# Indian Monsoon Variability at Different Time Scales: Marine and Terrestrial Proxy Records



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***Status Report***

# Indian Monsoon Variability at Different Time Scales: Marine and Terrestrial Proxy Records

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Here, we present a review of the work done in India during 2007-2011 on various proxy records of monsoon variability preserved in the marine (Central Indian Basin, western, northern and eastern Arabian Sea and the Bay of Bengal) and terrestrial (Northeast, Gujarat, Himalayan Lakes, Siwaliks, Thar Desert, Ganga Valley, etc.) realms, in order to understand how Indian monsoon has evolved through the Tertiary and Quaternary periods. Though, there are clear indications of heavy rainfall occurrence throughout the Palaeogene and early Neogene, seasonality in rainfall pattern becomes apparent only in the Late Miocene record. Northern Hemisphere Glaciation played a major role in the evolution of Pliocene monsoon, whereas, glacial-interglacial cycles influenced the Pleistocene monsoonal variability. The Holocene, which is characterized by millennial-scale climatic fluctuations, started with a strong monsoonal phase, often known as "Holocene Climatic Optimum", lasted till the mid Holocene. This strong phase was followed by weak phases causing drier conditions mostly around ~2,500-1,500, 1000, 650-450 yrs BP and the little Ice Age (AD. 1450-1850).

**Key Words: Monsoon Variability; Marine; Terrestrial; Proxy Record**

## Introduction

The Indian or South Asian monsoon is a unique climatic phenomenon characterized by seasonal reversal of winds over the Indian subcontinent. In summer, intense and moisture-laden winds blow from the southwest (Southwest monsoon or SWM) whereas in winter mostly dry and variable winds blow from the northeast (Northeast Monsoon or NEM), primarily caused by the difference in land-sea temperature and pressure conditions. Although, monsoon may appear to be regular in the sense that it occurs every year, its intensity, distribution and timings could vary considerably, often causing heavy floods and droughts. Since the Indian economy is heavily dependent on summer monsoon rains, understanding its past and present variability is of immense importance in making any attempt in predicting future monsoon variability.

Monsoon variability occurs at different timescales as revealed by the instrumental, historical and various proxy records. Among the terrestrial archives, tree rings, palaeosols, speleothemes, fluvio-lacustrine sediments, peat deposits, microfossils, magnetic minerals and plant phytoliths are widely used to reconstruct past monsoonal conditions. Marine proxy data derived mostly from foraminifer abundance, their stable oxygen and carbon isotope ratios, organic content of sediments, etc. Intraseasonal monsoon variability is associated with the break and active phases of the monsoon oscillating in the range of 30 to 50 days. These factors depend on tropical mid-latitude interactions, changing in sea surface temperature, atmospheric variability and soil moisture. Interannual fluctuations, such as droughts and floods are known to be controlled by El Nino-Southern Oscillations (ENSO), snow cover, land surface features and top layer characteristics of the tropical oceans. Variations on the decadal and century time scales are forced by changes in the monsoon circulation pattern, surface boundary conditions, north Atlantic deep water production and solar activity. Millennial and longer scale variations in the areal extent of monsoon are caused by global climatic excursions, solar insolation at the Milankovich scales (glacials/ interglacials), tectonics (Tibetan-Himalayan uplift) and changes in atmospheric $CO_2$. All these factors may function at the same time and over different time scales to intensify

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or reduce the seasonal progress of continental heating/ cooling, land-sea pressure gradients, latent heat transport, and moisture convergence, which in turn control the strength of the monsoon circulation (Gupta, 2010, Gupta, 2008; Tiwari *et al*., 2011 and references therein for a review).

## The Tertiary Period

### *Palaeogene*

While passing through the equator, in the Palaeogene time the Indian landmass had developed extensive wetlands and blanketed vegetation as evidenced by the occurrence of coal/lignite deposits in the northeast and northwestern parts of India (Fig. 1). Lakadong Formation (Late Palaeocene of Khasi Hills, Meghalaya), Mikir Formation (Early Eocene of Mikir Hills), Kopili Formation (Late Eocene, Meghalaya and Assam), Tikak Parbat Formation (Late Oligocene, Makum Coal fields, Upper Assam), Matanomadh and Naredi formations (Late Palaeocene and Early Eocene) of Kutch , Akli Formation (early Eocene ligtes of Rajasthan), contain large number of fossil plants derived from tropical-subtropical rainforests (Prasad *et al*., 2009; Saxena and Trivedi, 2009; Tripathi *et al*. 2009; Trivedi, 2009). Besides a diverse palynoassemblage, the coal of northeast India have yielded leaves (Mehrotra *et al*., 2009) and fruit legumes of drought resistant plant adapted to monsoon season (Srivastava and Mehrotra, 2010). Wang *et al*. (1997) subdivided the evolution of East Asian monsoon into four stages: Pre-monsoon (Palaeocene-Early Eocene), transitional stage (Late Eocene-Oligocene), stage I (Miocene-Pliocene) and stage II (Late Pliocene-Present).

## The Neogene

### *Miocene and Pliocene*

Collision of India with Asia, followed by an uplift of the Himalayan-Tibetan Plateau might have created a temperature-pressure difference facilitating monsoonal circulation. Although, India-Asia collision may have taken place ~55 or 34 Myr ago (Aitchison *et al*., 2007), the real deformation and Himalayan exhumation began only in the Early Miocene, some 23 Ma ago (Clift *et al*., 2008). Data from South China Sea, Arabian Sea and Bay of Bengal shows that, monsoon driven erosion intensified in the Miocene reaching its peak by ~15 Ma ago and remained

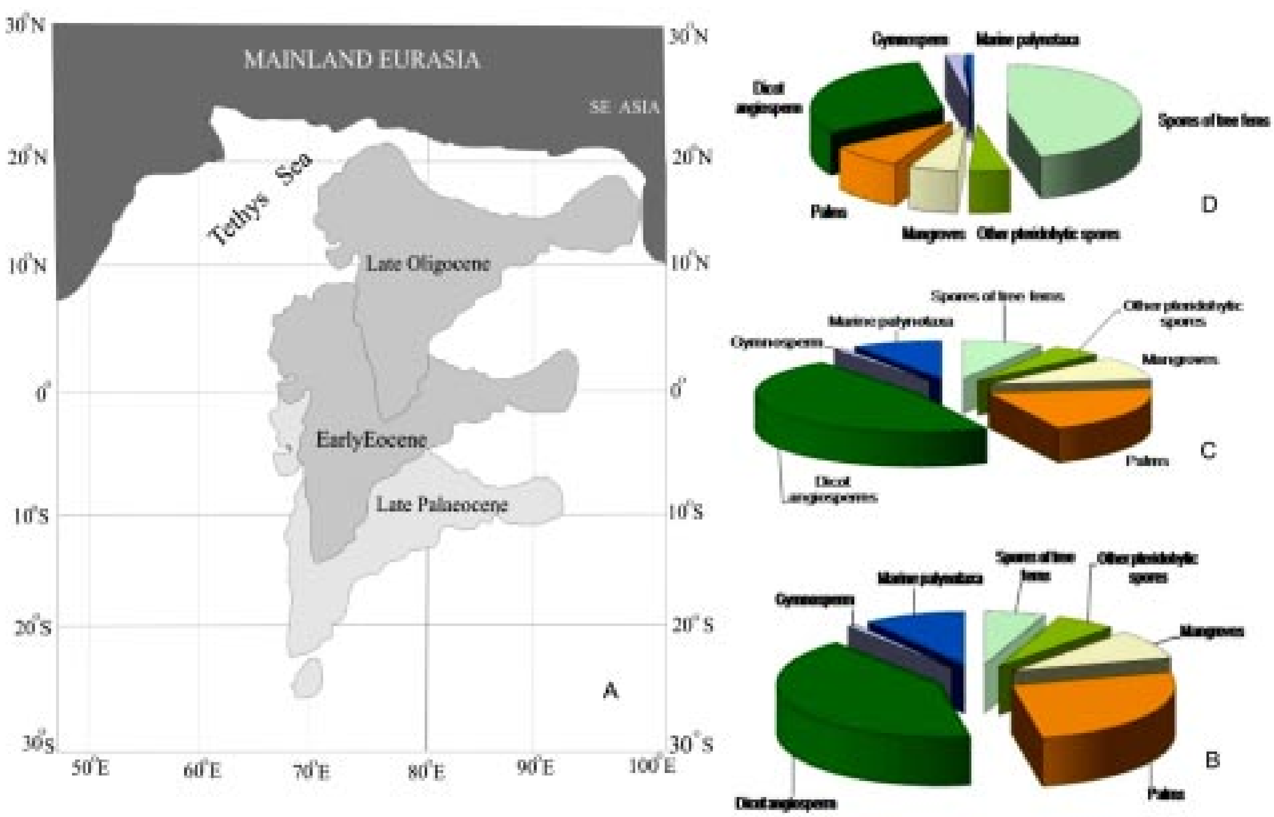


**Fig. 1: A. Position of India during Late Palaeocene-Late Oligocene. Pie diagram showing frequency of palynoassemblages derived from various plant groups from, B. Late Palaeocene sediments of Meghalaya, C. Early Eocene sediments of Assam, D. Late Oligocene sediments of Upper Assam (Madhav Kumar, unpublished data)**

high till 10.5 Ma; thereafter it kept on decreasing till 3.5 Ma, followed by another increase in the intensification during the late Pliocene and the Pleistocene (Clift *et al*., 2008). The Middle Miocene was the time of warm and humid climate and evergreen to deciduous tropical forests covered a large part of the northwestern part of the Indian subcontinent (Ashton and Gunatilleke, 1987). By the Late Miocene global cooling, spread of arid conditions and possibly monsoon intensification brought change in the vegetation marked by fragmentation of forests and spreading of grasslands (Quade *et al*., 1989; Retallack, 1991). This ecological shift has resulted in the extinction of Siwalik Miocene apes and diversification of monsoon adapted murine rodents (Nelson, 2007; Patnaik and Chauhan, 2009). Analogous Late Miocene (10-8 Ma) monsoon intensification is well documented in the marine record (Kroon *et al*., 1991; Gupta and Srinivasan, 1992).

The Indian Siwalik palaeosol oxygen isotope data indicates the existence of multiple phases of monsoonal intensification with peaks at ~11, between 6 and 3 Ma exhibit a change in C3 to C4 type of vegetation coinciding with the monsoonal intensification at the Miocene-Pliocene Boundary. An asynchronous emergence of C4 vegetation noticed because of variation in monsoonal distribution in time and space (Sanyal *et al*., 2010;Sanyal and Sinha, 2010). The diversity and ecomorphological aspects of fossil murine rodents of Siwalik indicate that monsoon probably initiated in the Indian subcontinent around 14 Ma ago. The present day climatic zonations cannot be directly applied to the Late Miocene; because the mean annual rainfall in the NW part of the Indian subcontinent in the Early Pliocene (5-4 Ma) were ranging between 1000 and 2000 mm with temperature variability of 20 and 25ºC. During the Late Pliocene (3-1.8 Ma), the region received 2000-4000 mm annual rainfall under the average temperature 22.5 to 27.5ºC (Patnaik, 2011). Between 5 and 2.75 Ma, when southern hemisphere was blanketed with ice, strong summer and winter monsoons were noticed under both obliquity and precession bands (Clemens *et al*., 2007). Orbital forcing accompanied by the Northern hemisphere glaciation controlled monsoon in the Pliocene. There are several evidences indicating northern hemisphere glaciation and phased uplift of the Tibetan Plateau, decrease in intensity of summer monsoon and increase in the winter monsoon.

## The Quaternary Period

### *Pleistocene*

The climatic variation during Pleistocene are characterized by Milankovich cycles, switching from dominantly 41-kyr mode in the Early Pleistocene to that of a 100-kyr mode in the Middle Pleistocene, in response to a change in Earth's orbital forcing. During this time the global climate experienced glacial-interglacial phases, which in turn influenced the temporal and spatial monsoon variability. Clemens *et al*. (2007) found that between 2.75 and 1.25 Ma, at both orbital bands, the phase of summer monsoon proxies began to drift toward ice minima while the phase of winter monsoon indicators began to drift toward ice maxima; these drifts indicate that increased glacial boundary conditions tend, on an average, to strengthen the winter monsoon while increased interglacial boundary conditions tend, on an average, to strengthen the summer monsoon. Further, they noticed that between 1.25 Ma and the Present, the phase of the summer monsoon proxies reside between ice minima and precession maxima (Indian Ocean latent heat export); direct northern hemisphere sensible heating appears to play a lesser role. For example, southwest monsoon was stronger (weaker) during the warmer (colder) periods, with a minimum during the Last Glacial Maximum (Gupta *et al*., 2011; Tiwari *et al*., 2011; Ramesh *et al*., 2010; Sinha *et al*., 2007). During deglaciation, monsoon fluctuated widely with weaker monsoons during colder episodes such as Younger Dryas and stronger monsoons during warm episodes such as Bølling-Allerød. During the last glacial, the summer monsoon oscillated with millennial-scale variability (the Dansgaard-Oeschger and Heinrich events), concentrated at periodicities of 1,100, 1,450, 1,750 and 2,300 yr (Naidu and Malmgren, 1995).

The well dated Pleistocene marine faunal and sedimentary records from the Central Indian Basin, the Arabian Sea and Bay of Bengal provide excellent archives for understanding palaeomonsoonal variability. The benthic faunal assemblage from the Ocean Drilling Program (ODP) Site 238 of Central Indian Basin show a major shift at ~0.7 to 0.6 Ma, coinciding with an increased amplitude of glacial cycles, which in turn appear to have influenced low latitude monsoonal climate as well as deep-sea conditions (Gupta *et al*., 2006). Gupta *et al*. (2010) noticed a major decrease in the strength of the Indian Ocean equatorial westerlies across the mid-Brunhes epoch (~300-250 Kyr BP) matching with a wetter equatorial East Africa, a drier Australasia and a stronger Indian summer monsoon (Fig. 2). This study provides the first direct evidence of Indian Ocean Dipole in a paleo record (Gupta *et al*., 2010).

The eastern Arabian Sea (EAS), sea surface temperature (SST), $\delta^{18}O$ of sea water ($\delta^{18}Ow$), and salinity data indicate that Mg/Ca derived SST record varied by 4°C; implying that the marine isotope stage (MIS) 4 was warmer than MIS 3, Last Glacial Maximum was 4°C cooler than the present, and there was a 2°C increase within the Holocene (Govil and Naidu, 2010). They found that the transition from MIS 4 to MIS 3 reflects a decrease in the evaporation precipitation budget in the EAS, perhaps due to the strengthening of southwest monsoon, because of

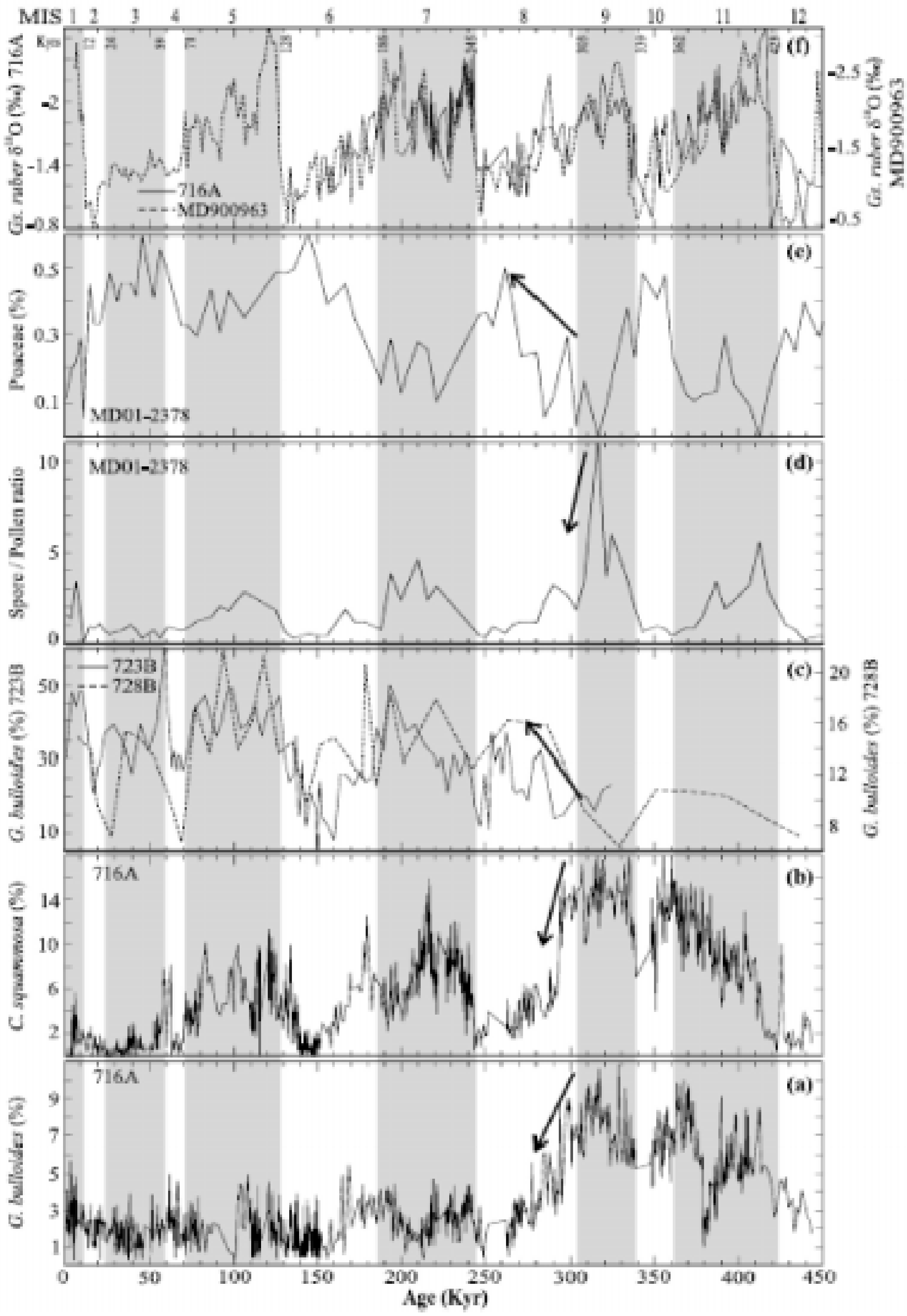


**Fig. 2: Proxy records from the equatorial Indian Ocean, NW Arabian Sea and Timor Sea. (a) Percent distribution of *Globigerina bulloides* at Hole 716A. A major decrease in *G. bulloides* percentages at ~300 Kyr coincides with a major shift in NW Arabian Sea, East African and Australasian climates; (b) benthic foraminifer *Cymbaloporetta squammosa* – an indicator of more productive waters, also shows a major decline at Hole 716A at ~300 Kyr; (c) *G. bulloides* percentages show a major increase at NW Arabian Sea Holes 723B and 728B (this study) indicating intensification of the SW monsoon between ~300 and 250 Kyr. (d) A decrease in spore/pollen ratio and (e) an increase in Poaceae population at ~300–250 Kyr suggest the beginning of an arid phase in Australasia including Indonesia, northern Australia and adjacent regions; (f) oxygen isotope values of *Globigerinoides ruber* from Hole 716A and core MD900963 used to interpolate ages of the samples after tuning with the SPECMAP stacked record. Grey bars illustrate interglacial marine isotope stages (MIS) (from Gupta *et al.*, 2010 and references therein)**

higher precipitation during MIS 3 and MIS 1 and lower during MIS 2 and MIS 4. The amplitude of monsoon fluctuations derived through upwelling indices and $\delta^{18}$Ow varies significantly, may indicate spatial variability of monsoon rainfall. Further Govil and Naidu, (2011) noticed that hydrography is highly influenced by the river runoff and rainfall during the southwest monsoon, with the SST during the LGM ~3.2ºC cooler than the present, and a ~3.5ºC rise is documented from 17 to 10 ka in the Bay of Bengal (Fig. 3). Both SST and $\delta^{18}$Osw exhibit greater amplitude fluctuations during MIS 2 which is attributable to the variability of NE monsoon rainfall and associated river discharge into the Bay of Bengal in association with strong seasonal temperature contrast. Onset of strengthening phase of SW monsoon was started during Bølling/Allerød as evidenced by the low $\delta^{18}$Osw values ~14.7 ka (Govil and Naidu, 2011). A conspicuous $\delta^{18}$Osw excursion between 15.8 to 12.8 ka represents highest rainfall and river discharge for the duration of 3 ka after the initiation of deglaciation in the BOB, which is consistent with the rainfall maximum in northern India from 15 to 13 ka as documented in the speleothem $\delta^{18}$O record from Timta Cave (Sinha *et al*., 2010). $\delta^{18}$Osw show consistently lower values during the Holocene (with an exception around 5 ka), which suggests that the freshening of the Bay of Bengal due to heavy precipitation and river discharge caused by the strong SW monsoon (Govil and Naidu, 2011; also see Anand *et al*., 2008; Rashid *et al*., 2007).

Gupta *et al*., (2011) analysed foraminiferal faunal,

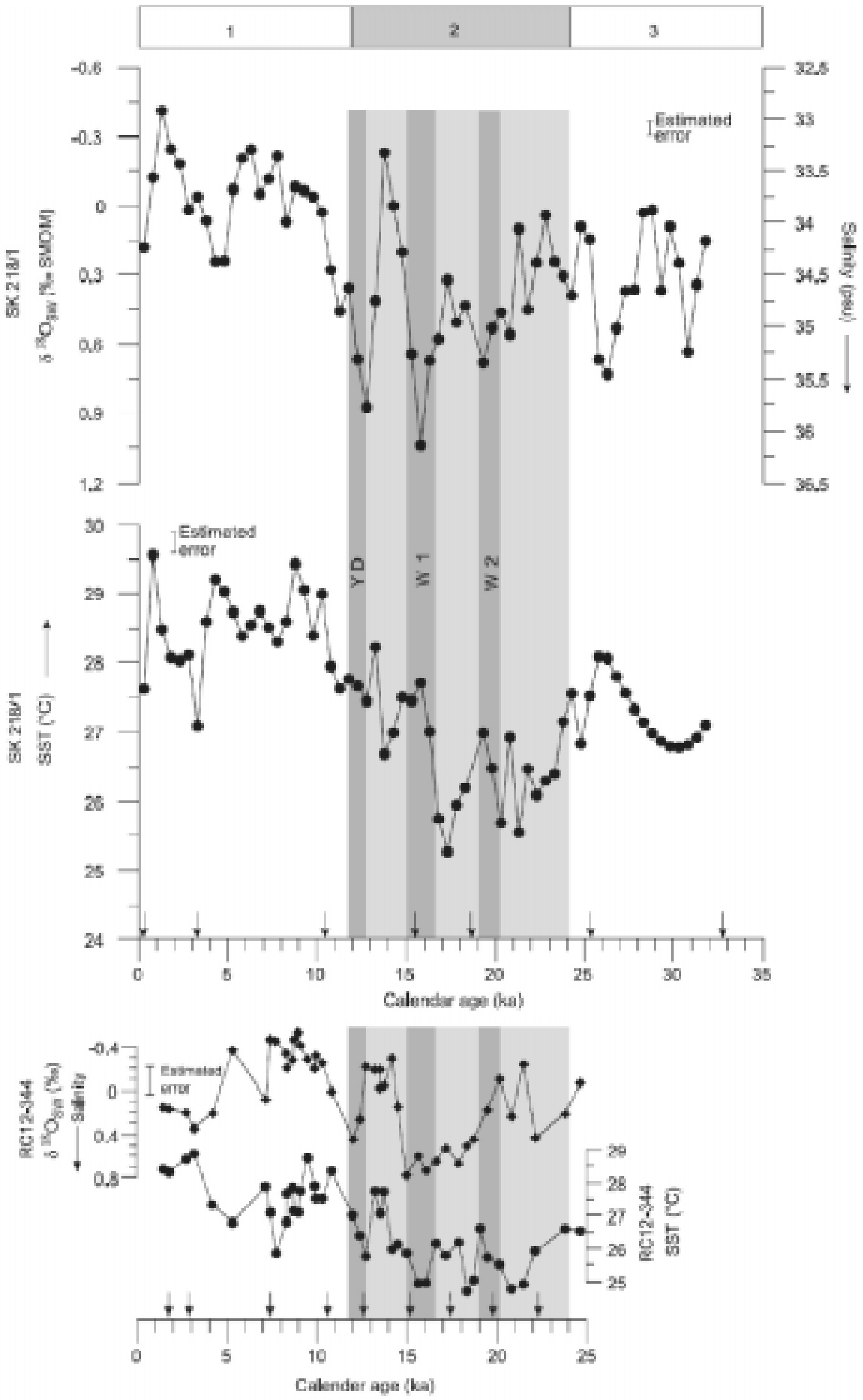


**Fig. 3: Profiles of $\delta^{18}$OSW, salinity and SST in core SK218/1. MIS 2 is shown on grey shade and Younger Dryas (YD) and W1 and W2 are marked with dark bands to show a negative relationship between SST and rainfall (as evident from $\delta^{18}$OSW and salinity) at this site during MIS 2. $\delta^{18}$OSW and SST in cores SK-218/1 are compared with another core (RC12-344) from the Andaman Sea[32]. Arrows pointing to X axis represent tie points of ages derived from C14 AMS dates (one 14C AMS date 36.8 ka falls beyond the scale of the above plot hence it is not shown here). (from Govil and Naidu, 2011and references therein)**

total organic carbon (TOC), Inorganic Carbon and stable isotope data to understand summer monsoon-driven changes in the western and eastern Arabian Sea and their impacts on deep-sea ventilation during the past 21 Kyr. During 14-12.5 Kyr increased TOC values, indicate weaker summer monsoon and weak oxygen minimum zone (OMZ). The TOC values and the abundance of the eutrophic benthic foraminifers species *Bulimina aculeata*, *Melonis barleeanum* and *Uvigerina peregrina* increased during the late Holocene whereas, well-oxygenated, low organic carbon benthic species *Sphaeroidina bulloides* as well as decreasing of pteropods indicate an intense OMZ and an increased supply of refractory organic material to the Arabian Sea (Fig. 4).

Godad *et al*., (2010) observed SST changes during May and August in the western Arabian Sea (WAS) over the last 22 kyr and found that in the WAS, the highest SST difference presently occur between May and August. The SST difference between May and August exhibits three distinct phases: i) a moderate SST difference in the late Holocene (0-3.5 ka) is attributable to intense upwelling during August, ii) a minimum SST difference from 4 to 12 ka is due to weak upwelling during the month of August, and iii) the highest SST difference during the last glacial interval (19 to 22 ka) with high *Globigerina bulloides* % could have been caused by the occurrence of a prolonged upwelling season (from May through July) and maximum difference in the incoming solar radiation between May and August. New evidence on the sequence of deglacial warming in the tropical Indian Ocean, indicates climate variability during the transition from the LGM to the Holocene, characterized by large spatial and temporal variations (Naidu and Govil, 2010).

The River Valleys of Peninsular India preserve a great wealth of palaeoclimate and archaeological data,

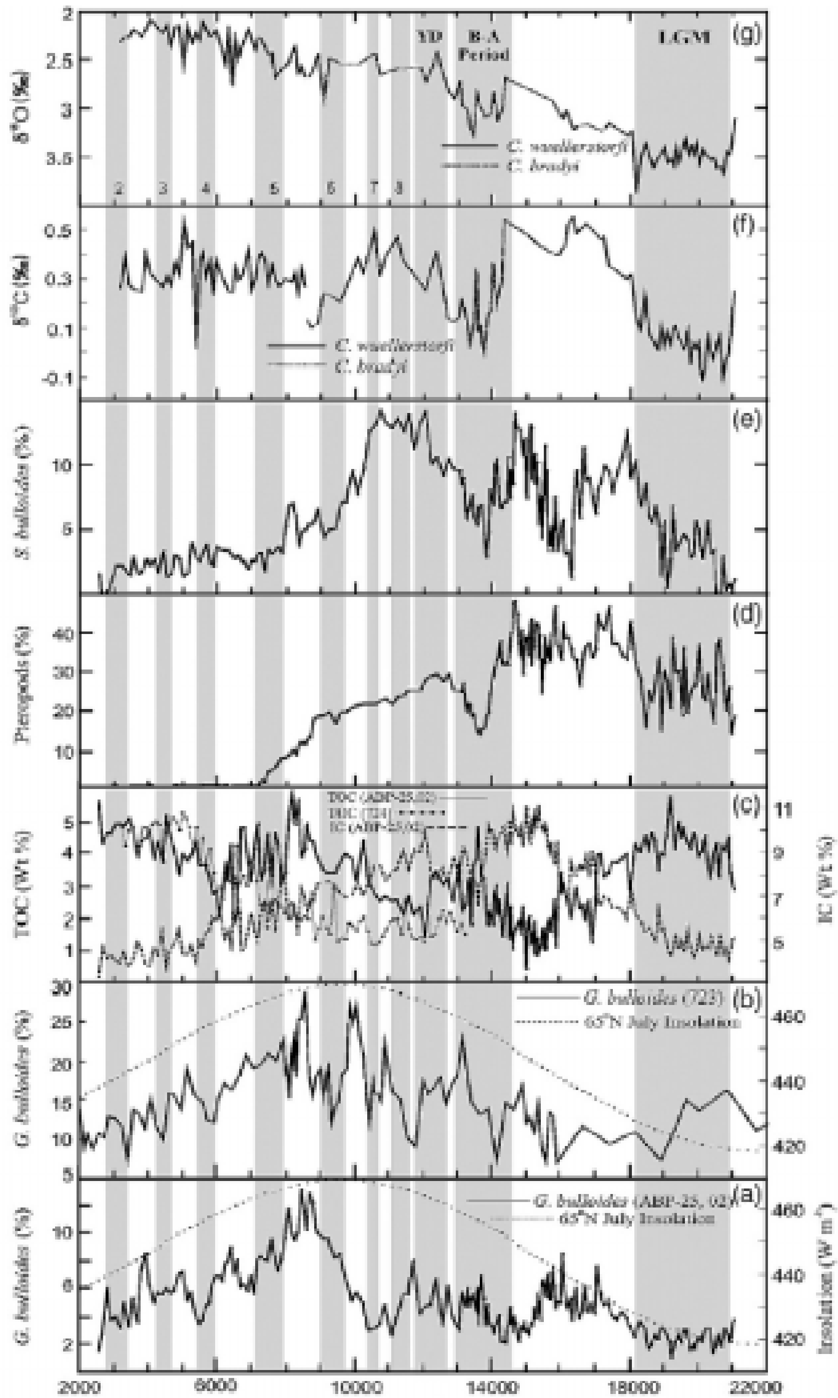


**Fig. 4: Proxy records from Holes ABP-25, 02, ODP 723A and ODP 724B (panels a-g). (a) and (b) Percent *Globigerina bulloides* from Holes ABP-25, 02 and ODP 723A, respectively, superimposed with 65o N July insolation, (c) total organic carbon and inorganic carbon (calcium carbonate) values from Holes ABP-25, 02 and ODP 724B, (d) percent pteropods from Hole ABP-25, 02, (e) percent distribution of benthic foraminifer *Sphaeroidina bulloides* at Hole ABP-25, 02, (f) carbon and (g) oxygen isotope values of benthic foraminifera *Cibicides wuellerstorfi* (continuous line) and *C. bradyi* (broken line) from Hole ABP-25, 02. Grey bars indicate major climate events including the Last Glacial Maximum (LGM), the Bølling-Allerød (B-A) interstadials, the Younger Dryas (YD) and Bond events 2-8 of the Holocene (from Gupta *et al*., 2011and references therein)**

spanning the last 100 ka. The $\delta^{18}O$ data of Ganga Plain show periodic change in rainfall amount between 100 and 18 ka with three peaks of higher monsoon at about 100, 40 and 25 ka comparable with the marine and other terrestrial records (Agarwal *et al*., 2012) (Fig. 5). The vegetation for the time period 84 to 18 ka based on δ13C values of soil carbonate indicate that relative abundances of C3 and C4 vegetations were mainly driven by variations in monsoonal rainfall (Agarwal *et al*., 2012). Magnetic susceptibility and geochemical data of the Goting Lake sediments from the higher central Himalayas indicate moderate to strong monsoon around 25 ka, 23.5 ka-22.5 ka, 22 ka-18 ka, 17 ka-16.5 ka and after14.5-13 ka, whereas at ~22 ka the early LGM climate was mainly arid (Juyal *et al*., 2009). Palynological analysis from Dokriani Valley, Garhwal Himalaya reveals that between 12,406 and 10,633 cal yr BP vegetation was sparse, because of cool and dry climate, attributable to the Younger Dryas event (Bhattacharyya *et al*., 2011a).

**The Holocene**

Holocene record shows millennial-scale weak and strong phases of the Indian summer monsoon. A weak monsoon event occurs at ~8,200 yrs BP between a generally strong early Holocene monsoon period ranging from 10,400 to 5500 yrs, which is followed by a very dry period (Gupta, 2008). Palynological data from Srinagar, southwest Tripura suggest existence of moist deciduous forest under warm humid climate with intermittent change in precipitation around 6.8 and 3.7-3.8 ka (Bhattacharyya *et al*., 2011b). The summer monsoon was weakest from ~2,500 to 1,500 cal yr BP in the late Holocene (Anderson *et al*., 2002; Gupta *et al*., 2003). Analysing *Globigerina bulloides* upwelling records from the Arabian Sea, Anderson et al. (2010) found that the weakest monsoon occurred 1500 yr BP, with an increasing trend towards the present (Fig. 6). Chauhan *et al*., (2010) observed weak summer monsoon ca. 1800-2200, 1000, 650-450 cal. a BP. During the Medieval Warm Period (AD. 900-1400), the summer monsoon strengthened (Gupta *et al*., 2003) whereas during the Little Ice Age (AD. 1450-1850), there was a drastic reduction (Gupta *et al*., 2003). Chauhan *et al*. (2010), however, noticed a humid climate and enhanced precipitation during the terminal stages of the Little Ice Age. The medieval warming (ca. AD 800-1300) is not synchronous either, and is punctuated by an arid event centred at 1000 a BP. High resolution speleothem oxygen isotope records from central and northeast India indicates that on centennial timescales (AD 1400-2008), precipitation variability from these two regions display contrasting pattern, fluctuating between periods with a persistently "active-dominated" (AD ~1700 to 2007) and a "break-dominated" (AD 1400 to ~1700) system (Sinha *et al*., 2011). Isotopic study of speleothems from Andaman Islands indicates that the summer monsoon few millennia ago was similar to the Present and the southwest monsoon was the main source of annual precipitation (Lashkar *et al*., 2011).

Over millennium long tree-ring chronology of Himalayan pencil juniper (*Juniperus polycarpos*), with centennial-scale variation, developed from Lahaul-Spiti, Himachal Pradesh exhibited strong negative correlation with mean regional summer May-August (MJJA) temperature series. Existence of such a strong relationship helped to developed mean summer temperature (May-August (MJJA)) back to AD 940, the period up to which the chronology contained sufficient sample replication (Yadav *et al*., 2009) (Fig. 7). The reconstruction showed protracted period of warm anomalies from $10^{th}$-$16^{th}$ centuries. This provides the state of summer temperature over global climate event of the Medieval Warm Period. Though, the mean summer temperatures decreased since the $15^{th}$ century, $18^{th}$ and $19^{th}$ centuries displayed very cool conditions in the western Himalaya in context of the last millennium. This cool period is consistent with the Little Ice Age period, is coherent with the expansion of glaciers in the western Himalaya and decreased Indian summer monsoon rainfall. The reconstructed mean summer temperatures showed increasing trend since the late $19^{th}$ century. However, the recent warmth could partly be underestimated due to the decrease in tree growth and mean summer temperature relationship in latter part of the $20^{th}$ century. This temperature record is further being updated with the development of more predictor chronologies from the region.

Recently, Yadav (2011a, b) developed tree-ring chronology of Himalayan cedar from the cold arid region of Lahaul and Spiti, Himachal Preadesh (also see Singh *et al*., 2009), India showing low-frequency variations and was used to develop August-July precipitation back to AD 1330 (Fig. 8). The reconstruction shows dominant signal of winter-spring precipitation as over 70% of the annual precipitation in Lahaul-Spiti region occurs during November-May, largely due to western disturbances. The reconstruction revealed very dry conditions during the $14^{th}$ and $15^{th}$ centuries, whereas the $20^{th}$ century was the wettest in context of the past seven centuries.

The long-term wet (dry) phases in the reconstructed precipitation series showed strong coupling with El-Nino (La-Nina) events. However, decoupling in ENSO/ precipitation relationship noted during 1930s-1950s has serious implications for hydrological predictability in cold desert region of the western Himalaya. This reconstructed precipitation series showed negative relationship with summer monsoon rainfall over central northeast India, which in turn shows that the long-term high-resolution precipitation records developed from the cold arid region

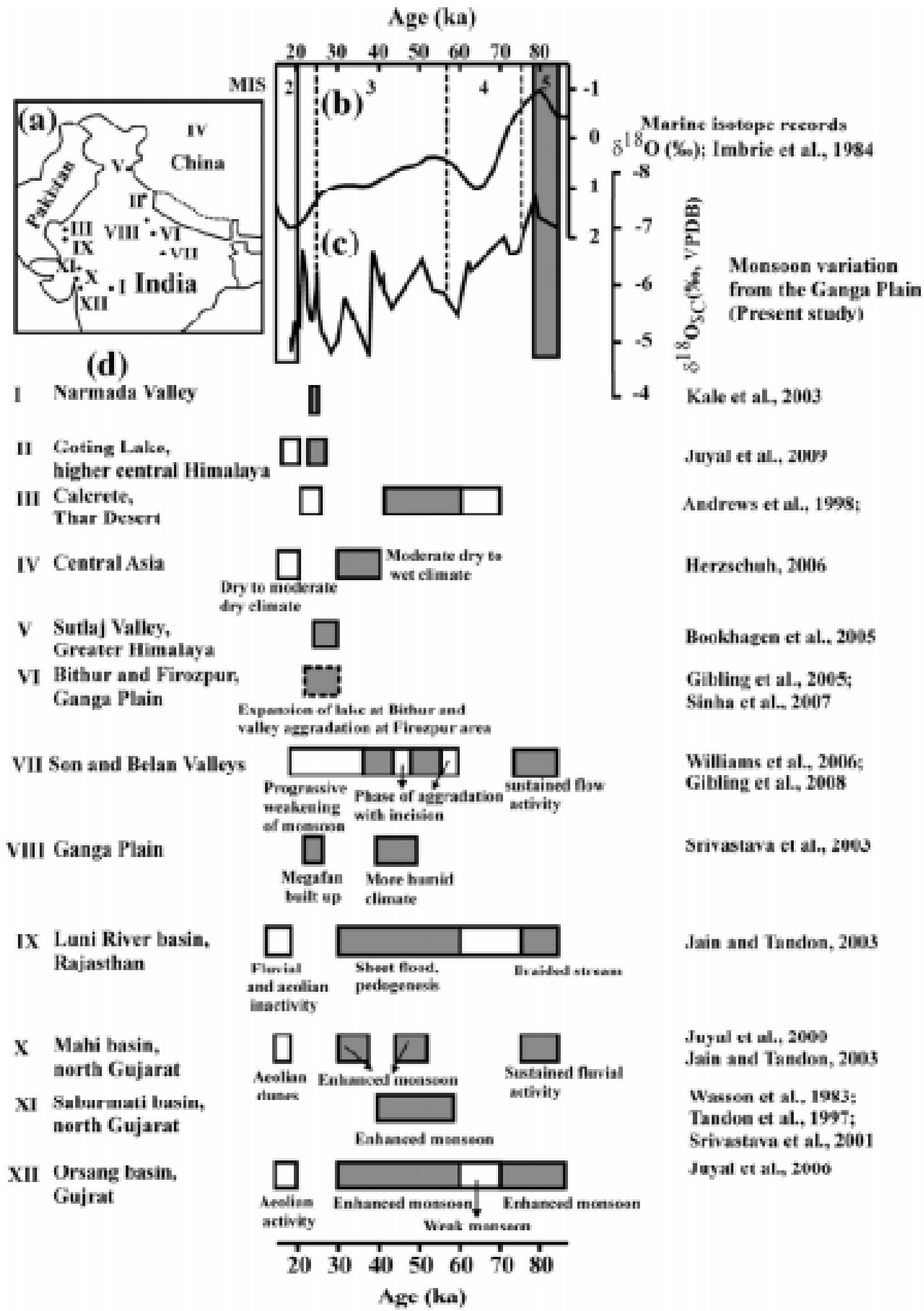


**Fig. 5:** **(a) The inset map shows spatial distribution of palaeoclimatic studies (numbers indicated refer to the records, see Fig. 7d) from different parts of the Indian subcontinent and Central Asia. (b) MIS in relation to (c) δ$^{18}$O values of soil carbonate from the Kalpi core of the Ganga Plain. In MIS, overall increase in oxygen isotope ratio with time is well-correlated with the δ$^{18}$Osc value change in the Kalpi core. A detailed comparison shows that the 84 ka monsoonal high (gray shade) and 18 ka monsoon low (white shade) are synchronous. (d) Sedimentary and geochemical records of the Indian subcontinent and Central Asia are compared with (c) monsoonal rainfall curve generated from the Ganga Plain. This comparison shows sedimentary sequences and geochemical proxies responded to high-amplitude climate changes (gray rectangles stand for high monsoon condition, white rectangles for weak monsoon and for dashed gray rectangle see section correlation of monsoonal rainfall with sedimentary sequence of the Indian subcontinent) (Agarwal *et al*., 2012 and references therein)**

of the western Himalaya should be useful in understanding the monsoon behavior in a long-term perspective. Recent studies (Shah *et al*., 2007; Ram *et al*., 2008; Managave *et al*., 2011) show that variations in ring-widths of teak can be used to reconstruct past monsoon rainfall. Bhattacharyya *et al*. (2007) have shown mean vessel area of the early wood in teak from southern India is correlated with the north-east (NE) monsoon rainfall of the previous year and based on this relation reconstructed past NE monsoon rainfall. Based on high resolution $\delta^{18}O$ composition of teak trees Managave *et al*. (20010 a, b) have demonstrated the possibility of reconstructing sub-annual monsoon rainfall.

## Concluding Remarks

Marine and terrestrial proxy data generated on Indian monsoon, in the last four-five years reveal both temporal and spatial variability. Monsoon, which is characterized by a long dry period interspersed with a rather short periods of heavy rainfall, cannot support rainforests on its own, which require rain throughout or most part of the year. Therefore, presence of tropical-subtropical rainforest in the Palaeogene of northwest and northeast Indian cannot explicitly point to the existence of monsoonal climatic conditions. Early Miocene Himalayan exhumation again suggest heavy rainfall, but seasonality, monsoon is known for, becomes evident only in the Late Miocene marine and terrestrial records. Pliocene monsoon is greatly influenced by the Northern Hemisphere Glaciation, where as that of Pleistocene is influenced by glacial-interglacial alterations,

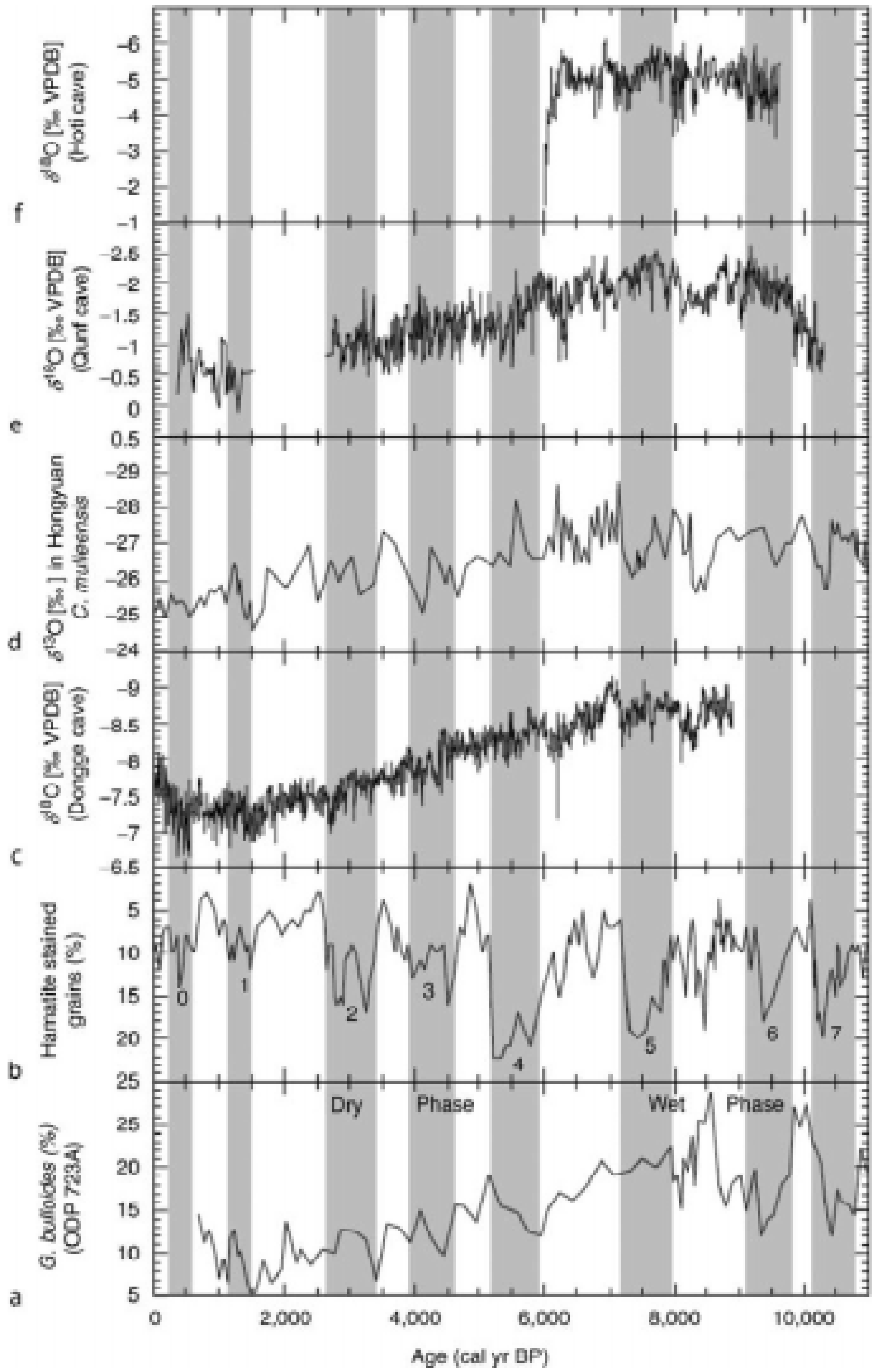


**Fig. 6: Visual correlation between proxy records. (a) *G. bulloides* percent from ODP Hole 723A, Arabian Sea[39]; (b) hematite % in core MC52 in the North Atlantic; (c) oxygen isotope record from Dongge Cave, southern China; (d) Hongyuan peat deposit, (e) Qunf Cave δ18O and (f) Hoti Cave δ18O Grey bars highlight the Bond events 0-7 (from Gupta, 2008 and references therein)**

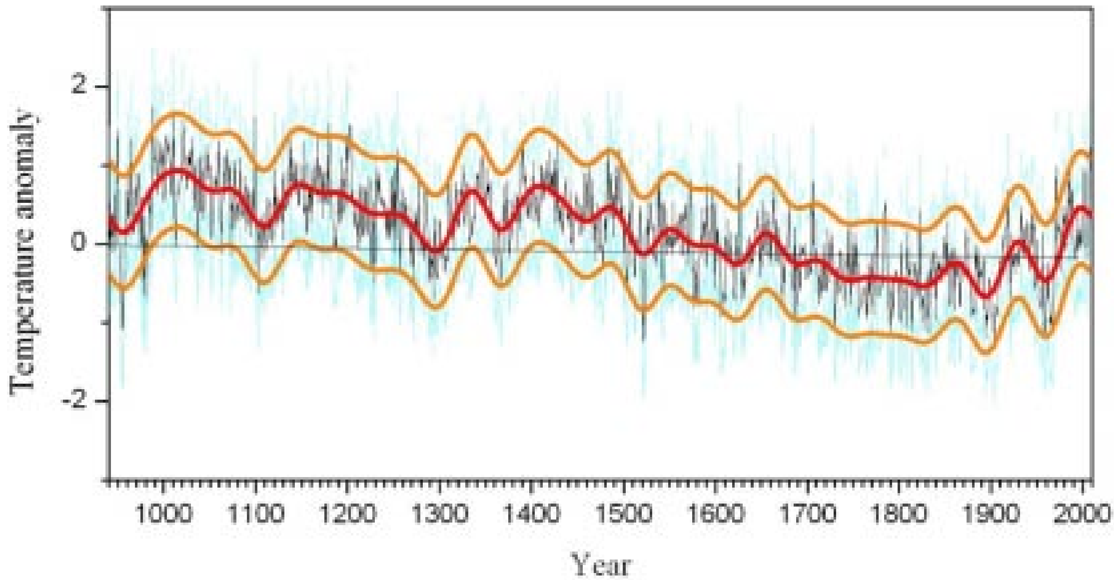


**Fig. 7:** **Mean annual summer (MJJA) temperature reconstruction for the western Himalaya (AD 940-2006). Reconstruction as well as lower and upper one standard errors were smoothed using 50-year low pass filter (Yadav *et al*., 2009)**

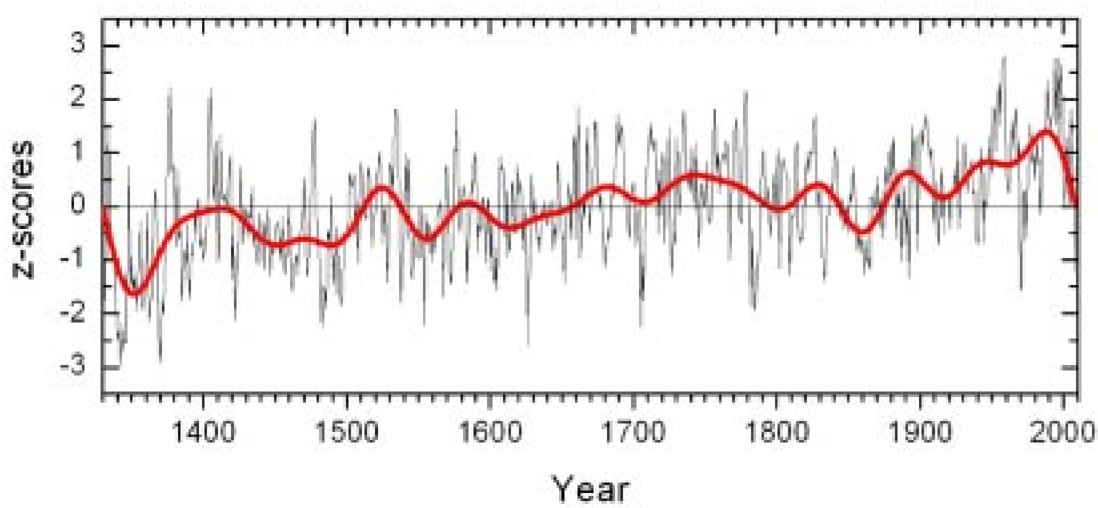


**Fig. 8:** **August-July precipitation reconstruction (AD 1330-2008) for Lahaul-Spiti region, Himachal Pradesh using tree-ring data network of Himalayan cedar. The thick line represents the 50-year low pass filter (Yadav, 2011a )**

with major shifts during 700 to 600 ka and 300 to 250 ka. The last 100 ka saw a strengthening of the southwest monsoon during the transition from MIS 4 to MIS 3, with precipitation being higher during MIS 3 and MIS 1 and lower during MIS 2 and MIS 4, with three peaks of higher monsoon at about 100, 40 and 25 ka and a very weak southwest monsoon during the Last Glacial Maximum. Since the LGM greater spatial variability among terrestrial and marine record has been recorded, for example, in Bay of Bengal 15.8 to 12.8 ka represents high rainfall, whereas in the Western Arabian Sea, 14-12.5 ka had a weaker summer monsoon, the Himalayan lake data indicate strong monsoon around 25 ka, 23.5 ka-22.5 ka, 22 ka-18 ka, 17 ka-16.5 ka and after14.5-13 ka. The Holocene saw millennial-scale monsoon variability, beginning with a strong phase till the mid Holocene, followed by weak phases at ~2,500, 2200-1800-2200, 1,500, 1000, 650-450 ka. In the last millennium, centennial scale monsoon variability display a "break-dominated" period ranging from 1400 to ~1700 AD and an "active-dominated" period from ~1700 to 2007 AD, 10th to 16th centuries exhibit, prolonged period of warm phase, interspersed with dry 14th and 15th centuries, 18th and first part of the 19th were particularly cold, had weak summer monsoon, followed by wetter conditions in the 20th century.

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