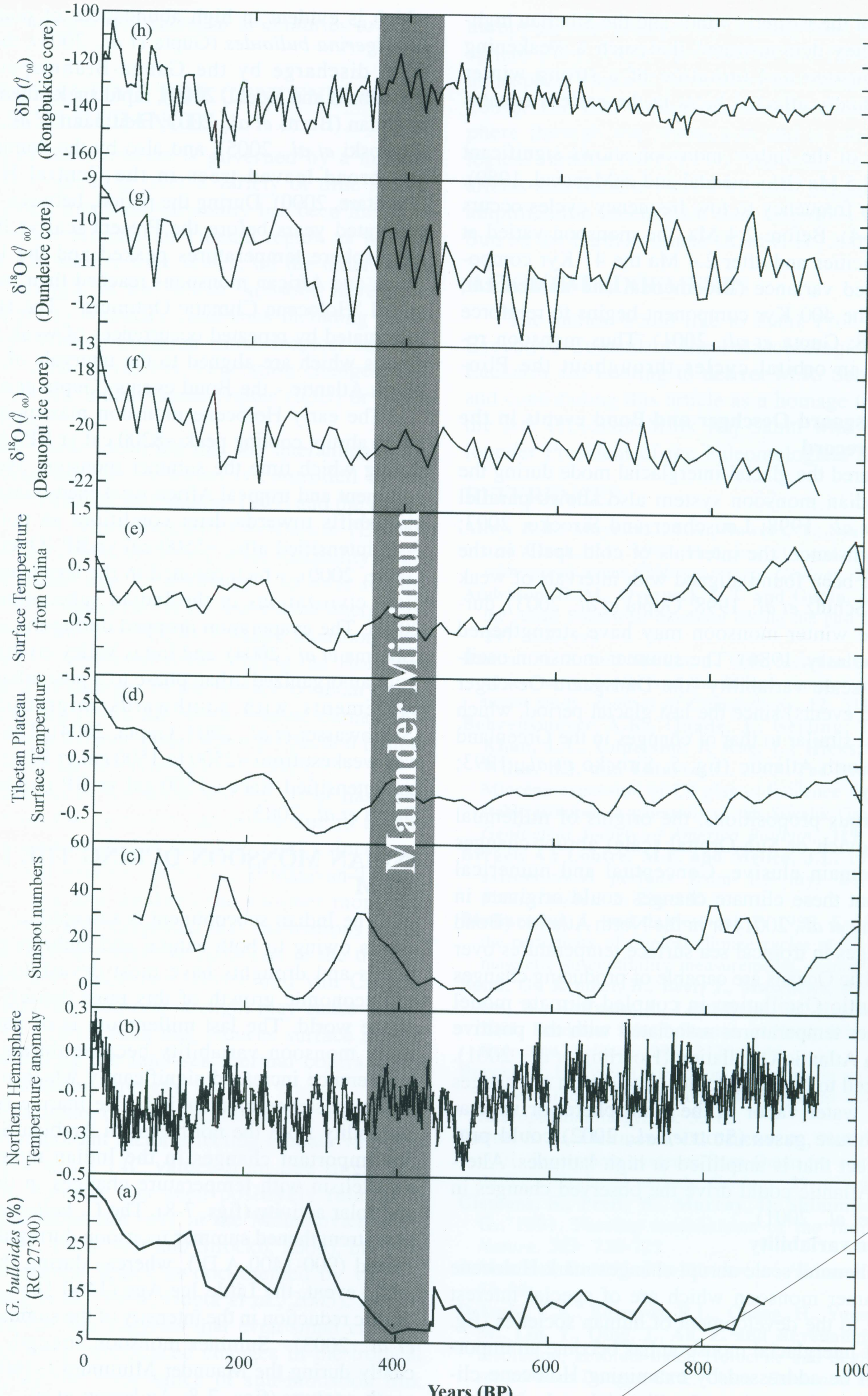

Fig. 8 (a) Percent *Globigerina bulloides* in core RC 2730 (Anderson *et al.*, 2002), (b) Northern Hemisphere temperature variations (Mann *et al.*, 1999), (c) Sun spot numbers (Solanki *et al.*, 2004), (d) Tibetan plateau surface temperature (Feng and Hu, 2005), (e) Surface temperature from China (Yang *et al.*, 2002), (f-g) ^{18}O variations in Dasuopu and Dunde ice cores from China (Thompson *et al.*, 2000), (h) ^{18}O variations in Rongbuk ice core from China (Hou *et al.*, 2007).

latitudinal position of the westerly winds and the Siberian high-pressure system. They demonstrated that such a weakening of the summer monsoon and initiation of a strong winter monsoon has profoundly affected biota in the eastern equatorial Indian Ocean.

Orbital forcing of the Indian monsoon shows significant shifts over the last 3.5 Ma (Bloemendal and deMenocal, 1989). A change from high frequency to low frequency cycles occurs at ~2.4 Ma (figs. 2-4). Before 2.4 Ma, the monsoon varied at precessional periodicities and after 2.4 Ma the 41 Kyr component shows increased variance (Bloemendal and deMenocal, 1989). At 0.9 Ma, the 400 Kyr component begins to reinforce (Berger *et al.*, 1998; Gupta *et al.*, 2001). Thus monsoon responded to shifts in orbital cycles throughout the Plio-Pleistocene.

Heinrich, Dansgaard-Oeschger and Bond events in the summer monsoon record

As climate entered the glacial-interglacial mode during the Quaternary, the Indian monsoon system also shows parallel changes (Schulz *et al.*, 1998; Leuschner and Sirocko, 2003; Gupta, 2008). For instance, the intervals of cold spells in the North Atlantic have been found aligned with intervals of weak summer monsoon (Schulz *et al.*, 1998; Gupta *et al.*, 2003), during which time the winter monsoon may have strengthened (Fontugne and Duplessy, 1986). The summer monsoon oscillated at millennial-scale variability (the Dansgaard-Oeschger (D-O) and Heinrich events) since the last glacial period, which has a pattern almost similar to that of changes in the Greenland and the northern North Atlantic (fig. 5; Sirocko *et al.*, 1993; Schulz *et al.*, 1998).

Despite numerous propositions, the origins of millennial scale (Sub-Milankovich or Sub-Orbital scale) abrupt changes in the monsoon remain elusive. Conceptual and numerical models indicate that these climate changes could originate in the tropics (Hoerling *et al.*, 2001) or in the North Atlantic (Bond *et al.*, 2001). Changes in tropical sea surface temperatures over the Indian and Pacific Oceans are capable of producing changes in the North Atlantic Oscillation in coupled climate model simulations, warmer temperatures associated with the positive phase of the North Atlantic Oscillation (Hoerling *et al.*, 2001). Other changes linked to the hydrological cycle such as changes in the quantity of water vapor in the atmosphere, or vegetation-driven greenhouse gases (Stott *et al.*, 2002) could produce a cooling effect that is amplified at high latitudes. Alternately, the North Atlantic could drive the observed changes in the tropics (Bond et al., 2001).

Holocene monsoon variability

Century to millennial scale abrupt changes mark Holocene record of the summer monsoon which are of special interest for their relevance to the development of human societies (fig. 6). The stability of interglacial monsoon has become an important issue that can be addressed by examining Holocene climate records. Recent observation of millennial scale variability during the last glacial, has been found to extend into the Holocene, challenging the hypothesis that stable conditions prevailed during the Holocene as observed in Greenland ice cores. The monsoon records from the Asian continent and the Indian Ocean document significant variability at centennial to millennial scales during the Holocene supporting the above argument (fig. 6; also Dykoski *et al.*, 2005).

The summer monsoon was intense in the early Holocene, which is evident in high abundances of planktic foraminifer *Globigerina bulloides* (Gupta *et al.*, 2003), the enormous sediment discharge by the Ganga-Brahmaputra river system (Goodbred and Kuehl, 2000), rapid speleothem growth reported in Oman (Burns *et al.*, 2001; Fleitmann *et al.*, 2003) and China (Dykoski *et al.*, 2005), and also by the dominance of conifers and broad-leaved trees in the Central Higher Himalaya (Phadtare, 2000). During the period, between ~10400 and 5500 calibrated years before the Present (cal yr BP), the Northern Hemisphere temperatures peaked and the Indian, Southeast Asian and African monsoons reached their maximum - the so-called "Holocene Climatic Optimum". The Holocene record is punctuated by repeated occurrences of weak summer monsoon phases which are aligned to the intervals of cold spells in the North Atlantic - the Bond events (Gupta *et al.*, 2003).

The early Holocene monsoon maximum was interrupted by an abrupt cooling peak ~8200 cal yr BP (Alley *et al.*, 1997), during which time the summer monsoon over the Indian subcontinent and tropical Africa weakened (Gasse, 2000). Successive shifts towards drier conditions in northern Africa and Asia intensified after ~5500 cal yr BP (Overpeck *et al.*, 1996; Gasse, 2000), which resulted in the termination/shifting of ancient civilizations in the Indian subcontinent (Gupta *et al.*, 2006). The evaporation dropped during the arid phase in India (Sharma *et al.*, 2004), and Indus Valley civilization transformed from an organized urban phase to a post urban phase of smaller settlements with southwards migration of population (Staubwasser *et al.*, 2003; Gupta, 2004). The summer monsoon was weakest from ~2500 to 1500 cal yr BP in the Holocene and has intensified since ~1500 cal yr BP (Anderson *et al.*, 2002; Gupta *et al.*, 2003).

INDIAN MONSOON DURING THE LAST MILLENNIUM

The Indian subcontinent is susceptible to extreme weather events owing to both natural and anthropogenic causes. Both floods and droughts have most devastating effect on health and economic growth of this most densely populated region of the world. The last millennium is of particular interest to study monsoon variability because during this time human interference increased significantly which may be detrimental to the existence of mankind. The marine record of monsoon variability from the northwestern Arabian Sea suggest numerous important changes in the Indian monsoon which show parallelism with temperature changes in the Tibetan plateau and solar activity (figs. 7-8). The *G. bulloides* percentages suggest strengthened summer monsoon during the Medieval Warm Period (900-1400 A.D.), whereas during the most recent climatic event, the Little Ice Age (1450-1850 A.D.), there was a drastic reduction in the intensity of the summer monsoon (Gupta *et al.*, 2003). Summer monsoon strength decreased significantly during the Maunder Minimum (~1600 A.D.) and eighteenth century (figs. 7-8; Anderson *et al.*, 2002), which coincide with devastating Indian droughts caused by summer monsoon failures. The eighteenth century summer monsoon failure coincides with repeated occurrences of El Niño-Southern Oscillation (ENSO) events (Trenberth and Hoar, 1996). The summer monsoon intensified since the Maunder Minimum paralleling an increase in surface temperatures in the Himalayan-Tibetan and Eurasian regions and increased sun spot numbers, indicating that as the Asian landmass will be hotter the summer monsoon strength will increase in centuries to come (figs. 7-8).

FORCING MECHANISMS FOR CENTENNIAL TO MILLENNIAL SCALE CHANGES

The strength of the monsoon is governed by a number forcing mechanisms operating over a variety of time scales. On longer time scales, monsoon variability has been linked to altitude changes in the Himalayan-Tibetan region as well as orbital variations. On shorter time scales, on the other hand, variations in the strength of the Indian monsoon have been explained by changes in boundary conditions including variations in Himalayan-Tibetan-Eurasian snow, tropical sea surface temperatures, solar variability, vegetation changes, and linkages with the ENSO, Indian Ocean Dipole (IOD) or North Atlantic Oscillations.

Previous studies have identified that the interannual variability of the Indian summer monsoon is controlled by the Eurasian snow cover (Bamzai and Shukla, 1999) and the amplitude and period of El Nino-Southern Oscillation (ENSO) (Krishna Kumar *et al.*, 1999). Millennial scale variability in the Indian summer monsoon may be attributed to solar forcing (Fleitmann *et al.*, 2003; Gupta *et al.*, 2005) and glacial-intergalcial boundary conditions (Burns *et al*, 2001). Major monsoon fluctuations in decadal to centennial scales during the Holocene have been attributed to changes in the surface boundary conditions (Overpeck *et al.*, 1996), the North Atlantic climate due to changes in the North Atlantic Deep Water production (Gupta *et al.*, 2003), and solar activity (Gupta *et al.*, 2005).

The extent/duration of the snow and ice cover has been suggested as a driver for monsoon variability due to albedo feedback of snow and ice-fields (Shukla, 1987). Increased albedo delays seasonal heating cycle over Himalayan-Tibetan region and thus delays and weakens the summer monsoon. Previous workers have also suggested that fluctuations in the monsoon intensity may be forced by variations in the tropical land-surface boundary conditions (Gasse and van Campo, 1994). Increased summer monsoon strength may cause denser vegetation cover in Central Asia that reduces surface albedo of the region, increasing the land-sea thermal contrast and thus the summer monsoon strength.

Stuiver and Braziunas (1993) suggested that solar variability could be an important factor affecting climate variation during the Holocene using 14C values from tree rings. The summer monsoon intensity has been positively related to changes in solar insolation not only in the Milankovich scale (Clemens *et al.*, 1991; Leuschner and Sirocko, 2003), but also in decadal and centennial scales as documented by various recent studies (Wang *et al.*, 1999; Gupta *et al.*, 2005). Recent study from the Arabian Sea suggests that even small changes in solar output can bring pronounced changes in the tropical climate, emphasizing the importance of sun-monsoon link (Gupta *et al.*, 2005). Gupta *et al.* (2005) found a correlation between increased sun spot numbers (Solanki *et al.*, 2004) and intense summer monsoon (high *G. bulloides* population) during the Holocene. The production rates of cosmogenic nuclides (14C and 10Be) that reflect changes in solar activity, are closely correlated to the Bond Cycles (Bond *et al.*, 2001).

Although the exact process by which global climate and solar forcing is linked is not clearly understood, variations in Ultra Violet radiation and cosmic ray flux (CRF) may trigger abrupt climate changes by altering heat budget of the stratosphere (Ney, 1959; van Geel et al., 1999) and changing the atmosphere's optical parameters and radiation balance (Kodera, 2004). The CRF affects electrical conductivity of the atmosphere through ion production and is an important meteorological variable (Ney, 1959). The solar-modulated CRF, which affects atmospheric electricity, may trigger a sufficiently large amplification mechanism that magnifies the influence of the Sun on the Earth's climate (Ram et al., 2009).

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