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Cenozoic Climatic Record for Monsoonal Rainfall over the Indian Region

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1. Introduction

Atmospheric carbon dioxide level is one of the major drivers responsible for the global temperature change (Lacis et al., 2010). The role of carbon dioxide as an important greenhouse gas, and its contribution towards regulation of global surface temperature has been recognized for over a century (Arrhenius, 1896; Chamberlin, 1899; Royer, 2006). The ice core records along with other proxy based records provides an evidence signifying a strong coupling between CO_2 and global temperature for at least the last ~65 m.y. (million years) (Petit et al., 1999; Siegenthaler et al., 2005, Zachos et al., 2001). The intensification of convective hydrological cycle inducing heavy rainfall during high pCO_2 condition is both simulated and estimated from General Circulation Models (GCM) and geochemical analyses of fossil record respectively (Kutzbach & Gallimore, 1989). The evidences of intensification of monsoon, which refer to the rainfall due to seasonal reversal of the wind direction along the shore of the Indian Ocean especially in the Arabian Sea and surrounding regions, are preserved in the sedimentary records from continental and oceanic region (Fig.1). The other factor which affected the regional hydrological cycle apart from the concentration of CO_2 in the atmosphere is tectonic rise of Himalayan mountain. Proxy record based on parameters like stomata index, alkenones and boron isotopes clearly suggested high concentration of CO_2 in the atmosphere (~400 ppm) during Miocene time. The estimated concentration of CO_2 observed in the atmosphere was rather similar to the concentration of CO_2 in the atmosphere measured in the recent years at Mauna Loa (Thoning et al., 1989). The effect of such high CO_2 concentration is seen to have significantly modulated and altered the pattern of rainfall distribution, intensity and its spatial variability. Record from sedimentary archives from the continental and marine sites over the Indian region yielded evidence of warmer, wetter and higher temperature seasonal climate for the Miocene period. A similarity of signature both from continental region and the marine archives support the argument for the change in hydrological condition during last 20 m.y. The marine records are only a few but the largely scattered along the continental margin and central Indian Ocean. A more recent study of such sedimentary sequences lying on the western and eastern India provided glimpses of spatial variability of regional climate. The chapter will narrate the long term variation in Miocene monsoonal rainfall and its spatial pattern using large set of available observations from the palaeo record. This narration will include

- Large synthesis of proxy based parameter from of geological, palentological, biogeochemical and isotopic proxies.

- Implication of upliftment rate of Himalaya and formation of Tibetan Plateau on monsoonal intensity.

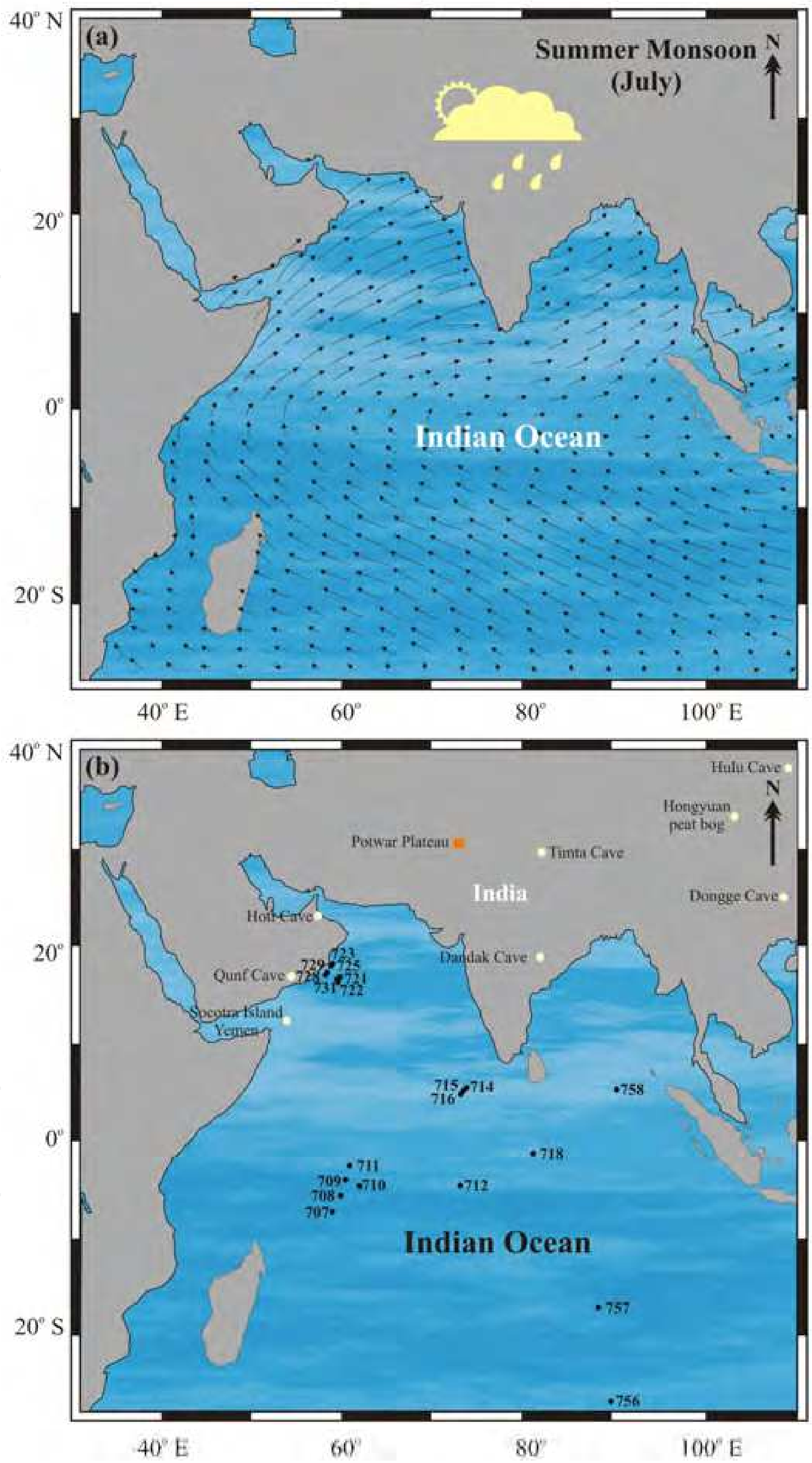

Fig. 1. (a) Climatology of the summer monsoon circulation (July) marked by surface winds (Figure after Wang et al., 2005) and (b) Indian Ocean along with Ocean Drilling Program (ODP) Holes and Cave deposits from south Asia.

The scenario is complex but possible to model with simultaneous observation on rate of upliftment of Tibetan Plateau and pCO_2 level both playing role in influencing the land-sea thermal contrast and strengthening of wind (Hahn & Manabe, 1975). The dual reason provided here will cause strong intensification of the Indian monsoons, with large increment in the summer precipitation (summer monsoon) over the Indian subcontinent and a cold dry winter season (Ruddiman & Kutzbach, 1989; Hastenrath, 1991). According to the tectonic upliftment hypothesis, monsoonal intensification due to changes in the elevation of the Himalayan-Tibetan region had greater role driving the climate in Indian region (Molnar et al., 1993; Clift et al., 2002).

2. Paleoclimate proxies

The evidence of the monsoon rainfall is registered in variety of proxies both over land and oceanic region, arguably the most reliable continuous record of the monsoonal variability comes from the sediments deposited on the ocean floor, from the Arabian Sea region. We cannot directly observe the climates of the geological past in the way that we do record climate parameters since the modern centuries with instrumental record (Plaut et al., 1995). The key variables measured are temperature and humidity. The indirect mean to retrieve such information from sedimentary archives is through analyses of physical and chemical parameter in the sedimentary deposits and these includes physical proxies like clay mineralogy, grain size distribution, heavy minerals, organic bio markers, chemical and isotopic proxies like isotopes in ice core, alkenones, sedimentary organic fraction, inorganic carbonates and organic deposits. Isotopic compositions provide direct relationship allowing understanding of temperature, atmospheric composition and seawater salinity respectively. Along with this multi parameter approach, the importance of distribution of microfossil assemblages and abundances of particular species from marine platform provide direct measure on the response of physical and chemical factors on the biota.

2.1 Paleontological proxies

The empirical use of fossils and fossil assemblages as palaeoclimate indicators can be traced back at least as far as Lyell (1830) and their value was even recognised in classical biostratigraphic literatures (Imbrie & Newell, 1964). The oceanic sedimentary records have been successfully used for reconstruction of monsoon intensity through time (Kroon et al., 1991; Nigrini, 1991; Prell et al., 1992; Prell & Kutzbach, 1992). The biological, chemical and sedimentological parameters indentified as important for understanding the intensity of upwelling phenomena having direct linkage with the strength of the monsoonal winds. Monsoonal variability can therefore be constrained in terms of wind strength through time. The intense seasonal upwelling is induced by the south-westerly monsoon winds from Arabian Sea (Anderson & Prell, 1993). Sediments in the Arabian Sea exhibit a characteristic fauna (foraminiferal species) that are endemic to areas of upwelling. Majority of these species are encountered only in cool temperate water, and therefore, their appearance and abundance in sediment record indicate parameter for upwelling (index). Cullen and Prell (1984) provided the high-quality core-top data banks to indentify, establish and evaluate proxies for monsoonal upwelling. The northern Indian Ocean based on the analysis of 251 core-top samples shows the distribution of planktic foraminiferal dwelling on the surface water and provide tool for paleoceanographic reconstructions. Together with isotopic ratios and trace element concentration, the assemblage of foraminiferal species capture changes in surface water hydrographic condition (e.g. Wefer et al., 1999; Thunell & Sautter, 1992; Kennett & Srinivasan, 1983; Kennett, et al., 1985). The oxygen and carbon isotope composition in shells of planktic foraminifera ($\delta^{18}O_{shell}$, $\delta^{13}C_{shell}$) provides one of the most widely used tools for reconstructing past changes in sea surface temperature and salinity (King & Howard, 2005). Also, high concentration of certain species like *Globigerina bulloides* mirror upwelling conditions in the tropical ocean (Prell et al., 1992; Overpeck et al., 1996; Peeters et al., 2002; Gupta et al., 2003). The $\delta^{18}O$ of planktic foraminiferal calcite records enable reconstruction of global surface temperatures, however suffers from effects like depth habitat reorganization due to climatic evolution (MacLeod et al., 2005). *Neogloboquadrina dutertri* is a significant component of the planktonic foraminiferal group in the Arabian Sea upwelling area (Be & Hutson, 1977; Kroon, 1988) associated with upwelling phenomena together with *G. bulloides* (Auras et al., 1989). The Globigerinoides species (*Globigerinoides sacculifer, Globigerinoides ruber, etc.,*) have occurrence subtropical- tropical surface waters with an oligotrophic character (Be & Hutson, 1977). Intervals with higher frequencies of Globigerinoides are characterized by surface waters condition different from the upwelling situations (Kroon et al., 1991). Paleotemperature reconstruction using benthic foramifera from different sites (Pacific, Atlantic and Indian Ocean) was achieved measuring parameter like $\delta^{18}O$ reflecting either temperature or ice volume effect (Zachos et al., 2001). Benthic foraminifera have ability to adapt changing environmental conditions, enabling them to survive in a wide range of marine environments. Their calcium carbonate shells represent a major and globally significant sink for carbon. Also their calcareous or agglutinated tests lend them good fossilization potential. Thus benthic foraminifera provide important tool for reconstruction of deep-sea paleoceanography and paleoclimatology, based on observations from modern day environmental niche (Gupta & Srinivasan, 1992; Sen Gupta & Machain-Castillo, 1993; Wells et al., 1994; Thomas et al., 1995; Gupta & Thomas, 1999, 2003).

2.2 Clay mineralogy

Composition of clay minerals in marine and terrestrial record reflects weathering processes in the catchment area and basin, adjoining depositional setup (Chamley, 1989; Sellwood et al., 1993; Chamley, 1998; Thiry, 2000). However, for example, the study conducted on the marine sedimentary record from southern latitudes documenting trend in the abundances of smectite, chlorite, and illite. A declining proportion of crystallized smectite and chlorite, and increasing illite in sediments was used to trace the transition from humid to sub-polar and polar conditions (Ehrmann et al., 2005). Ballantyne (1994) and Ballantyne et al. (2006) suggested that low gibbsite concentration in soils deposited in glacial terrain, which can also be used for dating the deposit involving cosmogenic isotopes.

2.3 Palaeosols

Stable isotopic (oxygen and carbon) ratios in paleosol carbonate nodules (Ghosh et al., 2004) and associated organic matter have been used as a proxy for vegetation and climatic changes. While oxygen isotopic composition of carbonate nodules act as an indicator of temperature and salinity condition of environmental water, carbon isotopic composition signifies presence of C_3, C_4 and CAM vegetation types (Cerling et al., 1989). Sellwood et al. (1993) and Retallack (2001) suggested palaeosols are one of the important palaeoclimatic proxy, in particular to estimate the palaeo-precipitation. Presence of palaeosols in a sedimentary record represents intervals with low or no sedimentation. Their mineral and chemical composition reflects the interaction between their source terrigenous clastic sediments and the processes of weathering, which can be physical, chemical and biological.

2.4 pH

The $\delta^{11}B$ of marine carbonate analyzed by Pearson and Palmer (1999) and Pagani et al. (2005a) to deduce pCO_2 estimation relies on the fact that a rise in the atmospheric concentration will mean that more CO_2 is dissolved in the surface ocean, causing a reduction in ocean pH. Sanyal et al. (1995) suggested the oceanic pH provides insights into how the carbonate chemistry of the oceans, including depth to lysocline, has changed through time. Royer et al. (2004) examined palaeo-pH of entire Phanerozoic period has also been estimated based on the calcium-ion concentration of seawater and modeled atmospheric CO_2 concentrations.

2.5 Atmospheric CO_2

Ice core data provided the atmospheric CO_2 concentrations that pre-date ~800 k.y. direct record (Siegenthaler et al., 2005) are of extreme importance for palaeoenvironmental reconstruction (Vaughan, 2007). Pagani (2002) suggested the oceanic proxies, $\delta^{13}C$ of organic materials has been particularly successful in pCO_2 determination, and has recently been refined with studies that focus on $\delta^{13}C$ derived from a single group of long chain hydrocarbons, for example, alkenones and rather than total marine organic material (Henderson, 2002). Pagani et al. (2005b) provided the alkenones record extends back to the Eocene–Oligocene boundary. $\delta^{13}C$ of carbonate, including both marine (Buggisch, 1991), freshwater (Yemane & Kelts, 1996) and pedogenic (Royer, 2001), has also been used as a proxy for atmospheric and oceanic carbon source and for rates of carbon burial (Schouten et al. 2000; Strauss & Peters-Kottig, 2003). Henderson (2002) and Pearson & Palmer (2002) recorded the oceanic pH has also been used as a proxy for atmospheric CO_2 concentration, although this requires assumptions to be made about the equilibrium condition between dissolved inorganic carbon and atmospheric CO_2. By this method, they have reconstructed atmospheric CO_2 concentration back to 60 m.y. ago (Pearson & Palmer, 2000; Demicco et al. 2003). Kasemann et al. (2005) recently extended the technique to deduce CO_2 from Neoproterozoic carbonates. Also GCM studies of monsoon climates have progressed extensively and contributed to our understanding of how the monsoon system evolved to various large scales forcing, including tectonics, CO_2, and orbital variations in incident solar radiation (Kutzbach et al., 1993; Wright et al., 1993, Wang et al., 2005). Atmospheric pCO_2 determination using parameters like stomatal index and carbon isotopic composition of nodules within soil carbonates provides efficient tool to estimate concentration of CO_2 in the atmosphere of geological past (Cerling, 1991; Yapp & Poths, 1992; van der Burgh et al., 1993; McElwain & Chaloner, 1995).

Chemistry-based proxies are the most direct measure and tend to focus on the specific chemical or isotopic system. The proxies are grouped by parameter estimated. Lithological proxies are directly related to the parameter estimated and are grouped by rock type, mineralogy or facies interpreted. Palaeontological proxies are subdivided by whether they use taxonomic methods or focus on some morphological aspect of a group of organisms. There are several proxies which were introduced over the past two decades (Royer et al., 2001, 2006); the $\delta^{13}C$ of pedogenic minerals (Cerling, 1991; Yapp & Poths, 1992); the stomatal densities and indices in plants (van der Burgh et al., 1993; McElwain & Chaloner, 1995); the $\delta^{13}C$ of long-chained alkenones in haptophytic algae (Pagani et al., 1999); the $\delta^{11}B$ of marine carbonate (Pearson & Palmer, 1999; Pagani et al., 2005b); and the $\delta^{13}C$ of liverworts (Fletcher et al., 2005). Some of the proxy records including soil carbonates (Quade et al., 1989; Sanyal et al., 2004), palaeosols (Rettallack, 1995; Thomas et al., 2002, Ghosh et al., 2004), microfossils (Phadtare et al., 1994), pollens (Hoorn et al., 2000), palaeomagnetic record (Sangode & Bloemendal, 2004) and general sedimentation parameter where precipitation pattern have been used to decipher the changes in the Indian monsoon strength. In the following sections we have provided overviews of monsoon variability.

3. Debate on impact of Cenozoic CO_2 and tectonic process on monsoon

The role of Cenozoic pCO_2 and its effect on monsoonal intensity, including rapid change in the precipitation cycles was discussed in the community through proxy based studies (McGowran, 1989; Brady, 1991; Worsley et al., 1994; Kerrick & Calderira, 1998; Beck et al., 1998; Pearson & Palmer, 2000). During this debate some researcher have stressed the importance of the changing inputs to the atmosphere such as volcanic and hydrothermal outgassing (Owen & Rea, 1985; Berner et al., 1993; Ghosh et al., 2001) or metamorphic decarbonation reactions (Kerrick & Calderira, 1998), while others have focused on outputs such as weathering of silicate minerals and limestone formation (Brady, 1991; Raymo & Ruddiman, 1992; Worsley et al., 1994) or organic carbon burial (Berger & Vincent, 1986; McGowran, 1989; Beck et al., 1998). Cenozoic time CO_2 concentration has gradually decreased from >1000 ppmv during the Paleocene and the beginning of the Eocene period to <300 ppmv during the Pleistocene. This long-term decrease is partly due to reduction in volcanic emissions, which were particularly large during the Paleocene and Eocene epochs, but which have diminished since then, and to changes in the rate of weathering of silicate rocks. Figure 2 shows the sharp decline in the CO_2 concentration is associated with a cooling from the warm conditions of the early Eocene climatic optimum between ~52-50 Ma. This shift is often referred to as a transition from a greenhouse climate to an icehouse, in which permanent ice sheets formation were initialized over Antarctica (starting ~35 Ma) and Arctic regions (around ~3 Ma at Greenland). The warming during the Paleocene Eocene Thermal Maximum (PETM) around ~55 Ma ago, which also had a major impact on life on Earth, is better documented (Fig. 2). During this event which lasted less than ~1.7 Ma, the global temperature increased by more than 5°C in less than 10,000 years. This period is also characterised by a massive injection of carbon into the atmosphere-ocean system as recorded by the variations in the $\delta^{13}C$ measured in the sediments. An apparent discrepancy between oxygen isotope data and other paleoclimate proxies for the span from ~26-16 Ma was resolved by comparison against eustatic estimates from continental margin stratigraphy. Ice-volume estimates from oxygen isotope data compare favorably with stratigraphic and palynological data from Antarctica, and with estimates of atmospheric carbon dioxide for the early Oligocene through early Miocene (34-16 Ma). These isotopic data suggest that the East Antarctic Ice Sheet grew to as much as 30% greater than the present-day ice volume at glacial maxima. This conclusion is supported by data on seismic reflection and stratigraphic data from the Antarctic margin that suggest that the ice sheet may have covered much of the continental shelf region at Oligocene and early Miocene glacial maxima. Palynological data suggest long-term cooling during the Oligocene, with near tundra environments developing

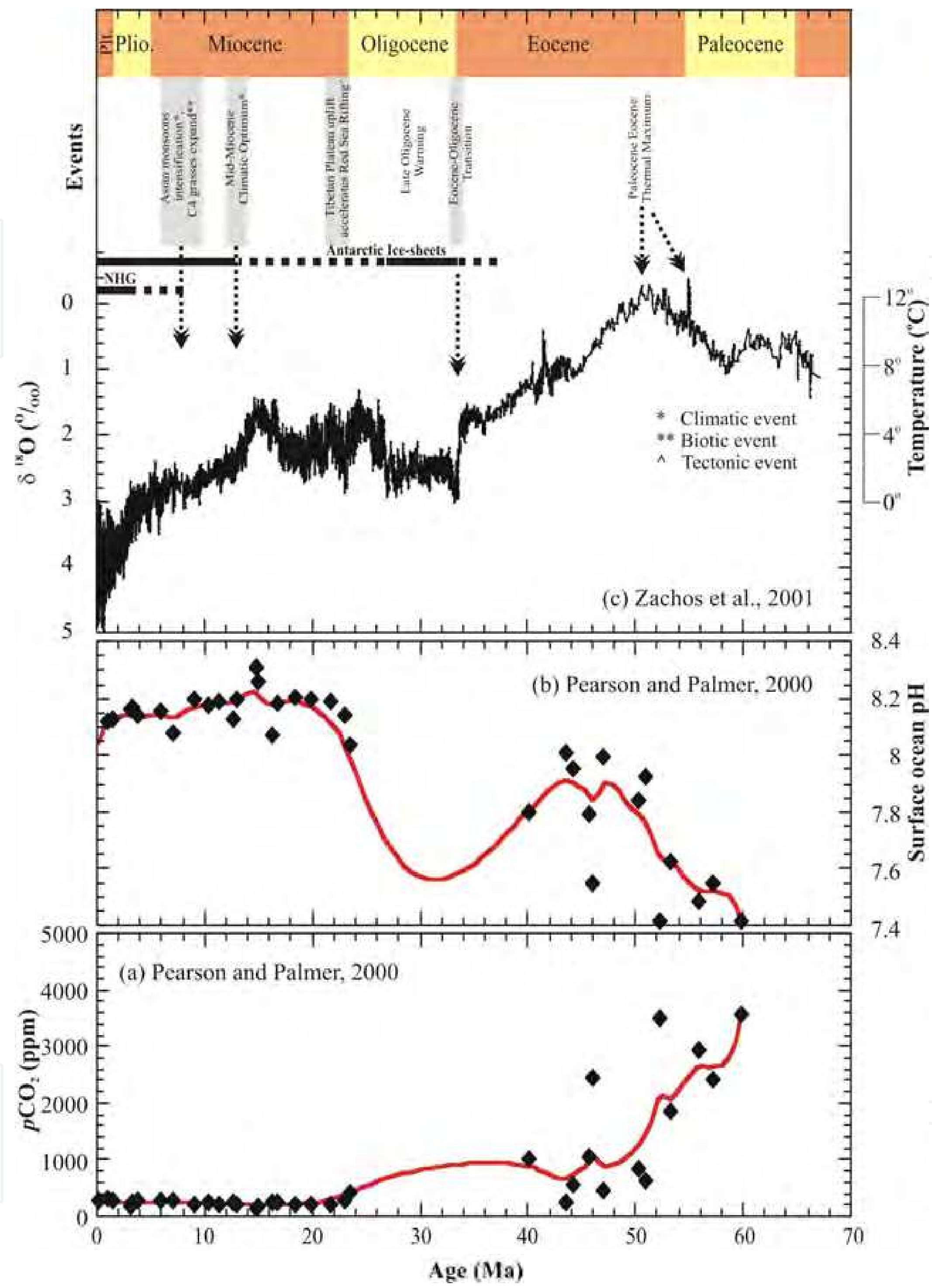

Fig. 2. The global climate over the past ~65 million years based on deep-sea oxygen-isotope measurements in the shell of benthic foraminifera. The $\delta^{18}O$ temperature scale, on the right axis, is only valid for an ice-free ocean. It therefore applies only to the time preceding the onset of large-scale glaciation in Antarctica (about 35 million years ago, see inset in the upper left corner), Figure from Zachos et al. (2008). Estimates of Sea surface pH and pCO_2 levels in Cenozoic climate change. The cure spanning most of the Cenozoic is estimated from the surface ocean pH as derived from the $\delta^{11}B$ ratio of planktonic foraminifers (Pearson & Palmer, 2000).

along the coast at glacial minima by the late Oligocene. Crowley (2000) has reported concentration of CO_2 changes in the atmosphere commonly regarded as a likely forcing mechanism on global climate over geological time scale because of its large and predictable effect on global temperature. An important factor of tectonic forcing associated with the reorganization of topography and elevation of the mountainous regions directly affect the atmosphere circulation and the wind prevailing over the region. Tectonic reorganizations by themselves were not the dominant cause of the warm climates of the Cenozoic, but they affected climate by controlling the processes that control ocean circulation, transport or trap solar heat, and maintain greenhouse gas levels in the atmosphere (Lyle et al., 2008). There was considerable fluctuation in these variables in the Eocene and Oligocene time period, but since the earliest Miocene the system seems to have been much steadier and more closely comparable to the present, despite continuing climate cooling. This suggests that other factors, such as complex feedbacks initiated by tectonic alteration of the ocean basins, were also important in determining global climate change (Pearson & Palmer, 2000).

3.1 Linkage between Indian monsoon system and uplift of the Tibetan Plateau

The major causes for the origin and intensification of the Indian monsoon system can be understood based on the upliftment rates recorded the Himalayas and Tibetan Plateau, because mountains modulate the land-sea thermal gradient. The growth of the Tibetan Plateau has been cited as being a triggering mechanism for a much stronger summer monsoon than might otherwise be predicted (Ruddiman & Kutzbach, 1989; Molner et al., 1993) and areas of rapid exhumation in the Himalayas have been correlated with zones of the most intense of summer monsoon (Thiede et al., 2004). The consequence of this was erosion which not only affected the mountain topography, but its transported large volumes of the sediments to the ocean. Due to this chemical reaction between silicate and acidic water was held responsible for the consumption of the atmospheric CO_2, a greenhouse gas, which in turn can drive long-term global cooling (Raymo & Ruddiman, 1992). Unfortunately, the timing of the tectonic events is poorly constrained relative to the timing of climate changes (Gupta et al., 2004) and Changes in continental position and height have frequently been invoked as causes of large-scale climatic shift during Cenozoic (Lyle et al., 2008). Data compilation of Aeolian records spanning early Miocene to late Pleistocene (central China) were used to infer aridity of the mid-continent was caused by global cooling or topographic uplift of the Tibetan Plateau, the latter might caused a rain shadow effect (Lu et al., 2004). A close association of drying proxies with global cooling and suggest a reduced role for topographic growth on climate change (Lu et al., 2004). The time between Miocene to recent was examined by Zheng et al. (2004), where sediments along the northern edge of the Tibetan Plateau in the Tarim Basin were investigated. The sections dated as ~8 Ma at the base grade upwards to a finer grade and more distal clastic sediments occurrences in a coarse alluvial fan deposits related to the uplift of northern Tibet were documented. At the same time aeolian dunes formation and Playa lakes started form after ~8 Ma, suggesting an enclosed desert basin from that time, suggesting that plateau uplift may be causing the regional climatic drying (Zheng et al. 2004). Gupta et al. (2004) examined the deep-sea benthic foraminiferal census data combined with published data shows that a major increase in biogenic productivity at ~10–8 Ma throughout the Indian Ocean, the equatorial Pacific, and southern Atlantic. The authors suggest that a change has been linked to initiation or intensification of the Indian Ocean monsoons. They argued, however, that the oceanic productivity was changed in a larger region than that affected by the monsoons, and secondly the effects of mountain uplift on global climate are not well documented (Hay et al., 2002). More recent workers have tried to estimates the rate of upliftment of mountains during Cenozoic time using soil carbonates, clays and organic component in the sedimentary deposits (Garzione et al., 2004; Chamberlin, 2006). Hoang et al.(2010) in the Gulf of Tonkin in the South China Sea, their study was based on a seismic stratigraphic analysis of sediments in the Song Hong-Yinggehai Basin formed as a pull-apart basin at the southern end of the continental Red river fault zone. During ~4-3 Ma the basin was filled by the sediments transported from the Red river, allowing periods of faster erosion and between ~15 and 10 Ma to be identified and linked to times of strong summer monsoon. However, the earliest pulse of fast erosion at 29.5–21 Ma may have been triggered by tectonic rock uplift along the Red river fault zone. The chemical weathering has gradually decreased in SE Asia after ~25 Ma, probably because of global cooling, whereas physical erosion became stronger, especially after ~12 Ma (Hoang et al., 2010). Kitoh, (2004) also studied the impact of Tibetan plateau uplift using by a GCM. Their model suggests that when there is no plateau present monsoon precipitation is limited in the deep tropics during northern hemispheric summer. However, as the Tibetan Plateau was uplifted, rainfall increases and affected climate over inland region over the south-eastern Tibetan Plateau. The Indian monsoon system got more intensified with upliftment of Tibet which commenced ~15 Ma. A rapid rate of upliftment during 10-8 Ma (1000-2000 m), coincides with the period of climate change or monsoon intensification (Harrison et al., 1992; Prell et al., 1992; Molnar et al., 1993). The Himalayas associated with Tibetan uplift and climate change connection is the classic example demonstrating a linkage between mountain growth and climate, yet despite significant attention and study a consensus history of the growth of the Himalayas and associated Tibetan Plateau has been elusive (Molnar et al., 1993; Copeland, 1997; Lyle et al., 2008).

4. Cenozoic climate record for estimated correlations: CO_2 and temperature (Greenhouse climates)

The link between the level of atmospheric CO_2 and global surface temperature is intensely important to understand. Solomon et al. (2007) suggested that implicated as the predominant cause of recent global climate change is release of fossil fuel CO_2 to the atmosphere by human activity. Zachos et al. (2001) has developed another database of $\delta^{18}O$ isotopes that cover the past 65 m.y., the time period when Antarctica glaciated over is clearly evident and based on which he showed that the trend in the $\delta^{18}O$ record suggesting a increasing and decreasing features reflect periods of global warming and cooling, and ice sheet growth and decay (Fig. 2). During Cenozoic period around ~59 Ma and 52 Ma (mid-Paleocene to early Eocene), the most distinct warming epoch, as expressed by enriched $\delta^{18}O$ composition was peaked at the PETM at ~52-50 Ma. The global average temperature exceeded by 6°C during the PETM and the periods witnessed deglaciation of Antarctica and Greenland glaciers started rebuilding at ~14 Ma. Zachos et al. (2001) showed that although having a high CO_2 concentration and high temperature, the Paleocene to early Eocene temperature record is a climate puzzle: globally averaged surface temperatures were significantly warmer than the present day and there is no convincing evidence for ice

(Frakes et al., 1992), but CO_2 estimates range from <300 to >2000 ppm (Fig. 3). Over the last 65 m.y., the CO_2 concentration has gradually decreased from more than 1000 ppmv during the Paleocene and the beginning of the Eocene epochs to less than 300 ppmv during the Pleistocene and the Cenozoic climate trend is characterized by a deep-sea cooling of approximately 12°C thought to have been forced by changes in atmospheric greenhouse gas composition (Hansen et al., 2008; Beerling & Royer, 2011). As suggested by Zachos et al. (1993), the transient climatic episode in this warm period, the PETM, at ~55.8 Ma and 52 Ma, may at face value seem an odd choice, given that it represents a period of extreme warmth. The Cenozoic warmth ~52 Ma corresponds with maximum reconstructed CO_2, and the rapid initiation of Antarctic glaciation to the Eocene-Oligocene boundary during ~34 Ma follows sharp fall in CO_2 (Royer, 2006). Crowley and Berner, (2001) and Royer et al. (2004&2006) reported clearest example in the Phanerozoic of a long-term positive coupling between CO_2 and temperature during the late Eocene to present day. The major changes in the Cenozoic earth's climate from a relatively warm and equable climate in the Paleocene to cold conditions with nearly freezing temperatures at the poles in the Pliocene, an evidence from various proxy record (Kennet & Barker, 1990; Lear et al., 2003; Zachos et al., 2001; Singh & Gupta, 2005), and like e.g. $\delta^{18}O$ for palaeotemperature (Zachos et al. 2001; Royer et al. 2004), $\delta^{11}B$ for palaeo-pH and palaeo-CO_2 (Pearson & Palmer, 2000), Sr/Ca from benthic foraminiferal calcite for weathering fluxes (Lear et al., 2003), etc. The late Oligocene and early Miocene an interval firmly linked with global warming and/or decreased ice mass in Antarctica (low $\delta^{18}O$ values; Miller et al., 1999 & 2004; Zachos et al., 2001; Barker & Thomas, 2004). The other section will narrate the vagaries of monsoon during last million years (glacial-interglacial cycles) and during last glacial maximum (LGM). A review of estimates of palaeo-atmospheric CO_2 levels from geochemical models, palaeosols, algae and foraminifera, plant stomata and boron isotopes concluded that there is no evidence that concentrations were ever more than 7500 ppm or less than 100 ppm during the past 300 m.y. (Crowley & Berner, 2001). Tripati et al. (2005) provide a strong, although indirect, case for three short lived glaciations at 42 – 41 Ma, 39 – 38 Ma, and 36.5 – 36 Ma based on independent reconstructions of sea level, temperature, and calcite compensation depth. The CO_2 record indicates moderately high CO_2 levels (1200 ppm) at 44 Ma, dropping to low levels (<500 ppm) just before the onset of the first cool event at 42 Ma (Fig. 3). This pattern of CO_2 (<1000 ppm during cool events and >1000 ppm elsewhere) generally persists across the remaining two events. Pearson and Palmer (2000) reported during Cenozoic pCO_2, indicating that this termination occurred at a time when green house gas levels were decreased or already low (Fig.2). ~25 Ma onwards the pCO_2 shows continuously decreasing trend due to the tectonic events such as mountain building or oceanic gateway reconfigurations, which can alter atmospheric circulation and water vapour transport, may have had a dominant role in triggering large scale shifts in climate as well (Kuztbach et al., 1993; Mikolajewicz et al., 1993; Driscoll & Haug, 1998, Zachol et al., 2001). Neftel et al. (1988) have been reported the atmospheric CO_2 concentration about 30% less than Holocene pre-industrial value during the last glaciation, although this change is thought to originate from ocean (Broecker & Pang, 1987; Archer & Maier-Reimer, 1994). The Cenozoic that represent examples of unique climate states of the Earth: the warm Eocene greenhouse world, which is associated with elevated atmospheric CO_2; the Miocene, when globally warm temperatures persisted but without the aid of high atmospheric CO_2 concentrations; and the Pliocene, a warm period that gradually cooled to become the Pleistocene ice ages (Lyle et al., 2008).

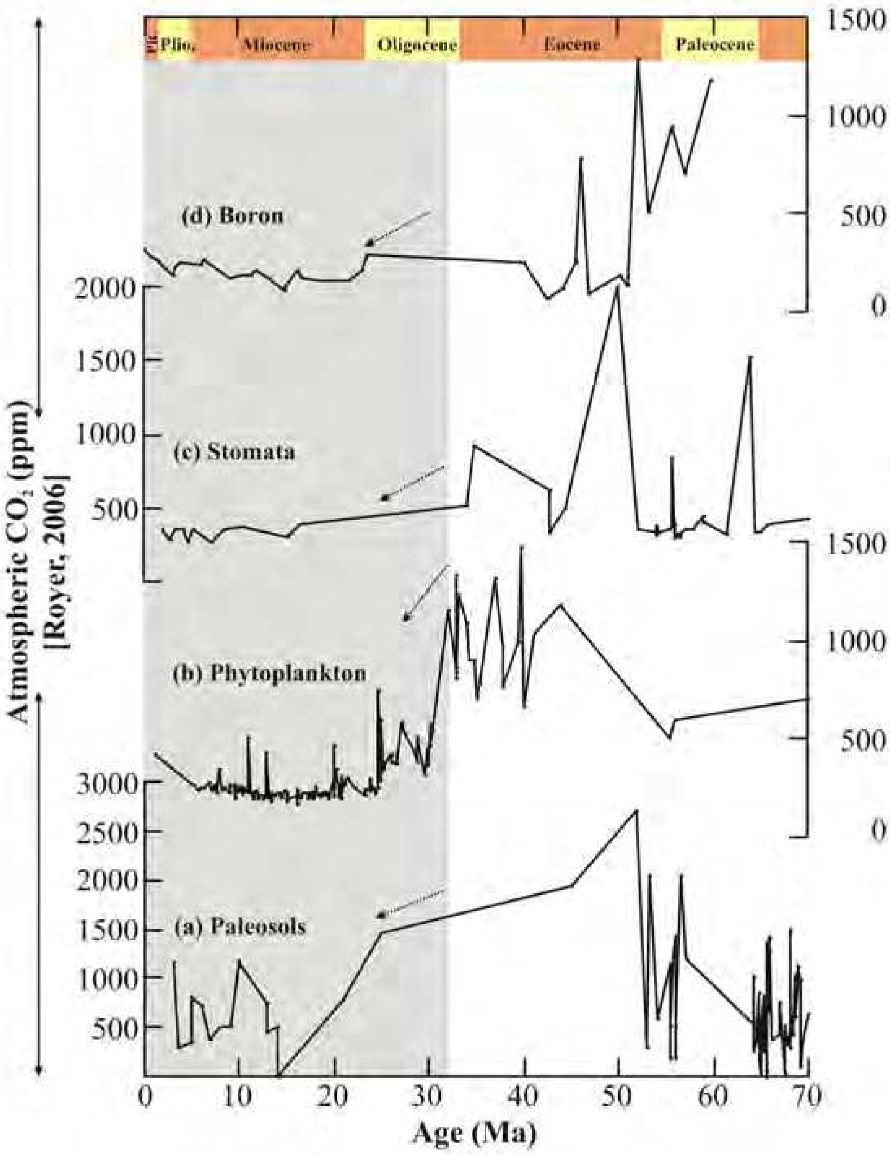

Fig. 3. Earth's atmospheric CO_2 history through the Cenozoic period. Various proxy generally track the estimates of atmospheric CO_2 (given separate panels a-d) reconstructed from the terrestrial and marine proxies following recent revisions (Royer, 2006). The gray zone shows from ~32 Ma to Holocene period, at 1000 and 500 ppm CO_2 represents the proposed CO_2 thresholds for, respectively, the initiation of globally cool events and full glacial (Royel, 2006). Early results provide increasing atmospheric CO_2 reconstructions.

5. Indian monsoon system and Cenozoic

The name "Monsoon" is derived from the Arabic word "mausim" meaning "season". Seasonal changes in the wind direction dominate the tropical climate of the Indian monsoon, bringing significant changes in ocean chemistry, biota and biogenic sedimentation budget. The seasonal reversal in the northern Indian Ocean wind system, called the Indian monsoon, is an important component of the global climate system affecting the weather and climate of the Asian and African regions between 30°N and 20°S latitudes (Fig. 1). The Indian monsoon is of great socio-economic importance, affecting the lives and livelihood of over 60 percentages of the world's population. In the Northern Hemisphere, during the summer monsoon, winds are southwesterly over the Arabian Sea and Bay of Bengal, whereas, during winter the monsoonal winds are directed away from the Asian continent, causing northeasterly wind transport over the area (Schott & McCreary, 2001). Several factors like the seasonal distribution of solar insolation, global ice volume, sea surface temperature, albedo, Himalayan-Tibetan heat budget, El Nino-Southern Oscillation (ENSO) and variability in the Indonesian Throughflow (ITF) may influence the monsoonal strength (Prell, 1984; Kutzbach, 1981; Quade et al., 1989; Clemens & Prell, 1990; Raymo & Ruddiman, 1992; Harrison et al., 1993; Kutzbach et al., 1993; Molnar et al., 1993; Raymo, 1994; Li, 1996; Gordon et al., 1997; An et al., 2001; Gupta et al., 2003, 2004). The monthly mean wind field for July (summer monsoon) are shown in Figure 1a., which induces ocean upwelling due to process of Ekman transport. During the summer monsoon season, the upwelling regions along the coasts of Somalia and Oman along with the southern coast of India experiences high productivity during the summer monsoon season (Banse, 1987; Nair et al., 1989; Naidu et al., 1992), making Arabian Sea one of the most productive location in the world (Fig. 1b). The summer monsoon season also accounts for 70-80% of the flux of organic matter to the sediments (Nair et al., 1989) and the southwest monsoon current drives upwelling and productivity changes in the Maldives region and other regions over Indian Ocean (Schulte et al., 1999). The summer monsoon anti-cyclonic gyre also results in weak upwelling along the coast of southwestern India and supports increased productivity during the summer season (Naidu et al., 1992). However local precipitation and land runoff results in a stratification and development of low-salinity layer (5-10m thick) over the up-welled waters (Schulte et al., 1999). Thus, inspite of a shallow thermocline the effect of upwelling is small. Due to the inflow of low salinity surface waters from the Bay of Bengal (Schulte et al., 1999) there is no increase in primary productivity during the northeast monsoon though there is deepening of the mixed layer during this time. Thus the upwelling and productivity changes in the Maldives region in the present day ocean could be related to summer monsoon variability. A spatial pattern documenting wide range of paleoceanographic and paleoclimatologic responses to the green house condition can be recorded from the sedimentary deposit from the Indian Ocean region. Monsoonal variability on different time scales ranging from annual cycles to long-term trends of millions of years can be well documented from these sedimentary records. A study on the records of the concentration and flux of biogenic components, such as calcium carbonate, organic carbon, and the composition of benthic and planktonic foraminiferal assemblages gives rise to indications of monsoonal variability as well as upwelling intensity (Prell et al., 1992).

5.1 The Indian summer monsoon (ISM) and climate change using multi proxy data

The ISM commenced about ~20 Ma ago because of uplift of the Tibetan Plateau beyond a critical height (Harrison et al., 1992; Prell & Kutzbach, 1992; Molnar et al., 1993). During summer, heating of the Tibetan plateau creates low atmospheric pressure, which acts as a powerful pump for moist air from the oceans to travel large distance, resulting in heavy rain (Quade et al., 1995). Reverse circulation of wind occurs during the winter; the radiative cooling of Tibetan Plateau causes flow of cold dry continental air towards the Indian Ocean. It has been argued that uplift of the Himalayas and Tibetan Plateau strengthen the land-sea thermal contrast (Hahn & Manabe, 1975) that led to the onset or major intensification of the Indian

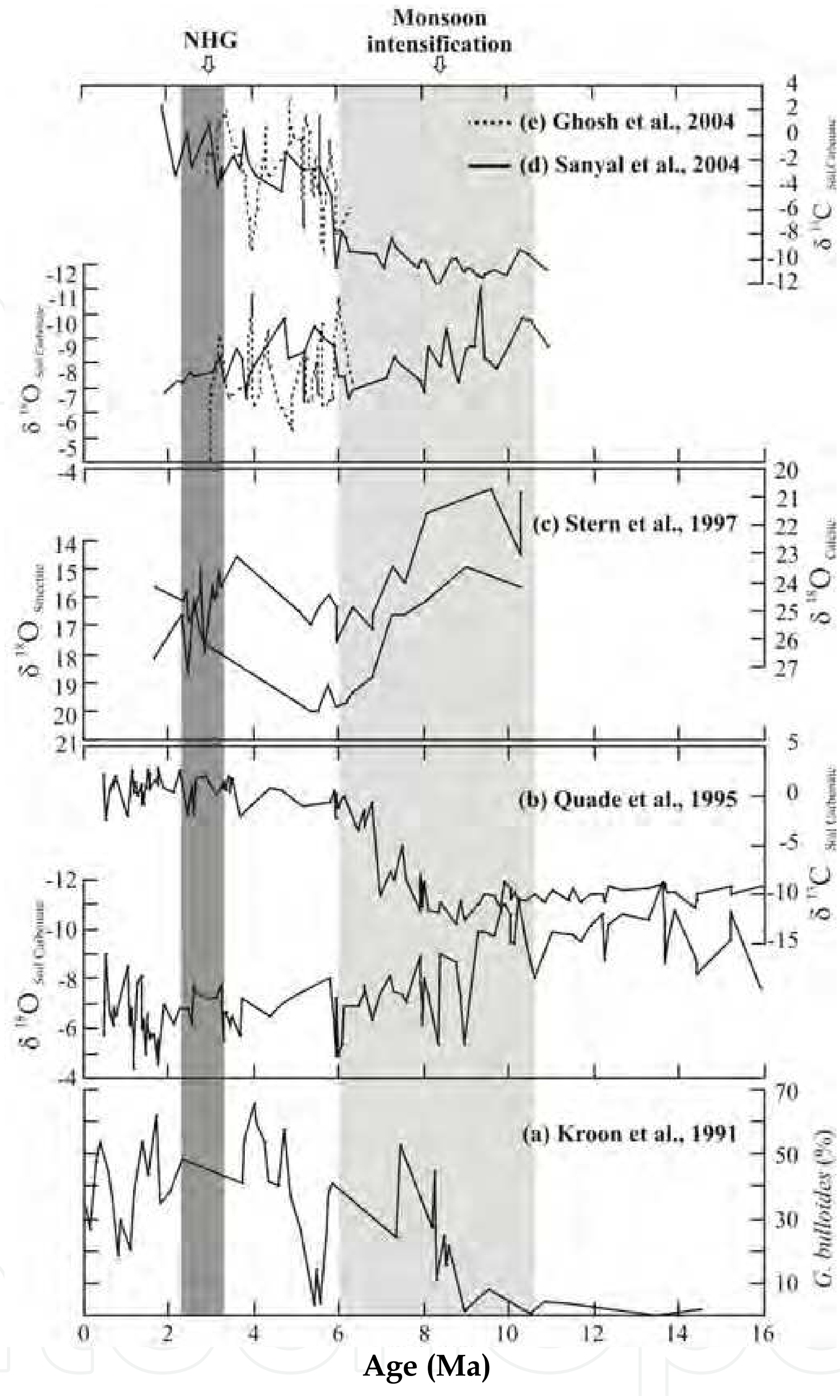

Fig. 4. (a) relative abundances of *G. bulloides* in time (Kroon et al., 1991); (b) The $\delta^{18}O$ and $\delta^{13}C$ of the soil carbonate from Siwalik Group fluvial sediments in the northern Pakistan (Quade et al., 1995); (c) $\delta^{18}O$ values of Calcite nodules and Smectite (Stren et al., 1997); (d) $\delta^{18}O$ and $\delta^{13}C$ of the soil carbonate from Siwalik basin (Sanyal et al., 2004); (e) $\delta^{18}O$ and $\delta^{13}C$ of the soil carbonate from Upper Siwalik (Ghosh et al., 2004). Grey zone shows a major excursion of monsoon intensification. Carbon isotope ratio of soil carbonate and associated organic matter indicates that vegetation was entirely of C_3 type from ~11 to 6 Ma. Post time is marked by appearance and expansion of C_4 grass. Dark grey zone (at ~3.2-2.3 Ma) indicates beginning of the major intensification of the Northern Hemisphere glaciation (NHG) (Zachos et al., 2001)

monsoon marked by the summer monsoon with heavy rainfall over the Indian subcontinent (Hastenrath, 1991) and cold dry winter monsoon. According to this hypothesis, changes in the elevation of the Himalayan-Tibetan region and its snow cover have modulated the development of the Indian or south Asian monsoons since the middle to late Miocene (Molnar et al., 1993; Clift et al., 2002). Many proxies indicate that the intensification of monsoonal winds led to increased upwelling over the Arabian Sea (Kroon et al., 1991) and eastern Indian Ocean (Singh & Gupta, 2004), a shift from C_3 to C_4 type vegetation on land (Quade et al., 1989), and increased terrigenous flux to the Indian Ocean as a result of increased weathering and erosion in the uplifted mountainous region (Prell & Kutzbach, 1992).

The high weathering rates increased nutrient flux (including phosphorus) to the oceans, increasing oceanic productivity (Filipelli, 1997). However, the timing of these events (like the uplift of the Himalayan-Tibetan Plateau, changes in the monsoon and changes in vegetation in south Asia) is not well constrained (Gupta et al., 2004). For instance, the dates for the elevation (~ 4 km) of the Tibetan Plateau, required to drive monsoon, vary from 35 to 8 Ma (Gupta et al., 2004). The long-term changes in the monsoon appear to have been linked with the uplift of the Himalayas and Tibetan Plateau (Rea, 1992; Prell et al., 1992; Molnar et al., 1993; Prell & Kutzbach, 1997; Filipelli, 1997) and Northern Hemisphere glaciations (Gupta & Thomas, 2003; Singh & Gupta, 2005). On the other hand, the short-term changes in the monsoon have been linked to the changes in the North Atlantic and Eurasian snow cover (Schulz et al., 1998; Anderson et al., 2002; Gupta et al., 2003), orbital cycle of eccentricity, tilt and precession (Prell & Kutzbach, 1987; Gupta et al., 2001) solar influence (Neff et al., 2001; Fleitmann et al., 2003; Gupta et al., 2005).

The Indian monsoon system is believed to have originated in the late Oligocene or early Miocene, and appears to have a major shift in its intensity at ~8.5 Ma. The ISM substantially increased its strength after 10-8 Ma (Quade et al., 1989; Kroon et al., 1991; An et al., 2001; Gupta et al., 2004). During middle Miocene (~15 Ma) the productivity record showed marked increment, significantly enhancing the productivity in all the oceans and reached maximum 10-8 Ma (Dickens & Owen, 1999; Hermoyian & Owen, 2001). The high productivity correspond to a "biogenic bloom" that began ~15-13 Ma and peaked 10-8 Ma (Pisias et al., 1995; Dickens & Owen, 1999; Hermoyian & Owen, 2001). Various proxy records have been interpreted as indicating that the monsoons started or strongly intensified between ~10 and 8 Ma, as a response to Himalayan-Tibetan uplift to at least about half of its present elevation (Prell & Kutzbach, 1992; Rea, 1992). The Indian Ocean high productivity event occurred (~10-8 Ma) at the end of a phase of build-up of the East Antarctic ice sheet and possibly the beginning of the formation of the west Antarctic ice sheet (Zachos et al., 2001, Barker & Thomas, 2004), and as well as in the global compilation of deep-sea oxygen isotope records (Zachos et al., 2001; Fig. 2). The biological, sedimentological and geochemical responses to this late Miocene event have been observed across the Indian, Atlantic and Pacific Oceans (Kroon et al., 1991; Dickens & Owen, 1999; Hermoyian & Owen, 2001; Gupta et al., 2004). During the late Miocene, the increased glaciation on Antarctica may have strengthened wind system, causing widespread open-ocean as well as coastal upwelling over a large part of the Atlantic, Indian and Pacific Ocean. This increased upwelling could have triggered the widespread biological productivity during the late Miocene (Gupta et al., 2004). Harrison et al. (1992) suggested rapid uplift around ~8 Ma of Tibetan Plateau, may have acted as an effective orographic barrier inducing depleted $\delta^{18}O$

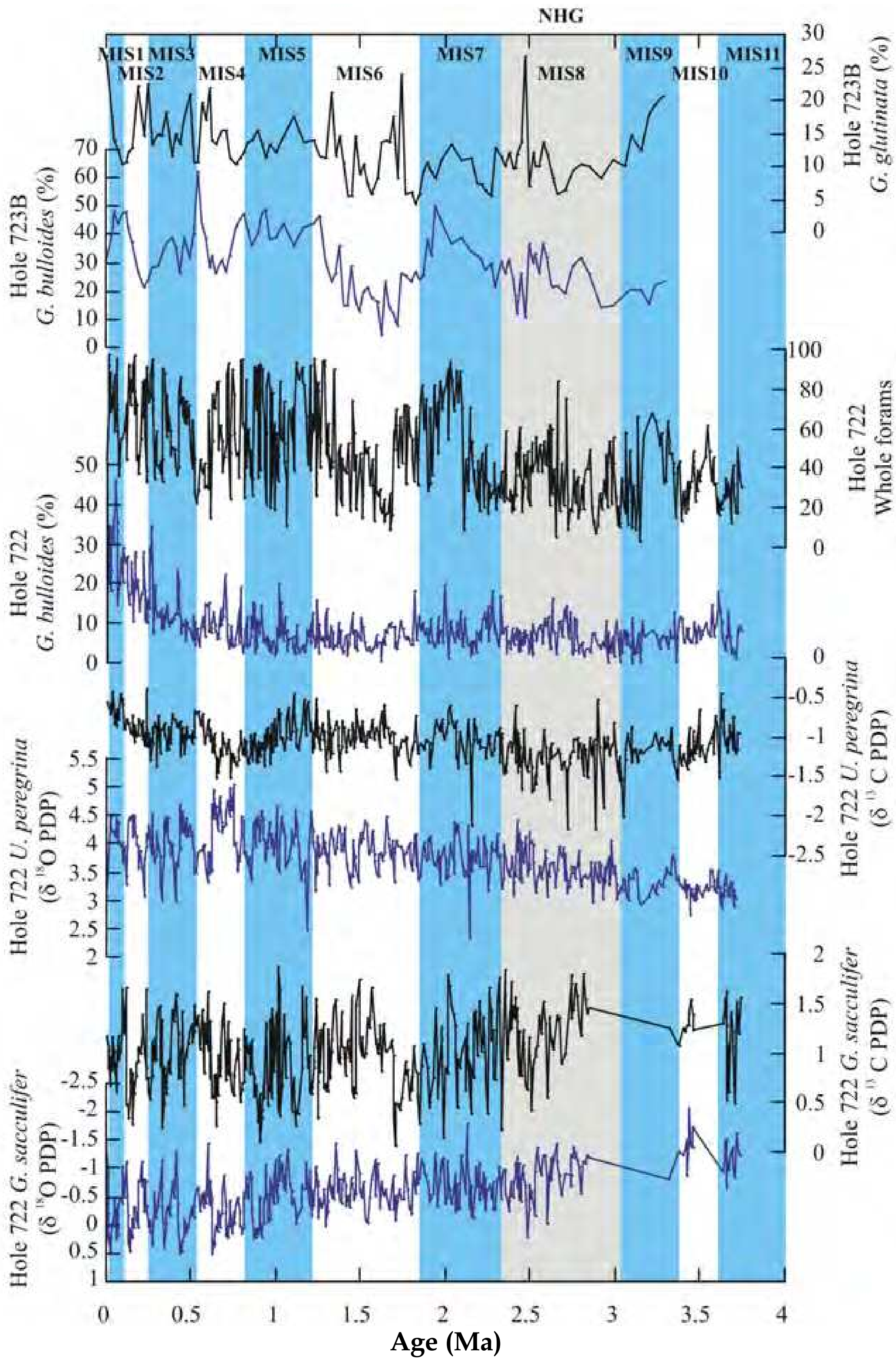

Fig. 5. Visual correlation between marine proxy records of the Indian summer monsoon during Plio-Pleistocene time at Arabian Sea using benthic and planktonic foraminifera with isotopes values (Clemens et al., 1996). Dark grey zone (at ~3.2-2.3 Ma) indicates beginning of the major intensification of the Northern Hemisphere glaciation (NHG) (Zachos et al., 2001; Clemens et al., 1991, 2003) and Blue color bars indicates Marine Isotope Stages (MIS 1-11)

value for the precipitation originating from the central Asia, while higher $\delta^{18}O$ values for the precipitation happening on the Potwar Plateau. The $^{18}O/^{16}O$ value increase observed at 8.5 and 6.5 Ma is impossible to distinguish and relate with the potential causal mechanism, but these clay mineral $\delta^{18}O$ values support that there was the significant climate changes (Stern et al, 1997; Fig. 4). Qaude et al. (1989) have been reported ~3.5 $^{o}/_{oo}$ increase in $\delta^{18}O$ in the soil formed calcite in this sequence of paleosols at ~8.5-6.5 Ma, so there is a significant isotope change which potentially could be preserved by the clay minerals and there is dramatic vegetation change from forest to grassland slightly postdating (7.7-6.5 Ma) the oxygen isotope ratio increase (Fig. 4), due to the intensification of monsoon in this region at 8.5-6.5 Ma (Quade et al., 1989, 1995). Sanyal et al. (2004) documented $\delta^{13}C$ values of soil carbonates show that, from 10.5 to 6 Ma, the vegetation was C_3 type and around ~6 Ma C_4 grasses dominated. The $\delta^{18}O$ variations of soil carbonates suggest that the monsoon system intensified, with one probable peak at around 10.5 Ma and a clear intensification at 6 Ma, with peak at 5.5 Ma. After 5.5 Ma, monsoon strength decreased and attained the modern-day values with minor fluctuations, which is supported by marine proxy of upwelling in the Arabian Sea. During the same time Ghosh et al. (2004) has proposed that greater moisture availability ~4 Ma due to the higher monsoonal activity, which caused an abrupt transient of C_3 dominance in this region and $\delta^{18}O$ of carbonate and δD of clay support depleting trend observed at ~4 Ma. One of the strongest arguments for an intensification of the Indian monsoon ~10 Ma came from deep-ocean drilling in the Arabian Sea by Kroon et al. (1991). During both summer and winter monsoons, steady winds induce upwelling of cold, deep, nutrient-rich water. Studies of *Globigerina bulloides* from this region shows that thrives in cold water where nutrients upwell, becomes abundant during monsoons, particularly in summer, and dominates organic sediment accumulation (Curry et al. 1992, Prell & Curry 1981). *G. bulloides* started at ~14 Ma, and for a few million years, it comprised only few percent of the annual mass accumulation of organic sediment in the western Arabian Sea. Then beginning near ~10 Ma, it suddenly comprised tens of percent of organic sediment deposition (Fig. 4), suggesting an intensification of monsoon winds at that time (Kroon et al., 1991). The $\delta^{18}O$ value variations at different Siwalik sections, sedimentary record from foothill of Himalya suggest three phases of lowering of monsoon strength at around ~10.5 Ma, ~5.5 Ma and ~3 Ma, which correspond to periods of intensification of the Indian summer monsoon (Fig. 4a–d based on Kroon et al. 1991; Quade et al., 1995; Stern et al. 1997; Sanyal et al. 2004 and Ghosh et al., 2004). Benthic and planktonic foraminiferal ratios were used to estimate Plio-Pleistocene climate changes (Fig. 5) and specially there is prominent presence of MIS 3 (Marine Isotope Stages) and DO cycles (Dansgaard-Oeschger); marine C^{14} reservoir ages and the incremental time scale of the ice cores are not known well enough to resolve potential difference in timing which would allow inferences about the mechanism of global energy transfer (Sarnethein, 2001, 2002; Wang et al., 2005). DO cycle are not restricted to MIS 3 and monsoon proxies from Arabian Sea records shows that the transition from interglacial MIS 5 to glacial MIS 4 occurred in discrete steps, characterized by a number of warming rebounds. This record, chronologically constrained by the well dated sediments from Arabian Sea (Fig. 5). Clemens et al. (1991, 2003) observed together with *G. bulloides* other various proxy records of summer monsoon strength at 3.5 Ma. They suggested that of strong monsoon has systematically drifted over the past 2.6 Ma (refer Fig 5). During the initiation and growth of NGH ice sheets, the phase of strong monsoon moves away from the phase maximum ice-volume, systematically shifting by ~83 and ~124 k.y., at the precession and obliquity bands, respectively (Clemens et al., 1991, 2003).

5.2 Quaternary climate change

During the Quaternary time the Indian monsoon system also underwent parallel changes like glacial and interglacial mode (Schuldz et al., 1998; Leuschner & Sirocko, 2003). The cold spell intervals in the North Atlantic have been found to be associated with intervals of weak summer monsoon (Schuldz et al., 1998), the winter monsoon strengthened during the same time (Fontunge & Duplessy, 1986). The summer monsoon oscillated with millennial scale variability (the Dansgaard-Oescher-DO) and Henrich events concentrated at periodicities of ~1100, 1450 and 1750 years during the last glacial cycles (Naidu & Malmgren, 1995; Siriko et al., 1996). The pattern is almost similar to that of changes recorded from the Greenland ice cores (Schulz et al., 1998; Leuschner & Sirocko, 2003; Fig. 6). The atmospheric CO_2, ice core data (GSIP2) with Asian speleothems and Arabian Sea cores uniformly indicate the monsoon was weaker during cold intervals such as Heinrich event 1 (H1) and the Younger Dryas (YD), and stronger during the warming period observed during the Bølling–Allerød (B–A) period (Fig. 6; Schulz et al., 1998; Altabet et al., 2002; Ivanochko et al., 2005; Sinha et al., 2005; Wang et al., 2001; Dykoski et al., 2005; Sirocko et al., 1996). The deglaciation affecting Asian monsoon intensity as recorded in the speoleothems samples analysed from the Chinese and Indian sites (see Fig. 6; Sinha et al., 2005; Dykoski et al., 2005). These findings suggest a strong solar influence on the monsoon during the Holocene (Dykoski et al., 2005; Fleitmann et al., 2003; Wang et al., 2005; Neff et al., 2001; Fig. 7). The speleothem record from China and India region shows similar trend from ~16000–11000 calendar years before present (cal yr B.P.) (Fig. 6). Oxygen isotopes anomaly at D4 are also thought to be driven by an amount effect and are interpreted to reflect changes in the strength of the Asian monsoon (Dykoski et al., 2005). In general, these all speleothem time series are remarkably similar (Fig. 6 & 7). Shakun et al. (2007) suggested these climate events affected a large area of the monsoon region in the same way and to the same degree because the magnitudes of the transitions into the Bølling, the YD, and the Holocene are nearly identical in the speleothem records from these caves. Even more interesting, the structures of these climate changes are similar. In particular, the transitions into the Bølling period and the YD in all these records, as well as the Timta Cave record from northern India (Sinha et al., 2005), are remarkably gradual and take place over several centuries (Fig. 6).

The characteristic feature of ISM during Holocene period is presentence presence oscillating intensity with frequency of occurrences of events coinciding with millennial time scale. Goodbred & Kuel (2000) reported the ISM was stronger in the early Holocene, which is evident from the huge sediments deposits at Ganges-Brahmaputra, the rapid speleothem growth (see Fig. 1) eg. Hoti Cave (Neff et al., 2001) and Qunf Cave at Oman margin (Fleitmann et al., 2003), Socotra Island Yemen at northwestern Indian Ocean (Shakun, et al. 2007), Timta Cave at western Himalaya (Sinha et al., 2005), Hulu Cave at eastern China (Wang et al., 2004), Dongge Cave at southern China (Wang et al., 2005), Dandak Cave at central India (Berkelhammer, et al. 2010) and *G. bulloides* census data from Arabian Sea (Gupta et al., 2004). Northern Hemisphere temperatures peaked at ~10400 and 5500 cal yr B.P., and during which time Asian and African monsoon reached their maximum so called "Holocene Climatic Optimum". Alley et al. (1997) suggested the early Holocene monsoon maximum was interrupted by an abrupt cooling peak ~8200 cal yr B.P., the summer monsoon over the Indian subcontinent and tropical Africa weakened during same time (Gasse, 2000). After ~ 5500 cal yr B.P., consecutive shifts towards drier condition in northern Africa and Asian was noted (Overpec al., 1996; Gasse, 2000), which led to termination of

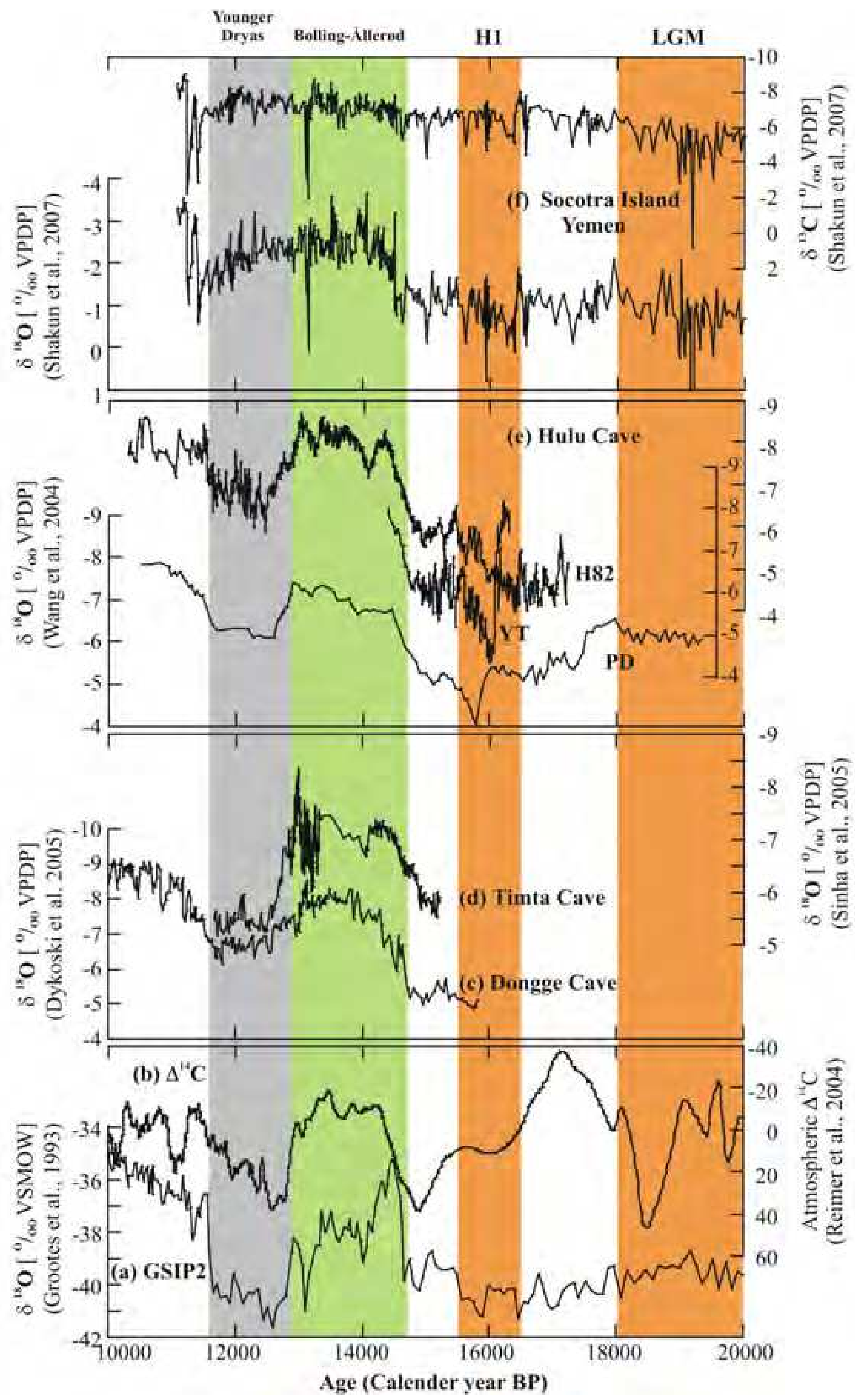

Fig. 6. Visual correlation between various proxy records of the Indian summer monsoon. a) GSIP2 δ18O record (Grootes et al., 1993); (b) Atmospheric Δ14C record (Remier et al., 2004); (c) Oxygen isotope record from Dongge Cave at southern China (Dykoski et al., 2006); (d) Timta Cave δ18O at Western Himalaya (Sinha et al., 2005); (e) δ18O record from Hulu Cave at eastern China (Wang et al., 2004) and (f) δ18O and δ13C record Socotra Island Yemen at Northwestern Indian Ocean (Shakun, et al. 2007). Grey band indicate the timing and duration of the Younger Dryas (YD) and green bars indicate Bølling–Allerød (B–A). Orange color bars number H1 indicates Henrich events with LGM (Last glacial maximum).

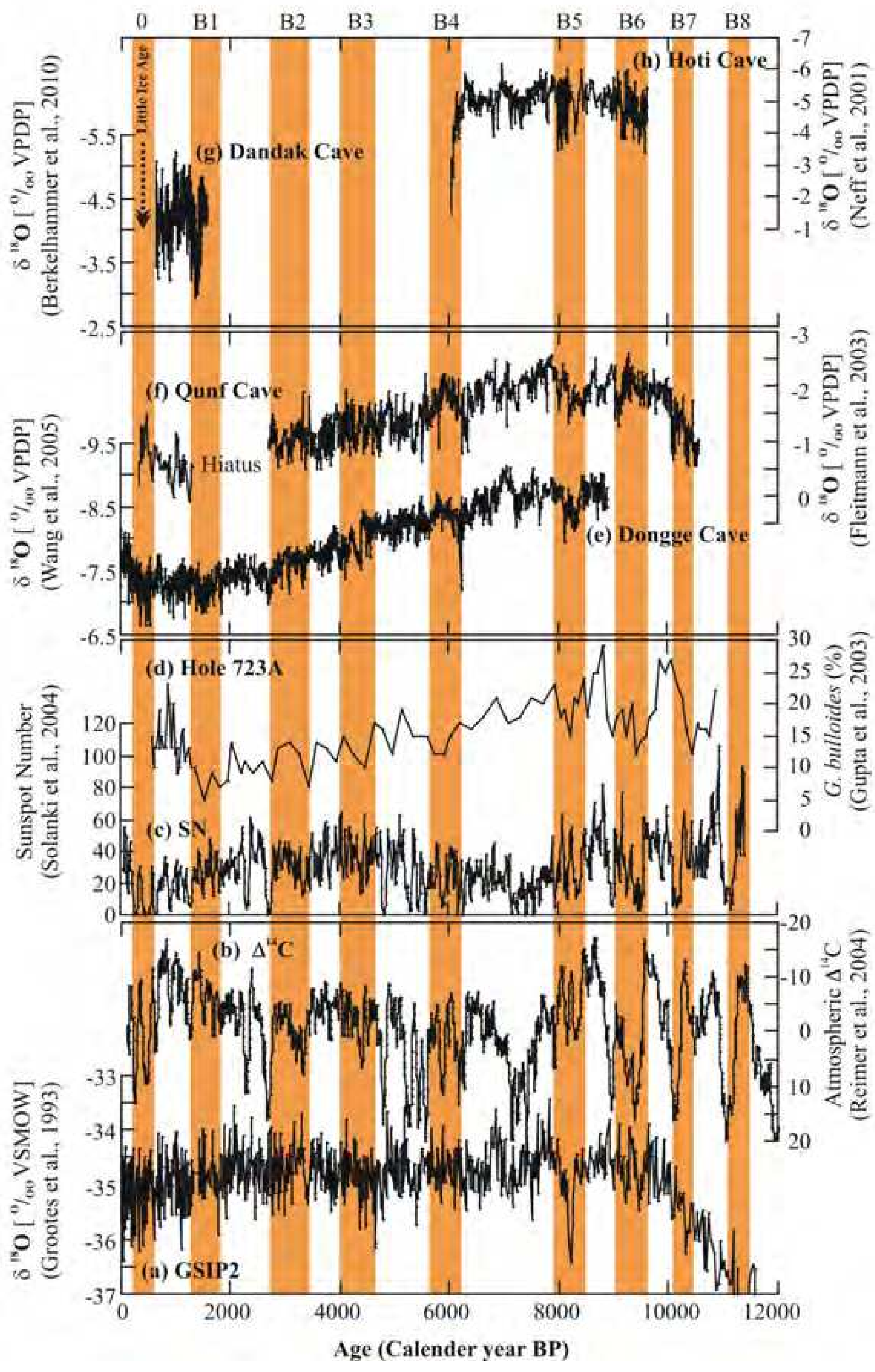

Fig. 7. Correlation between various proxy records of the Indian summer monsoon. (a) GSIP2 $\delta^{18}O$ record (Grootes et al., 1993); (b) Atmospheric $\Delta^{14}C$ record (Reimer et al., 2004); Sunspot number (Solanki et al., 2004); (d) *G. bulloides* census data from Arabian Sea (Gupta et al., 2004); (e) $\delta^{18}O$ data from Dongge Cave at southern China (Wang et al., 2005); (f) $\delta^{18}O$ record from Qunf Cave at Oman margin (Fleitmann et al., 2003); (g) , Dandak Cave $\delta^{18}O$ at central India (Berkelhammer, et al. 2010) and $\delta^{18}O$ Hoti Cave (Neff et al., 2001). Numbers from B1 to B8 indicate each of the eight Bond events and number 0 indicates Little Ice Age event (Bond et al., 2001)

several ancient civilization existed in this regions. During ~4200 cal yr B.P., persistent record of drought due to the aridification led to societal collapse both in the Egyptian and Mesopotamian civilization (Weiss et al., 1993; Cullen et al., 2000). The precipitation diminishes during the arid phase over Indian region as well (Sharma et al., 2004), and the Indus valley civilization transformed from an organized urban phase of smaller settlements migrated southward in search of water (Allchin & Allchin, 1997; Staubwasser et al., 2003; Gupta et al., 2004). The weakest summer monsoon occurred in the late Holocene around ~2500 to 1500 cal yr B.P. (Anderson et al., 2002; Gupta et al., 2003; Fig. 7). Gupta et al. (2003) have been suggested the summer monsoon intensified during the Medieval Warm Priod (AD. 900-1400) whereas during the most recent climatic event, the little ice age (AD. 1450-1850), there was a drastic reduction in the intensity of the ISM (Fig. 7; Gupta et al., 2003). The ISM is controlled by the Eurasian snow cover (Bamzai & Shukla, 1999) too and the amplitude and period of ENSO (Krishna Kumar et al., 1999). ISM variability on millennial scale may be attributed to the solar forcing (Neff et al., 2001; Fleitmen et al., 2003) and glacial and interglacial boundary condition (Burns et al., 2001). The $\delta^{18}O$ record from the stalagmite, which serves as a proxy for variations in the tropical circulation and monsoon rainfall, allows us to make a direct comparison of the $\delta^{18}O$ record with the $\Delta^{14}C$ record from tree rings (Fig. 7; Stuiver et al., 1998; Neff et al., 2001), which largely reflects changes in solar activity (Stuiver & Braziunas,, 1993; Beer et al., 2000). Bond et al. (2001) suggested that solar variability could be an affecting climate variation during the Holocene on the basis of analysis of ^{14}C records from the tree rings. The production rate of comogenic nuclides (^{14}C and ^{10}Be) that reflect changes in solar activity appear to closely follow the Bond Cycles (see Fig. 7). van Geel et al. (1999) has pointed out that the exact process responsible for the linking global climate through solar forcing is poorly understood, variations in ultra violet radiation and cosmic ray flux may trigger abrupt climate changes by the altering the heat budget of the stratosphere and changing the atmospheres optical parameters and radiation balance (Kodera, 2004). The summer monsoon strengthened has been related to changes solar insolation during Milankovitch cycles (Clemans et al., 1991), but centennial and decadal variations in the $\Delta^{14}C$ record are controlled by changes in solar activities or sun spot numbers, (see Fig. 7); Wang et al., 1999; Neff et al., 2001; Gupta et al., 2005). During the Holocene, the relation between intervals of low sunspot activity and low intensity of the summer monsoon were observed in marine records (Fig. 7; Gupta et al., 2005).

6. Conclusions

The past ~65 million years of geological history documented in the continental and marine sedimentary record allowed understanding the evolution of climate change in a longer time scales. The effect of green house gas concentration on global climate and moisture circulation can be explored through careful investigation of available proxy records. The impact of other factors like topography and albedo change can be addressed knowing the magnitude of upliftment rate of mountain region. The simultaneous analyses of proxy records from both land and marine location provided conclusive signature of climate change through time. The $\delta^{18}O$ of foraminiferal calcite from sea sediments provided independent estimate of time for commencement of Antarctic and Arctic glaciations (Zachos et al., 2001). The increasing and decreasing trends suggested based on the $\delta^{18}O$ record demarcate periods of global warming and cooling, responsible for growth and decay of ice sheet. The elevation (~ 4 km) of the Tibetan Plateau, required to drive the monsoon, varied significantly between

35 to 8 Ma. The combination of factors like pCO_2 concentration in the atmosphere and rate of uplift of Tibetan plateau were found responsible for modulating the intensity of Indian monsoon. The published data from land and oceanic region indicate a major changes land vegetation and oceanic productivity during period of intense monsoon. The high southern latitude cooling and increased volume of ice act as important factors responsible for lowering the strength of monsoonal circulation. The signature for strengthening of the upwelling, presumably from strengthening of seasonal winds over the Indian Ocean was noted as a proxy for monsoon intensification. The signature of increasing aridity and warming in northern Pakistan perhaps increases both in temperature and seasonal precipitation in the Indo-Gangetic plain just south of the Himalayas in Nepal and in the Himalayas are recorded from soil carbonate preserved in the sedimentary archives. Together with ice core record, speleothems occurrences from Asian region act as a recorder of Holocene climate and monsoonal variability. The role of sun, green house gases and tectonic adjustment of continental lithosphere played significant role controlling the monsoon over Indian region. Understanding the sensitivity of all these factors on rainfall or precipitation is yet to be understood.

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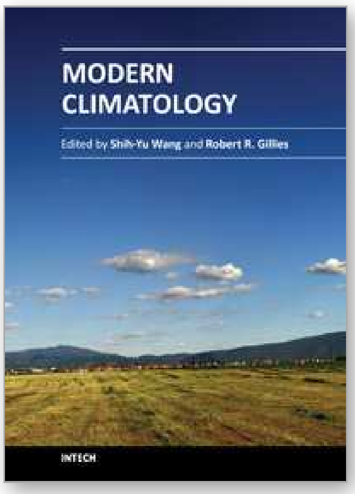

Modern Climatology

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Climatology, the study of climate, is no longer regarded as a single discipline that treats climate as something that fluctuates only within the unchanging boundaries described by historical statistics. The field has recognized that climate is something that changes continually under the influence of physical and biological forces and so, cannot be understood in isolation but rather, is one that includes diverse scientific disciplines that play their role in understanding a highly complex coupled "whole system" that is the earth's climate. The modern era of climatology is echoed in this book. On the one hand it offers a broad synoptic perspective but also considers the regional standpoint, as it is this that affects what people need from climatology. Aspects on the topic of climate change - what is often considered a contradiction in terms - is also addressed. It is all too evident these days that what recent work in climatology has revealed carries profound implications for economic and social policy; it is with these in mind that the final chapters consider acumens as to the application of what has been learned to date.

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