

## **Reconstruction of the variability of SW monsoon during the past 3 Ka from the continental margin of the South Eastern Arabian Sea**

Onkar S. Chauhan<sup>1</sup>, Elke Vogelsang<sup>2</sup>, Nathani Basavaiah<sup>3</sup>, U. Syed Abdul Kader<sup>1</sup>

1. National Institute of Oceanography, CSIR, Dona Paula, 403 004 Goa, India
2. Institut für Geowissenschaften, Universität Kiel, Olshausenstrasse 40, D - 24118 Kiel, Germany
3. Indian Institute of Geomagnetism, Kalamboli, New Panvel, Navi Mumbai 410218, India

### **Abstract**

From temporal variation in  $\delta^{18}\text{O}$  in *G. ruber* and *G. sacculifer* and geochemical indices of weathering/erosion (CIA, Al and Ti), we infer rapid Southwest Monsoon deterioration with dwindled fluvial and detrital fluxes during ~450-650, 1000, and 1800-2200 cal y BP during the Late Holocene. We have evaluated the role of solar influx (reconstructed) and high latitude climate variability (archived in GRIP and GISP-2 cores) on the Southwest monsoon precipitation (SWM). Broadly, our  $\delta^{18}\text{O}$  climate reconstruction is concordant with GRIP and GISP-2, and supports a teleconnection through atmospheric connection between the SWM and the North Atlantic climate -albeit temporal extents of LIA and MWP from high latitude are not entirely coeval. Moreover, there is humid climate and enhanced precipitation during terminal stages of the Little Ice Age. The medieval warming (~800-1300 AD) is not synchronous also, and is punctured by an arid event centered at 1000y BP. Though the delineation of the specific influence of solar influx on the SWM precipitation is elusive, we surmise that the SWM precipitation is a complex phenomenon and local orography along the SW India may have a role on the entrapment of moisture from the SW trade winds, while these brace land.

## Introduction

Long term variations in the Asian monsoon, the major source of moisture for the Indian Subcontinent, have been attributed to Milankovich orbital parameters; mountain plateau orography; atmospheric CO<sub>2</sub>; and glacial surface boundary conditions etc. (Prell and Kutzbach, 1992). In alternative thesis, solar forcing has been postulated to cause rapid climatic instabilities in the monsoonal regime during past two millennia (Eddy, 1976; Bond *et al.*, 2001; Haigh, 2001; Agnihotri *et al.*, 2002; Rind, 2002; Gupta *et al.*, 2003; Sinha *et al.*, 2007). Influence of ENSO through Walker circulation is also suggested (Hasternath and Greichar, 1993; Tiwari *et al.*, 2006; Saraswat *et al.*, 2007). It is postulated further that the monsoon regime has a teleconnection with that of the climate of the North Atlantic (Sirocko *et al.*, 1996; Gupta *et al.*, 2003). However, documentation and linkage of abrupt, high frequency instabilities in the SWM precipitation with that of internal as well as external forcing is still in the state of infancy. One of the reasons contributing to debilitate in such studies is scanty, high resolution, continuous records of climate variability with decadal – centenary scale resolution from the Arabian Sea. Some few millennia precipitation history is available (Sinha *et al.*, 2007; Yadava *et al.*, 2004; Tiwari *et al.*, 2005), though these records are either short - mostly for less than a millennium or are not continuous (Staubwasser *et al.*, 2003; Tiwari *et al.*, 2005). A five ka history of fluvial influx of the River Indus has been reconstructed (von Rad *et al.*, 1999), Staubwasser *et al.* (2003) have also provided some insight in the climate variability since the Mid Holocene. The Indus has large catchment and hinterland in semi-arid region. Considerable flux of the Indus is known to advect along-shelf and sinks inland (Chauhan *et al.*, 2006). Being fed from the Himalayan ice melt as well as precipitation in the hinterland, its discharge is governed by a complex phenomenon.

Conventionally, the Western Arabian Sea, having dominant evaporation over precipitation - even though does not receive precipitation, has been a host of studies for estimating SWM intensity from upwelling or wind strength (Overpeck *et al.*, 1996; Sirocko *et al.*, 1996; Schulz *et al.*, 1998; Anderson *et al.*, 2002; Gupta *et al.*, 2003). However, there exists no clue whether such proxies of climate reconstruction are coeval with the SWM precipitation in the other climatological regime with dominant precipitation over evaporation. We have attempted to decipher variations in the SWM precipitation during past three millennia from orographically enhanced precipitation (interception of the southwest trade wind by the hilly region, elevated about 2000-3000 m above MSL, known as the Western Ghat) along the SW continental margin of India, carried as fluvial flux into the sea influencing regional sea surface salinity (SSS). Even though our core is not laminated, a high sedimentation rate and sub-century

scale resolution of our data have helped us to assess if there exists any link among the SWM precipitation, the multi-decadal the North Atlantic climate instabilities (GRIP and GISP-2 cores) and reconstructed variability in solar irradiance.

### **Modern climatology and Hydrography**

The South East (SE) Arabian Sea, in contrast to its western counter part, has seasonal sea surface salinity (SSS) variation regulated by the SWM precipitation (Fig. 1). Inflowing into the Indian subcontinent, the SWM trade winds braces landmass here. Except during SWM, the region receives insignificant and episodic rainfall. During SWM (June-September) in the coastal region precipitation is moderate (150 cm)., Because of enhanced moisture condensation by orographic influence, inland precipitation enhances to ~500 cm in the Western Ghats, the hilly region adjacent to coastal plains (Chauhan and Gujar, 1996; Fig. 1). The SE Arabian Sea, therefore, receives high freshwater influx through several short, high gradient rivers; important among these are the Kallada, the Karmana, the Neyyar and the Tamraparni (Fig. 1). Except for the Kallada, these rivers do not have an estuarine mouth (Table 1; Aswathanarayana, 1964; Mallik *et al.*, 1989; Nagam Aiya, 1989) and the fluvial flux is directly carried into the sea. Lower salinity has been observed in the vicinity of the core site (June- January; values 34-33‰; Levitus and Boyer, 1994; Levitus *et al.*, 1994; Fig. 1). In contrast to the wind driven upwelling regions of the Western Arabian Sea, our study area has dominance of precipitation over evaporation. The SSS is significantly modified by the SWM precipitation, though there is no distinct SST instability associated with the SWM winds (Levitus and Boyer, 1994; Levitus *et al.*, 1994). The hydrography of the area is seasonal. Polewards surface circulation during the North Eastern Monsoon (NEM) reverses and turns equatorwards during SWM.

### **Materials and Methods**

From the low SSS regions off the precipitation dominated the SW continental margin of India (June-January; values 33-34‰ Fig. 1; Levitus and Boyer, 1994; Levitus *et al.*, 1994), during various cruises of the R.V. *Gaveshni* and ORV *Sagar Kanya*, 23 cores were collected. Among these, Core 3904 only (Latitude 08° 07.9' N; Longitude 76° 01.8' E; water depth 1400 m; Fig. 1) was found to be: (i) suitable for deciphering the sub-century scale events on account of the highest rate of sedimentation, (ii) located in the area in which, as determined from the available data, the overlying waters are anoxic and bioturbation is expected to be minimal due to high productivity (Sarkar *et al.*, 2000; Qasim, 1977), (iii) entirely turbidity free, and (iv) regional SSS in this area is influenced by SWM.

The upper 58 cm of the core was sub-sampled (44 levels). For  $\delta^{18}\text{O}$  ( $^{18}\text{O}/^{16}\text{O}$ ) sample/ $(^{18}\text{O}/^{16}\text{O})_{\text{standard}} - 1$  measurements, about 25-30 tests (size 315-400  $\mu\text{m}$ , crushed, ultrasonic cleaned in ethanol) of *G. ruber* (s.s. white), *G. sacculifer* (without sac), and *G. menardii* were analyzed on a VG Micromass against the Peedee Belemnite (PDB) standard at the University of Kiel. The results have a standard deviation of 0.04‰ for oxygen and 0.05‰ for carbon isotopic analyses.

The samples were wet sieved to separate out +63  $\mu\text{m}$  fractions, while other fractions (+ 45, 25, 15, and 4  $\mu\text{m}$ ) were separated using standard pipette methods (Folk, 1966). Associated with high productivity in the region (Qasim, 1977), we have found high shell contents in +63 and +45  $\mu\text{m}$  fractions, which are likely to influence estimation of continental Ca and terrigenous flux. We have treated bulk samples with 10% acetic acid to remove the biogenic carbonate. Chemical analyses are made on treated samples on a Philips XRF (model PW 1480) at UCSI, Delhi. The precision is obtained against standard SDO-1-3, SCO-1 and MAG-1, and it is better than  $\pm 5\%$ . Chemical Index of Alter (CIA) is obtained from  $\text{CIA} = \text{molar ratio of } \text{Al}_2\text{O}_3 / (\text{Al}_2\text{O}_3 + \text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}) \times 100$  (Nesbitt and Young, 1982).

Al and Ti are viewed as resistant elements, and an indicator of detrital flux (Juyal *et al.*, 2009). Besides detrital contribution, Aluminum may be contributed from scavenging, volcanic dust and aeolian sources in the marine environment (Murray *et al.*, 1993; Murray and Leinen, 1996; Pattan and Shane, 1999). In order to nullify any effect of these, we have estimated structurally unsupported detrital terrigenous fraction from following relation (Pattan and Shane, 1999):

$$\text{Al}_{\text{exc}} = \text{Al}_{\text{total}} - (\text{Ti}_{\text{sample}} \times (\text{Al}/\text{Ti}_{\text{shale}})) \text{ Where } \text{Ti}_{\text{shale}} = 0.6$$

The excess Al, even though is found to be very negligible, has been subtracted from the total Al content to obtain detrital Al.

## **Chronology**

For most of the studies in the Arabian Sea, ages are measured at irregular intervals of several tens of cm (Anderson *et al.*, 2002; Gupta *et al.*, 2003; Tiwari *et al.*, 2005, 2006; and references cited therein). The events are then assigned ages by extrapolation, assuming a constant rate of sedimentation between these levels. However, due to variations in the environment of deposition, variable productivity and terrigenous and biogenic flux, sedimentation is unlikely to be linear. The points of measurements,

though presumed, may not be actual levels on which the sedimentation rates actually altered in a sediment column. Therefore, for most of the paleoclimatic reconstruction from this region, specifically in the areas of non-linear sedimentation, the age models by and large have poor chronometric control.

In order to have better chronometric control, four AMS ages (Poznan radio carbon laboratory, Poland) were obtained on *G. ruber* (250-415  $\mu\text{m}$  size) to specifically date the pronounced instabilities (Table 2). In addition 8 bulk  $^{14}\text{C}$  ages obtained at each 4-6 cm interval were also used to refine the chronology (Table 2; Fig. 2). Because bulk ages may suffer from contamination by older carbonates or from bioturbation, we obtained ages on hand picked planktonic foraminifers to nullify any influence of land derived contamination. However, due to dearth of sufficient biogenic material, we have combined two slices of 1 cm for age determination. Ages are converted to calendar years for construction of the age model (Fig. 2) after standard correction for the reservoir ages; (Stuiver and Braziunas, 1993). Because of high productivity, prevalence of anoxia, a high rate of sedimentation, and negligible bioturbation (Sarkar *et al.*, 2000), it is expected that the fine resolution climate signatures are not homogenized significantly.

## Results

The temporal variations in  $\delta^{18}\text{O}$  *G. ruber* and *G. sacculifer* are well pronounced and concordant (Fig. 3). In general,  $\delta^{18}\text{O}$  varies between  $-2.3\text{‰}$  and  $-1.64\text{‰}$  for *G. sacculifer* and  $-2.87$  and  $-2.25\text{‰}$  for *G. ruber* (Fig.3). These values for *G. ruber* are similar to the regional values reported earlier (Duplessy *et al.*, 1981; Duplessy, 1982). We have found heavier excursion centered at 450-650 ( $-2.49\text{‰}$ ; Arid I),  $\sim 1000$  y ( $-2.29\text{‰}$ ; Arid II), and 2200-1800 ( $-2.25\text{‰}$ ; Arid III), which are about 0.24, 0.45 and 0.51 $\text{‰}$  heavier for *G. ruber*. *G. sacculifer* depicts a concordant trend, although the event of 2200-1800 only is pronounced in *G. menardii* (Fig. 3). During the intervening periods we have observed persistently lighter incursions ( $-2.87 \text{‰}$  at 3000-2300;  $-2.76\text{‰}$  at 1800-1100;  $-2.81\text{‰}$  at 900-650 and  $-2.78\text{‰}$  at  $<400$  cal y BP). The variations in CIA, Al and Ti are concordant with  $\delta^{18}\text{O}$  variability (Fig. 3). Al varies between 3.06 and 5.74%. During all heavier  $\delta^{18}\text{O}$  incursion episodes I – III, a reduced abundance of Al is found (3.06-3.72%), though its content is enhanced during 3000-2300, 1800-1100, 900-650, and  $<400$  y BP (3.5-5.74%; Fig. 3). In general, CIA is found to vary between 60.93 and 74.36%. An enhanced CIA during 3000-2300 (71.4%), 1800-1100 (72.83%), 900-650 (74.11%) and  $<400$  y (74.36%) has persistent reduction during 650-450 (60.93%),  $\sim 1000$  (64.18%) and 1800-2200 (64.54 %) cal y, Ti shows the least variability (range 0.51 to 0.64%; Fig. 3).

## Discussions

### Climate reconstruction from the Oxygen isotope and chemical proxies

We have found reduced (enhanced) CIA and Al during all the events of heavier (lighter) incursion of  $\delta^{18}\text{O}$ . CIA has been considered as an indices of weathering and erosion (Nesbitt and Young, 1982; Colin *et al.*, 1998). However, there exist few clues on response time of these proxies to the climate variability. We rely on the studies of Warrier and Shankar (2009), who have demonstrated decadal scale variability in chemical weathering regime in response to climate variability at the SW continental margin of India. Multiproxies data of  $\delta^{18}\text{O}$ , CIA, Al and Ti are, therefore, used for climatic reconstruction.

During the Holocene, SST variability is rather negligible ( $0.05^{\circ}\text{C}$ ; Rostek *et al.*, 1993; Rolinski *et al.*, 2001), and is not likely to influence  $\delta^{18}\text{O}$  signature significantly. The global ice meltdown is negligible during past 3 ka (Fairbanks, 1989) and changes in SSS due to global meltdown may be considered insignificant in the SE Arabian Sea.  $\delta^{18}\text{O}$  variability in our core may, therefore, be attributed to the magnitude of fluvial discharge regulated by the SWM precipitation. *G. ruber* and *G. sacculifer*, are surface dwelling species (Hilbrecht *et al.*, 1996; Guptha *et al.*, 1997; Tiwari *et al.*, 2005), and are expected to archive fluvial flux induced changes in surface salinity. Because *G. menardii* dwells into the deeper waters (about 100-150 m; Hilbrecht *et al.*, 1996), its  $\delta^{18}\text{O}$  may not be a sensitive indicator of surficial, shallow waters salinity variations. We have evaluated, therefore,  $\delta^{18}\text{O}$  variability in surface dwelling species in order to archive SSS variability.

In the Arabian Sea, 0.33‰  $\delta^{18}\text{O}$  changes are found to be related with 1 psu changes in SSS (Duplessy *et al.*, 1981; Sarkar *et al.*, 2000). The heavier incursions of  $\sim 0.45\text{‰}$  and  $0.51\text{‰}$  have been observed in the  $\delta^{18}\text{O}$  of *G. ruber* during the “Arid Events” II and III respectively (Fig. 3), which suggest  $> 1$ psu changes in SSS due to persistent reduction in the fluvial influx. The heavier incursion is less pronounced ( $\sim 0.24\text{‰}$ ) but persists also during the episode I. This implies that there have been pronounced century scale instabilities in the SWM precipitation during 450-650,  $\sim 1000$ , 1800-2200, cal y BP. Synchronized with heavier  $\delta^{18}\text{O}$ , a decline in CIA, Ti, and Al is observed during these Arid Events I-III, which is interpreted as prevalence of a weaker SW monsoon during 2200-1800,  $\sim 1000$  y, and 650-450y cal BP. From the lighter  $\delta^{18}\text{O}$  values and from enhanced Al and CIA during 3000-2300, 1800-1100, 900-650 and  $< 400$  y BP, we infer an amelioration in the magnitude of the SW

monsoon during these times. It is surmised, therefore, that during the Late Holocene there were several rapid phases of significant variability in the SWM precipitation.

### **Comparison with other continental and the Western Arabian Sea records**

Because of paucity of any other suitable turbidity free core with high chronometric resolution, in order to obtain a regional trend in climate variability during the past 3 ka, we have compared our results with available marine as well as continental records from the Indian Subcontinent ( Fig. 6; von Rad *et al.*, 1999; Anderson *et al.*, 2002; Gupta *et al.*, 2003; Tiwari *et al.*, 2005, 2006; Sinha *et al.*, 2007; refer Fig. 1 for their location). For 2800 y cal BP Tiwari *et al.* (2006), and for past 800 AD (~1150 y BP), Tiwari *et al.* (2005) have obtained  $\delta^{18}\text{O}$  variability in surface dwelling *G. ruber* and *G. sacculifer* from the farther north location (Central Eastern Arabian Sea; Fig. 1) in an AMS dated core. This region lies also in the area in which a high SWM precipitation influences SSS (Fig. 1). Although these studies do not cover full span of our studies, there are heavier incursions centered at 2000, 1110 and 550 y cal BP, which are synchronous with our reconstruction within chronometric errors of extrapolation between larger core levels (Fig. 6). Based upon varved chronology, von Rad *et al.*(1999) documented high fluvial discharge of the River Indus at 900, 1200-1700, and 2600 yr BP, which are concordant)with higher riverine influx in our studies (Fig. 6). Gupta *et al.* (2003) from wind induced upwelling indices *G. bulloides* have observed also a coeval warming off Oman during 600-1200 BP, though an enhancement of monsoon during LIA is found by Anderson *et al.* (2002), synchronous within chronometric uncertainties (assigned ages) to our studies (Fig. 6).

The humid phase of 2300-3000 BP, though concordant in the Indus discharge (von Rad *et al.*, 1999), is, however, at variance to a long drought in the Thar Desert, India between 3600-2000 yr BP (Swain *et al.*, 1983; Bryson, 1996) or aridity and weakening of the summer monsoon between 4500-2000 y BP (Swain *et al.*, 1983) in the continental records (Fig. 6). We attribute this to different local response to monsoonal forcing. The Thar Desert, under modern climatology, lies in an extreme temperature belt with scanty rainfall (< 50 cm of annual rainfall compared to ~500 cm in the study area) in the terminal path of SWM. An erratic marginal increase in the precipitation in this arid region can possibly not alter the vegetation pattern or pollen flux, which is the basis of such climate reconstruction (Swain *et al.*, 1983). The observed discordance therefore stems from the archival efficiency of climatological proxies. Along the SW continental margin of India, precipitation intensity is the basis of

SWM reconstruction, and is perhaps not comparable with a secondary vegetation response manifested in the pollen flux into a lake located in a semi-arid, high- temperature desert region.

The lighter incursion of  $\delta^{18}\text{O}$  in  $< 400$  y BP is coeval with the other records of an enhanced SWM from the SE Arabian Sea (Fig. 6; Anderson *et al.*, 2002; Gupta *et al.*, 2003; Tiwari *et al.*, 2005). In the yearly resolved  $\delta^{18}\text{O}$  of speleothem, Yadava *et al.* (2004) have found also the highest rainfall at 1666 AD and it endorses our reconstruction. Concurrent regional variability in  $\delta^{18}\text{O}$  and other proxies synchronous to our reconstruction, therefore, endorse regional climate variability in the Indian Subcontinent during the Late Holocene (Fig. 6).

### **Comparison with climate of other regions**

Important climate instabilities at the high altitude during the Late Holocene are Little Ice Age (LIA) and Medieval Warm Period (MWP). In order to estimate a linkage between regional climate of the Arabian Sea and that of the High Latitude and the Europe, we have compared our reconstruction with available climate variability from these regions (Fig. 6). Crowley and Lowery, (2000) have described LIA as mild cooling in the Northern Hemisphere. Hu *et al.* (2001) have found it  $\sim 1.7^\circ\text{C}$  cooler in the Alaska, though its termination is found to vary (in the Sargasso Sea at 400 BP; Keigwin, 1996). LIA, therefore, has a large temporal and spatial variability across the globe (IPCC report, 2001). Cronin *et al.*, (2001, 2003) have found several 100 y incursions in the Chesapeake Bay during 2100, 1600, 950, 650, 400 y BP which have similar century scale climatic instabilities observed in SWM in our studies. Large temperature excursions ( $2\text{-}4^\circ\text{C}$ ) during MWP (about 800–1300 AD) and during LIA (about 1350–1850 AD; Cronin *et al.*, 2001, 2003; Wanner *et al.*, 2008) are also archived in the Chesapeake Bay, and these have been linked with changes in the strength of the North Atlantic thermohaline circulation (Diamond, 2005). We document a phase with declined precipitation during MWP, and a wetter phase during terminal stage of LIA (Fig. 6), and reaffirm that LIA is not synchronously distributed in the Tropics, the high latitude and the North Atlantic Ocean. Even though Zonneveld *et al.* (1997) have suggested a rapid teleconnection between the climate of high latitude with SWM through atmospheric connection, there exists no clue if such response is synchronously archived in sub-century scale orographically induced SWM precipitation. Variable chronometric resolution, a poor understanding of the response timing and the archival accuracies of the land and the marine proxies prelude to elucidate the factors influencing these intraregional climate variability.



## Correlation with GRIP and GISP-2 cores data

We have compared climate instabilities in our data with that of GRIP and GISP-2 ice core to decipher a possible linkage between them (Grootes and Stuiver, 1997; Vinther *et al.*, 2006). During the Late Holocene, the  $\delta^{18}\text{O}$  incursions in GISP-2 and GRIP have far too less amplitude compared to LGM – the Upper Holocene, although there is distinct high frequency  $\delta^{18}\text{O}$  variability (range -33.98- -35.72‰) corresponding to temperature variability of -32.18 - -29.57°C during last 3150 y BP (Alley, 2004). Our data have sub-century scale resolution while GRIP and GISP-2 data have annual resolution. Irrespective of the thesis of teleconnection of SWM with high latitude climate through rapid atmospheric connection (Zonneveld *et al.*, 1997), high resolution annual climatic variability of GRIP and GISP-2 data may not be fully mirrored in our core because of inherent archival limitation dictated by uncertain response time of land and marine proxies. However, in order to aid a comparison between both series (Vinther *et al.*, 2006), we have smoothed and 3- point averaged GISP-2 and GRIP data using Fast Fourier Transformation (FFT) filtering (Cooley and Tukey, 1965; Duhamel and Vetterli, 1990). For peak type data, optimum smoothing level is generally between 25-40% (Duhamel and Vetterli, 1990; Unser *et al.*, 1993). We have used 25% smoothing. Unfiltered as well as smoothed time series data are presented (Fig. 4).

Varying within limits of above referred chronometric control, we have observed near equivalent trends among GISP-2, GRIP and our data (Fig. 4). Terminated at ~ 2300 y in our record, the oldest wetter phase of our study is reasonably coeval with the lighter  $\delta^{18}\text{O}$  during 3000- 2400 y in GRIP, which is persistent also in GISP-2 data. Perturbation of ~2200-1800 (Event III) is also present in both these records. The wetter climate of 1800-1200 y and the reduced fluvial flux during event II are equitable with the trend of lighter  $\delta^{18}\text{O}$  incursion and an abrupt shift toward heavier  $\delta^{18}\text{O}$  at ~ 1100 y in GRIP and GISP-2. The event I (650-450 cal y) has more high frequency variations in the GISP-2 and GRIP data (Fig. 4).

As described earlier, a teleconnection between SWM and the climate changes in the North Atlantic has been proposed through an atmospheric connection (Zonneveld *et al.*, 1997). It is suggested that any change in the boundary conditions at high latitude influences amount and timing of southward transport of cold winds (Colin *et al.*, 1998) which in turn influences the low pressure area over the Tibetan Plateau through the processes described in the westerly swing model of Fang *et al.* (1999). Our data is time averaged for multi decadal periods compared to annual GISP- 2 and GRIP data. Within

these limitations, from the broader concordant trends, we endorse a teleconnection between century scale instabilities at the high latitude as well in SWM during the Late Holocene through an atmospheric connection (Zonneveld *et al.*, 1997).

### **Solar irradiance (Sun spot) vis-à-vis SWM précipitation**

Small variation (0.25%) in solar irradiance has been linked with large global climate and the SWM instabilities (Rind and Overpeck, 1993; Overpeck *et al.*, 1996, Bond *et al.*, 2001; Neff *et al.*, 2001; Agnihotri *et al.*, 2002; Agnihotri and Dutta, 2003). Total Solar Irradiance (TSI) data are available only for 1961-850 AD, which preclude its correlation with our data. Because reconstructed sunspot data are available for the entire Holocene, we have obtained a correlation between reconstructed TSI and sun spot number (Rz); (Bard *et al.*, 2000; Solanki *et al.*, 2004) which is rather good ( $r=0.79$ ;  $p=0.01$ ). We have used available sun spot data (spline for 10 y and 5 point running averaged) for a comparison.

A reducing trend in Rz has been observed during wetter phases of 3000- 2200, 1800-1200 and 900-650 cal y BP (Fig. 5). The last wetter event of <400 y persisted during known episodes of reduced sunspot activity during Sporer Minimum (1410-1515 AD;  $Rz\ 0.06\pm\ 0.59$ ; Riogozo *et al.*, 2001). The Maunder and Dalton Minimums (1641-1715;  $Rz\ 3.56\pm\ 8.72$ ; 1790-1825 AD;  $Rz\ 26.10\pm\ 23.40$  respectively; Riogozo *et al.*, 2001; Bard *et al.*, 2007) overlapped with a wetter climate of < 400y, which were persistent also in the Western Arabian Sea (Fig. 6; Anderson *et al.*, 2002; Gupta *et al.*, 2003). Significantly, the drier events, specifically the event I and III, were found to prevail when the sunspot number or TSI have a reducing trend (Fig. 5). The marine core data of Tiwari *et al.* (2005, 2006), from a northwards location in the Arabian Sea (Fig. 1) have been evaluated also to verify if there is a consistency in these trends from the other region with similar climatology and SSS variability. Concordant with our studies, in this study also, heavier  $\delta^{18}O$  incursion at 800 AD predates the Oort Minimum (1090-1140 AD) while other incursion at 1050-1100 AD (~950 BP) is post dated to it (Fig. 6). A wetter climate (lighter  $\delta^{18}O$  incursion) is found to prevail post 800 AD during Oort Minimum, and subsequent to 1400 AD (0-550 BP) there are lighter as well as heavier incursions during Sporer Minimum (Fig. 3 of Tiwari *et al.*, 2005). Our data, therefore, do not endorse direct or robust relation between solar flux enhancements with that of regional wetter climate along the SE Arabian Sea, though some drier phases do occur during the episodes of reduced sun spot activities (Fig. 6). Because our climate reconstruction is concordant with other regions of the world, specifically from the Chesapeake Bay and from the Sargasso Sea (Fig. 6), we infer discordant, diverse climatic response to the solar

irradiance in the different regions of the globe, more specifically on the SWM precipitation, which perhaps underline the importance of factors which regulate precipitation mechanism.

With respect to SWM, solar irradiance is expected to regulate: (i) evaporation, (ii) extent of low pressure at the Tibetan Plateau - enhancing land-ocean heat gradient, (iii) position of Inter Tropical Convergence Zone (ITCZ), and (iv) glacial-interglacial boundary conditions vis-à-vis the Himalayan albedo (Pant and Kumar, 1997; Beer *et al.*, 2000; Sinha *et al.*, 2007; Juyal *et al.*, 2009). It has been postulated that a pronounced low due to heating of the Tibet during high solar insolation deflects westerlies northwards as explained in the swing model of Fang *et al.* (1999), and aids in a northward migration of ITCZ, together with an enhanced moisture advection into arid and semi arid regions of the Himalayas (Pant and Kumar, 1997; Fang *et al.*, 1997; Bookhagen *et al.*, 2005; Juyal *et al.*, 2009). The available data are, however, not adequate to decipher which of these specific processes influence the SWM precipitation in the different climatic zones of India. We could not obtain also a clue on the extent of modification of the low pressure region over the Tibet by marginal century scale variability of solar activity and its impact on regional SWM precipitation. We tentatively suggest that there is a different response to solar forcing in the different climatic zones of the India. We propose that the SW continental margin of the India has a different response to solar forcing owing to ample availability of moisture, and the precipitation mechanism, which enhances by orographic influences. We speculate that a small increase in solar activity even-though may influence the (i) global evaporation and (ii) northward moisture advection in the terminal stages of SWM, at least, does not amend significantly the magnitude of precipitation at the SW continental margin of India. We propose a significant role of the condensation mechanism. The orographic induced variability in the magnitude of condensation mechanism of moisture from the SWM trade winds by about 2000-3000m topographic high the Western Ghat appears to play a vital role on the entrapment of moisture and magnitude of precipitation from the SWM winds along the SW continental margin of India.

## **Conclusions**

The late Holocene climate is marked with regional century scale rapid instabilities in the SWM precipitation at 2200-1800, ~ 1000 and 650-450 y cal BP, and we affirm a teleconnection between the North Atlantic climate and SWM through atmospheric connection.

In an evaluation of the influence of solar forcing on the SWM precipitation, we have found variable climate during the known period of reduced sun spot activities (Fig. 6). We propose, therefore,

that the SWM precipitation is a complex phenomenon, and our data does not endorse a direct and robust relation between solar influence (sun spot number) and orographically enhanced regional SWM precipitation along the SW continental margin of India.

**Acknowledgement:** - Authors thank Director NIO for providing facilities. They also acknowledge efficient, constructive review processes by Professor Chris Turney. Authors are grateful also to professor Sander van der Kaars, and an other anonymous reviewer for their very constructive review which has contributed immensely to improve scientific contents of ms. The authors are thankful to M. Saranthein, for the help with carbon and oxygen isotopic analysis and interpretation. Kumud K Singh, P. Verma and Dinesh are thanked for EPMA analysis. A. Mascarenhas and Arun Mahale are thanked for editorial help and drafting the illustrations. We also acknowledge ISRO-IGBP and INSA-DFG grants (for the visit of osc to the University of Kiel). Contribution no 5411

## References

Agnihotri R, Dutta K, Bhushan R, Somayajulu BLK. 2002. Evidence for solar forcing on the Indian monsoon during the last millennium. *Earth and Planetary Science Letters* **198**: 521-527. DOI:10.1016/S0012-821X(02)00530-7.

Agnihotri R, Dutta K. 2003. Centennial scale variations in monsoonal rainfall (Indian, east equatorial and Chinese monsoons): Manifestations of solar variability. *Current Science* **85**: 459–463.

Alley RB. 2004. *GISP2 Ice core temperature and accumulation Data*. IGBP Pages/World Data center for Paleoclimatology Data Contribution Series #2004-2013. NOAA/NGDC Paleoclimatology Program, Boulder CO, USA.

Anderson DM, Overpeck JT, Gupta AK. 2002. Increase in the Asian Southwest Monsoon during the past four centuries. *Science* **297**: 596-599. DOI:10.1038/nature01340.

Aswathanarayana U.1964. Origin of the heavy mineral beach sand deposits of the southwest coast of India. In *Advancing frontiers in geology and geophysics*, Subramanian R, Balkrishna S (eds). Indian Geophysical Union: Hyderabad; 481-489.

Bard E, Raisbeck GM, Yiou F, Jouzel J. 2000. Solar irradiance during the last 1200 yr based on cosmogenic nuclides. *Tellus B* **52**: 985-992. DOI:10.1034/j.1600-0889.2000.d01-7.x

Bard E, Raisbeck GM, Yiou F, Jouzel J. 2007. Comment on “Solar activity during the last 1000 yr inferred from radionuclide records” by Muscheler et al. (2007). *Quaternary Science Reviews* **26**: 2301-2304. DOI:10.1016/j.quascirev.2007.06.002.

Beer J, Mende W, Stellmacher R. 2000. The role of the sun in climate forcing. *Quaternary Science Reviews* **19**: 403-415. DOI:10.1016/S0277-3791(99)00072-4.

Bond G, Kromer B, Beer J, Muscheler R, Evans MN, Showers W, Hoffmann S, Lotti-Bond R, Hajdas I, Bonani G. 2001. Persistent solar influence on North Atlantic climate during the Holocene. *Science* **294**: 2130-2136. DOI:10.1126/science.1065680.

Bookhagen B, Thiede RC, Strecker MR. 2005. Abnormal monsoon years and their control on erosion and sediment flux in the high, arid northwest Himalaya. *Earth and Planetary Science Letters* **231**: 131-146. DOI:10.1016/j.epsl.2004.11.014.

Bryson RA. 1996. Proxy indications of Holocene winter rains in southwest Asia compared with simulated rainfall. In *Third Millennium BC; Climate Change and Old World Collapse*, Dalfes HN, Kukla G, Weiss H (eds). NATO ASI Series **49**: Springer-Verlag; 465-473.

Chauhan OS, Gujar AR. 1996. Surficial clay mineral distribution on the southwestern continental margin of India: evidence of input from the Bay of Bengal. *Continental Shelf Research* **16**: 321-333. DOI:10.1016/0278-4343(95)00015-S.

Chauhan OS, Jayakumar S, Menezes AA, Rajawat AS, Nayak SR. 2006. Anomalous inland influx of the River Indus, Gulf of Kachchh, India. *Marine Geology* **229**: 91-100.

Colin C, Kissel C, Blamart D, Turpin L. 1998. Magnetic properties of sediments in the Bay of Bengal and the Andaman Sea: impact of rapid North Atlantic Ocean climatic events on the strength of the Indian monsoon. *Earth and Planetary Science Letters* **160**: 623-635. DOI:10.1016/S0012-821X(98)00116-2.

Cooley JW, Tukey JW. 1965. An Algorithm for the machine computation of the Complex Fourier Series. *Mathematics of Computation* **19**: 297-301.

Cronin TM, Dwyer GS, Thunell RC, Dowsett HJ, Tappa E, Vann CD. 2001. Medieval Warm Period, Little Ice Age, and 20th century climate variability from Chesapeake Bay. *EOS Transactions AGU Fall Meeting Supplement* **83**: n. 47.

Cronin TM, Dwyer GS, Kamiya T, Schwede S, Willard DA. 2003. Medieval Warm Period, Little Ice Age and 20th Century temperature Variability from Chesapeake Bay. *Global and Planetary Change* **36**:17-29.

Crowley TJ, Lowery TS. 2000. How warm was the medieval warm period? *Ambio* **29**: 51–54.

Diamond J. 2005. *Collapse: How Societies Choose to Fail or Succeed*. Penguin Books: New York, ISBN 0143036556; pp. 575.

Duhamel P, Vetterli M. 1990. Fast Fourier Transforms: A tutorial review and a state of the art. *Signal Processing* **19**: 259-299.

Duplessy JC, Be AWH, Blanc PL. 1981. Oxygen and carbon isotope composition and biogeographic distribution of planktonic foraminifera in the Indian Ocean. *Palaeogeography, Palaeoclimatology, Palaeoecology* **33**(1-3): 9-46.

Duplessy JC. 1982. Glacial to interglacial contrasts in the northern Indian Ocean. *Nature* **295**: 494-498. DOI:10.1038/295494a0.

Dupont LM, Mook WG. 1987. Palaeoclimate analysis of  $^2\text{H}/^1\text{H}$  ratios in Peat sequences with variable plant composition. *Chemical Geology* **66**: 323-333.

Eddy JA. 1976. The Maunder Minimum. *Science* **192**: 1189–1202.

Fairbanks RG. 1989. A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-Ocean circulation. *Nature* **342**: 637-642. DOI:10.1038/342637a0.

Fang XM, Li JJ, Van der Voo R, Niocail CM, Dai XR, Kemp RA, Derbyshire E, Cao JX, Wang JM, Wang G. 1997. A record of the Blake event during the last interglacial paleosol in the western Loess Plateau of China. *Earth and Planetary Science Letters* **146**: 73–82. DOI:10.1016/S0012-821X(96)00222-1.

Fang XM, Ono Y, Fukusawa H, Pan BT, Li JJ, Guan DH, Oi K, Tsukamoto S, Torii M, Mishima T. 1999. Asian summer monsoon instability during the past 60,000 years: magnetic susceptibility and pedogenic evidence from the western Chinese Loess Plateau. *Earth and Planetary Science Letters* **168**: 219–232. DOI: 10.1016/S0012-821X(99)00053-9.

Folk RL. 1966. A review of grain-size parameters. *Sedimentology* **6**: 73-93. DOI:10.1111/j.1365-3091.1966.tb01572.x.

Grootes PM, Stuiver M. 1997. Oxygen 18/16 variability in Greenland snow and ice with  $10^{-3}$ - to  $10^5$ -year time resolution. *Journal of Geophysical Research* **102**: 26455-26470. DOI:10.1029/97JC00880. [http://depts.washington.edu/qil/datasets/gisp2\\_main.html](http://depts.washington.edu/qil/datasets/gisp2_main.html).

Gupta AK, Anderson DM, Overpeck JT. 2003. Abrupt changes in the Asian southwest monsoon during the Holocene and their links to the North Atlantic Ocean. *Nature* **421**: 354–357. DOI:10.1038/nature01340.

Guptha MVS, Curry WB, Ittekkot V, Muralinath AS. 1997. Seasonal variation in the flux of Planktic foraminifera: Sediment trap results from the Bay of Bengal, Northern Indian Ocean. *Journal of Foraminiferal Research* **27**: 5–19.

Haigh JD. 2001. Climate variability and the influence of the sun. *Science* **294**: 2109-2111. DOI:10.1126/science.1067013.

Hasternath S, Greichar L. 1993. Changing predictability of Indian monsoon rainfall anomalies? *Journal of Earth System Science* **102** (1): 35-47. DOI: 10.1007/BF02839181.

Hilbrecht H, Frieg C, Tröger KA, Voigt S, Voigt T. 1996. Shallow water facies during the Cenomanian–Turonian anoxic event: bio-events, isotopes, and sea level in southern Germany. *Cretaceous Research* **17**: 229–253. DOI:10.1006/cres.1996.0017.

Hu SF, Ito E, Brown TA, Curry BB, Engstrom DR. 2001. Pronounced climatic variations in Alaska during the last two millennia. *Proceedings of the National Academy of Sciences* **98(19)**: 10552-10556. DOI:10.1073/pnas.181333798.

IPCC. 2001. Climate Change 2001: In *The Scientific Basis*. Contribution of Working Group I in The Third Assessment Report of the Intergovernmental Panel on Climate Change, Houghton JT, Ding Y, Griggs DJ, Noguer M, van der Linden PJ, Dai X, Maskell K, Johnson CA (eds). Cambridge University Press: Cambridge.

Juyal N, Pant RK, Basavaiah N, Bhusan R, Jain M, Saini NK, Yadava MG, Singhvi AK. 2009 Reconstruction of Last Glacial to early Holocene monsoon variability from relict lake sediments of the Higher central Himalaya, Uttarakhand, India. *Journal of Asian Earth Sciences* **34**: 437-449. DOI:10.1016/j.jseaes.2008.07.007.

Keigwin LD. 1996. The Little Ice Age and Medieval Warm Period in the Sargasso Sea. *Science* **274**: 1503-1508. DOI:10.1126/science.274.5292.1503.

Levitus S, Burgett R, Boyer T. 1994. World Ocean Atlas: *Salinity* NOAA Atlas. *NESDIS* **3**: pp. 99.

Levitus S, Boyer T. 1994. World Ocean Atlas: *Temperature* NOAA Atlas. *NESDIS* **4**: pp.117.

- Mallik TK, Vasudevan V, Verghese PA, Machado T. 1987. The black sand placer deposits of Kerala, southwest India. *Marine Geology* **77**: 129-150. DOI:10.1016/0025-3227(87)90088-0.
- Murray RW, Leinen M, Isern AR. 1993. Biogenic flux of Al to sediment in the Central Equatorial Pacific Ocean: evidence for increased productivity during glacial episodes. *Paleoceanography* **8**: 651–670.
- Murray RW, Leinen M. 1996. Scavenged excess aluminum and its relationship to bulk titanium in biogenic sediment from the Central equatorial Pacific Ocean. *Geochimica et Cosmochimica Acta* **60**: 3869–3878. DOI:10.1016/0016-7037(96)00236-0.
- Nagam Aiya V. 1989. *The Travancore state manual*. Asian Educational services: New Delhi: 1-20.
- Neff U, Burns SJ, Mangini A, Mudelsee M, Fleitmann D, Matter A. 2001. Strong coherence between solar variability and the monsoon in Oman between 9 and 6 kyr ago. *Nature* **411**: 290–293. DOI:10.1038/35077048.
- Nesbitt HW, Young GM. 1982. Early Proterozoic climates and plate motions inferred from major element chemistry of lutites. *Nature* **299**: 715-717. DOI:10.1038/299715a0.
- Overpeck J, Anderson D, Trumbore S, Prell W. 1996. The southwest Indian Monsoon over the last 18000 years. *Climate Dynamics* **12**: 213–225. DOI:10.1007/BF00211619.
- Pant GB, Kumar RK. 1997. *Climates of South Asia*. John Wiley and Sons: Chichester, UK.
- Pattan JN, Shane P. 1999. Excess aluminum in deep sea sediments of the Central Indian Basin. *Marine Geology* **161**: 247-255. DOI:10.1016/S0025-3227(99)00042-0.
- Prell WL, Kutzbach JE. 1992. Sensitivity of the Indian monsoon to forcing parameters and implications for its evolution. *Nature* **360**: 647-652. DOI:10.1038/360647a0.
- Qasim SZ. 1977. Biological productivity of the Indian. Ocean. *Indian Journal marine Sciences* **6**: 122-137.
- Rigozo NR, Echer E, Vieira LEA, Nordemann DJR. 2001. Reconstruction of Wolf Sunspot Numbers on the Basis of Spectral Characteristics and Estimates of Associated Radio Flux and Solar Wind Parameters for the Last Millennium. *Solar Physics* **203**: 179-191. DOI: 10.1023/A:1012745612022.
- Rind D, Overpeck J. 1993. Hypothesized causes of decade-to century-scale climate variability: climate model results. *Quaternary Science Reviews* **12**: 357–374. DOI: 10.1016/S0277-3791(05)80002-2.



- Rind D. 2002. The Sun's role in Climate Variations. *Science* **296**: 673-677. DOI: 10.1126/science.1069562.
- Rolinski HD, Rogalla U, Scheeder G, Luckge A, von Rad U. 2001. High – resolution Temperature and Evaporation changes during the late Holocene in the Northeastern Arabian Sea. *Paleoceanography* **16**: 358-367.
- Rostek F, Ruhland G, Bassinot F, Muller PJ, Labeyrie, LD, Lancelot Y, Bard E. 1993. Reconstruction of sea surface temperature and salinity using  $\delta^{18}\text{O}$  alkenone records. *Nature* **364**: 319-321. DOI:10.1038/364319a0.
- Saraswat R, Nigam R, Weldeab S, Mackensen A. 2007. A tropical warm pool in the Indian Ocean and its influence on ENSO over the past 137,000 yrs BP. *Current Science* **92(8)**: 1153-1156.
- Sarkar A, Ramesh R, Somayajulu BLK, Agnihotri R, Jull AJT, Burr GS. 2000. High resolution Holocene monsoon record from the eastern Arabian Sea. *Earth and Planetary Science Letters* **177**: 209-218. DOI:10.1016/S0012-821X(00)00053-4.
- Schulz H, von Rad U, Erlenkeuser H. 1998. Correlation between Arabian Sea and Greenland climate oscillations of the past 110,000 years. *Nature* **393**: 54–57. DOI:10.1038/31750.
- Sinha A, Cannariato KG, Stott LD, Cheng H, Edwards RL, Yadava MG, Ramesh R, Singh IB. 2007. A 900-year (600 to 1500 AD.) record of the Indian summer monsoon precipitation from the core monsoon zone of India. *Geophysical Research Letters* **34**: L16707. DOI:10.1029/2007GL030431.
- Sirocko F, GarbeSchonberg D, McIntyre A, Molfino B. 1996. Teleconnections between the Subtropical Monsoons and High-Latitude Climates during the Last Deglaciation. *Science* **272**: 526–529. DOI:10.1126/science.272.5261.526.
- Solanki SK, Usoskin IG, Kromer B, Schussler M, Beer J. 2004. Unusual activity of the Sun during recent decades compared to the previous 11,000 years. *Nature* **431**: 1084-1087. DOI:10.1038/nature02995. [www.ncdc.noaa.gov/paleo/forcing.html](http://www.ncdc.noaa.gov/paleo/forcing.html).
- Staubwasser M, Sirocko F, Grootes PM, Segl M. 2003. Climate change at the 4.2 ka BP termination of the Indus valley civilization and Holocene south Asian monsoon variability. *Geophysical Research Letters* **30(8)**: 1425; DOI:10.1029/2002GL016822.
- Stuiver M, BraziunasTF. 1993. Modeling atmospheric  $^{14}\text{C}$  influences and  $^{14}\text{C}$  ages of marine samples to 10 000 BC. *Radiocarbon* **35**: 137-191.

- Swain AM, Kutzbach JE, Hastenrath S. 1983. Estimates of Holocene precipitation for Rajasthan, India, based on pollen and lake – level data. *Quaternary Research* **19**: 1-17. DOI:10.1016/0033-5894(83)90024-8.
- Tiwari M, Ramesh R, Somayajulu BLK, Jull AJT, Burr GS. 2005. Solar control of southwest monsoon on centennial timescales. *Current Science* **89**: 1583-1588.
- Tiwari M, Ramesh R, Yadava MG, Somayajulu BLK, Jull AJT, Burr GS. 2006. Is there a persistent control of monsoon winds by precipitation during the late Holocene?. *Geochemistry Geophysics Geosystems* **7**: Q03001. DOI: 10.1029/2005GC001095.
- Unser M, Aldroubi A, Eden M. 1993. B-Spline signal processing: Part I-Theory. *IEEE Transactions on Signal Processing* **41**: 821-833.
- Vinther BM, Clausen HB, Johnsen SJ, Rasmussen SO, Andersen KK, Buchardt SL, DahlJensen D, Seierstad IK, Siggaard AML, Steffensen JP, Svensson A, Olsen J, Heinemeier JA 2006. Synchronized dating of three Greenland ice cores throughout the Holocene. *Journal of Geophysical Research* **111**: D13102. DOI:10.1029/2005JD006921. <http://www.iceandclimate.nbi.ku.dk/data>.
- von Rad U, Schaaf M, Michels KH, Schulz H, Berger WH, Sirocko F. 1999. A 5000–yr Record of Climate Change in Varved Sediments from the Oxygen Minimum Zone off Pakistan, Northeastern Arabian Sea. *Quaternary Research* **51**: 39–53. DOI:10.1006/qres.1998.2016.
- Wanner H, Beer J, Butikofer J, Crowley TJ, Cubasch U, Fluckiger J, Goosse H, Grosjean M, Joos F, Kaplan OJ, Kuttel M, Muller AS, Prentice C, Solomina O, Stocker TF, Tarasov P, Wagner M, Widmann M. 2008. Mid- to Late Holocene climate change: an overview. *Quaternary Science Reviews* **27**: 1791-1828. DOI:10.1016/j.quascirev.2008.06.013.
- Warrier AK, Shankar R. 2009. Geochemical evidence for the use of magnetic susceptibility as a paleorainfall proxy in the tropics. *Chemical Geology* **265**: 553-562. DOI:10.1016/j.chemgeo.2009.05.023.
- Yadava MG, Ramesh R, Pant GB. 2004. Past monsoon rainfall variations in peninsular India recorded in a 331-year-old speleothem. *Holocene* **14**: 517–524. DOI: 10.1191/0959683604hl728rp.
- Zonneveld KAF, Ganssen G, Troelstra S, Versteegh GJM, Visscher H. 1997. Mechanisms forcing abrupt fluctuations of the Indian Ocean summer monsoon during the last deglaciation. *Quaternary Science Reviews* **16**: 187–201. DOI:10.1016/S0277-3791(96)00049-2.

## Figure captions

Fig. 1. Location of core, regional sea surface salinity (p.s.u) during August, the Southwest Indian summer monsoon mean precipitation rate (mm/day; modified from NCEP/NCAR dataset of June to September), and major fluvial sources in the study area (inset). The geographical locations of other available studies from the continental India, and the Arabian Sea used for reconstruction and comparisons of regional trend in SWM are shown.

Fig. 2. The age model of the core (AMS dates are shown with open circles). Ages shown are corrected cal ages (refer methodology, Table 2).  $\pm$  indicates 1 s.d. (y).

Fig. 3. Temporal variations in *G. ruber*, *G. sacculifer*, *G. menardii*, CIA, Al and Ti. Heavier  $\delta^{18}\text{O}$  and reduced Al and CIA during 450-650 (Arid I), ~1000 (Arid II) and 1800-2200 y BP (Arid III) interpreted as the events of reduced precipitation and fluvial flux are identified by gray shades. For better chronometric precision, ages are obtained on specific instabilities (bold arrows represent AMS ages; thin arrows represent bulk  $^{14}\text{C}$  ages in panel (a).

Fig. 4. Comparison of: (a) GISP-2, (b) GRIP (Filtered and unfiltered data), (c-d)  $\delta^{18}\text{O}$  of *G. ruber* and *G. sacculifer* and (e) CIA. Correspondence among climate episodes I - III is highlighted by gray bars.

Fig. 5. Comparison of available (a) Total Solar Irradiance (b) spline and 5 point running averaged Solar Sun Spot data (Rz Solanki *et al.*, 2004) with (c)  $\delta^{18}\text{O}$  of *G. ruber* and (d) CIA. All known events of sun spot minima are shown in Panel (b). Drier climate during 650 and 450 y cal BP predates Sporer Minimum (1410-1515 AD) while lighter  $\delta^{18}\text{O}$  (enhanced fluvial flux) is found during Maunder (1640-1715 AD) and Dalton Minima (1790-1825) during LIA.

Fig. 6. Schematic diagram of climate variability during the Late Holocene from available continental and marine studies in the Arabian Sea, and the North Atlantic together with their correspondence with global climate. The levels on which  $^{14}\text{C}$  ages are obtained and time span covered by each of these studies are shown. The episodes of reduced sunspot activities are shown also for comparison with global climate variability.

**Table 1.**

Characteristic of the rivers in the study area (data source Aswathanarayana, 1964; Mallik *et al.*,1987; Nagam Aiya, 1989)

River	Elevation at point of origin (m)	Length (km)	Gradient	Rainfall	Peak discharge	Discharge ( $10^6 \text{ m}^3 \text{ y}^{-1}$ )
Kallada	1878	121	1:65	480 cm	J – S*	1.14
Karmana	1697	68	1:40	510 cm	J – S	1.14
Neyyar	1848	56	1:30	490 cm	J – S	1.13
Tambraparni	1697	61	1:35	480 cm	J - S	1.12

- J= June, S= September, Rainfall is averaged from 20 yr data

**Table 2**

Un-calibrated and calibrated calendar ages (with 1 standard deviation) used in the construction of age model for the core.

level (cm)	Lab ID	Species	<sup>14</sup> C Age (y BP)	Calibrated Age (y BP)	1 s.d. (y)
10	RG1673904/1	Bulk**	840	460	90
13	POZ-22003	<i>G.ruber</i> *	950	590	50
14	POZ-22021	<i>G.ruber</i> *	1050	630	40
15	RG1673904/2	Bulk**	1160	710	30
20	RG1673904/3	Bulk**	1440	960	110
24	RG1673904/4	Bulk**	1600	1140	100
28	RG1673904/5	Bulk**	1760	1300	30
30	RG1673904/6	Bulk**	1870	1370	40
38	RG1673904/7	Bulk**	2190	1730	110
42	POZ-22062	<i>G.ruber</i> *	2450	2020	50
46	POZ-22064	<i>G.ruber</i> *	2730	2350	30
50	RG1673904/8	Bulk**	2940	2600	130

\*\* unbroken planktonic foraminifers;

\* AMS age

Figure 1

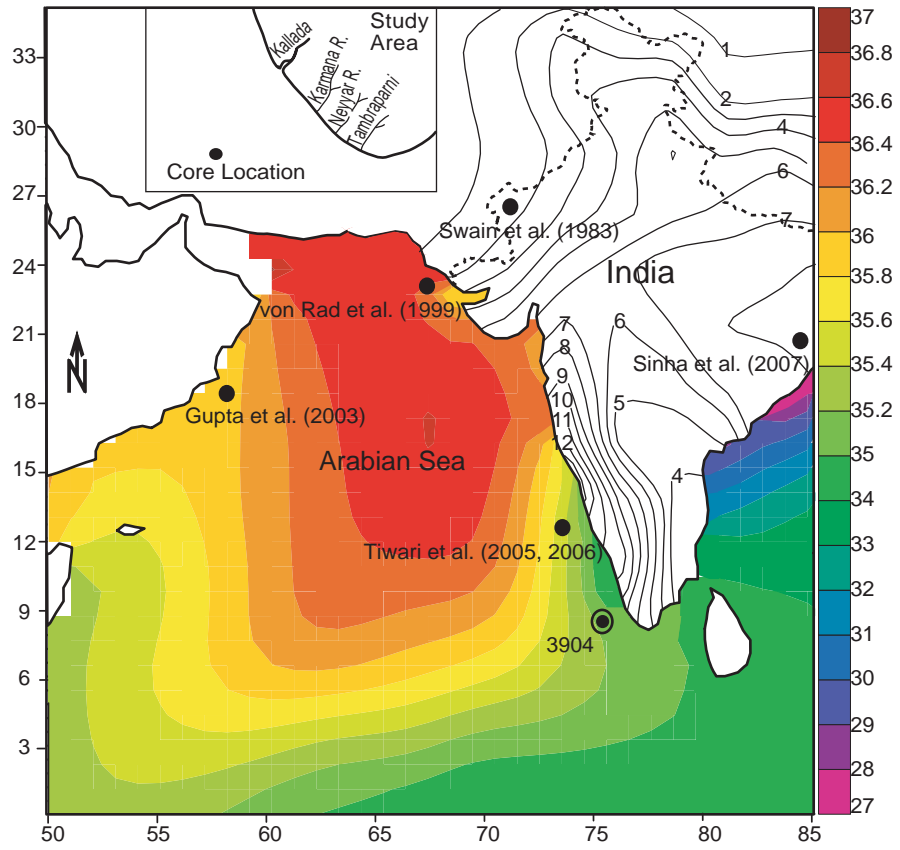


Figure 2

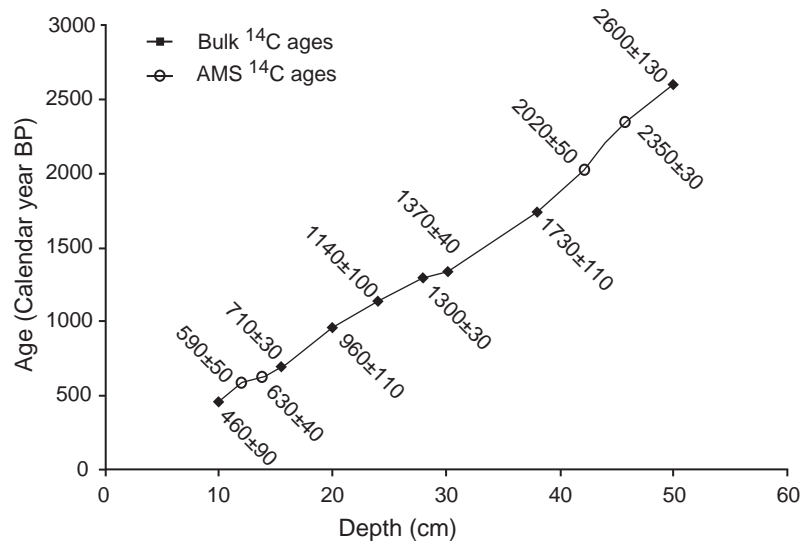


Figure 3

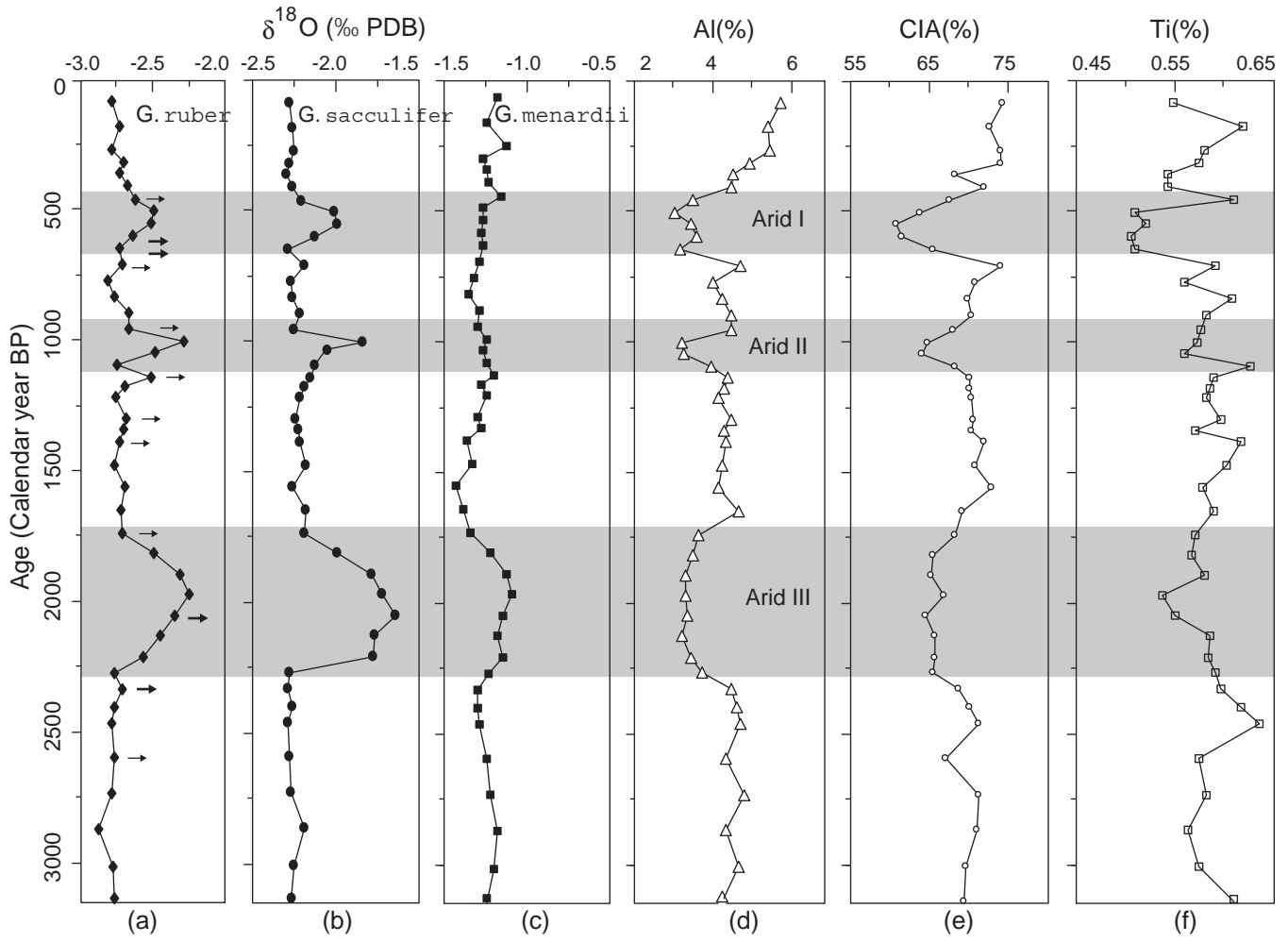




Figure 4

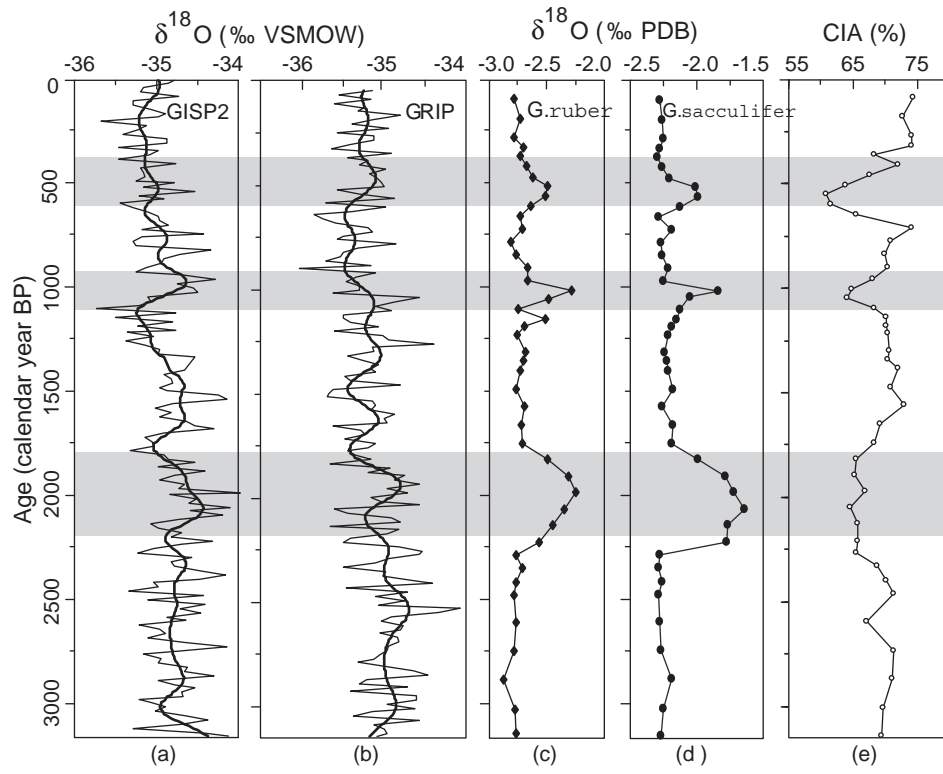


Figure 5

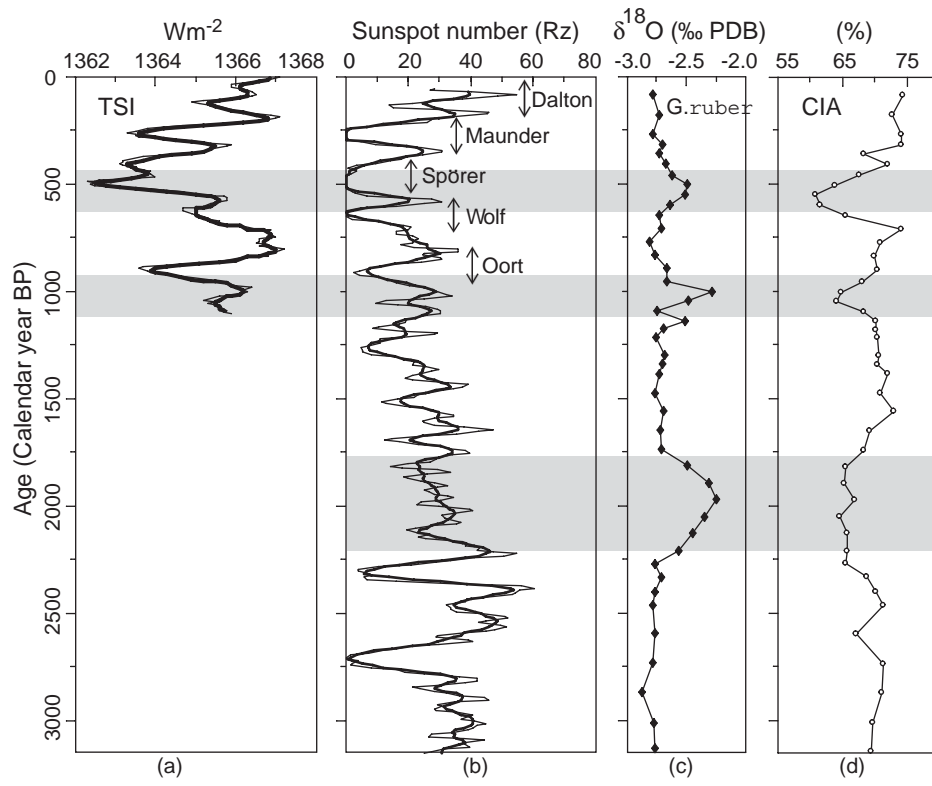


Figure 6

