Variation in the Indian Summer Monsoon intensity during the Bølling-Ållerød and Holocene

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Abstract

Variations in the Indian summer monsoon (ISM) intensity during the last 16.7 ka have been studied using organic carbon (C_{org}), $\delta^{15}N$ of sedimentary organic matter, $CaCO_3$, sediment texture, $\delta^{18}O_C$ and Mg/Ca derived sea surface temperature, $\delta^{18}O$ of sea water and sea surface salinity, in a ^{14}C -dated sediment core from the eastern Arabian Sea. The $\delta^{18}O$ in water and planktonic foraminifera shells off the central west coast of India may be controlled by the ISM intensity as this area receives high precipitation and land runoff. Also, the C_{org} and $CaCO_3$ contents of sediments and $\delta^{15}N$ of sedimentary organic matter may be linked to ISM-induced productivity and denitrification. The results of the present study reveal that between 16 and 15.2 ka BP the ISM was weak with minor fluctuations and started intensifying around 15.2 ka BP, at the onset of the Bølling-Ållerød (B-A) event. The B-A event is characterized by higher water column denitrification rates comparable to the present day. The ISM signatures observed in the $\delta^{18}O_C$ record of B-A event compare well with those from Timta cave of the western Himalayas and also the Asian summer monsoon signatures from the Hulu caves in China and warming signatures in Greenland Ice Sheet Project 2 (GISP2) suggesting atmospheric tele-connections through Intertropical Convergence Zone (ITCZ). The boundary between the Younger Dryas and the Holocene is discernible with small episodes of abrupt events of increased ISM intensity. This decrease in $\delta^{18}O_C$ values at ~ 11.8 ka BP is contemporary with June solar insolation maximum at 30° north and the increase in methane in the GISP2 ice core supporting episodes of warmer climate and increase in ISM intensity. The ISM seems to have been most stable between 7 and 5.6 ka BP. The core exhibits periodicity of 500 that is comparable to the Atlantic water formation and the Chinese monsoon.

Key words:

Monsoon, Bølling-Ållerød, Holocene, India, Arabian Sea, Sediment

Introduction

Climatologically, the monsoon regions are important areas on the Earth and are responsible for the transport of major part of global atmospheric heat and moisture [Clemens et al., 2003]. One of the most important of such systems is the Indian Summer Monsoon (ISM). Both modern and paleo- studies in the Indian Ocean have demonstrated that the ISM is highly variable over annual to orbital time scales [e.g. Altabet et al., 1999; Clement et al., 1999; Saji et al., 1999; Clark et al., 2000]. The past ISM intensities and environmental changes observed in proxy records from the Arabian Sea sediments have been related to changes in the Atlantic Ocean [e.g. Suthhof et al., 2001; Altabet et al., 2002] and in the Pacific Ocean [Saraswat et al., 2007] suggesting atmospheric and/or oceanic teleconnections. The ISM is a part of the larger phenomenon, the Asian monsoon, associated with the Intertropical Convergence Zone (ITCZ) which connects the rainfall regions of Asia. Variations in the mean latitudinal position of ITCZ influence tropical and extratropical climate [Fleitmann et al., 2007]. Changes in the ISM rainfall have been observed to be coupled with variations in the East Asian monsoon and the North Atlantic climate [Sinha et al., 2005]. The Holocene ISM records from the Arabian Sea have revealed a link with climate variability over continental China and the South China Sea suggesting simultaneous fluctuations of monsoon intensities in these places [Jung et al., 2004]. The ISM variability during Bølling-Ållerød (B-A) event has been linked to that of the East Asian monsoon [Sinha et al., 2005] based on stalagmite oxygen isotope record from the Timta caves. Staubwasser and Weiss [2006] utilized the data from the Mediterranean and South Asia to compile the abrupt events in these regions and suggested that the ISM became part of the Asian Monsoon since the Holocene time.

The Arabian Sea-based paleo- ISM reconstructions are mostly derived from sediment cores from the northern and western Arabian Sea, with relatively few studies in the eastern Arabian Sea. The ISM proxies used for the northern and western Arabian Sea are mostly upwelling and wind related [e.g. Schulz et al., 1998; Sirocko et al., 1993; Gupta et al., 2003], whereas in the eastern Arabian Sea they are precipitation related like oxygen isotopes, Mg/Ca ratios, and terrigenous material [Thamban et al., 2002; Kumar et al., 2005; Anand et al., 2008; Banakar et al., 2010; Rao et al., 2010; Govil and Naidu., 2010; Mahesh et al. 2011]. The high sedimentation rates and varved sediments in the northern Arabian Sea help in deciphering proxy variations of the past at a high temporal resolution. In comparison, because of lower sedimentation rates in the eastern Arabian Sea with consequently coarser temporal resolution fewer proxy records are generated so far [Thamban et al., 2001; Tiwari et al., 2005; Singh et al., 2006; Kessarkar et al., 2007, 2010; Govil and Naidu, 2010; Banakar et al., 2010, Rao et al., 2010] there is a need to study this area to infer past ISM variations on decadal to centennial time scales. There have been a few high resolution studies, though, namely by Agnihotri et al. [2008] off Goa,

covering the past seven centuries, and by Thamban et al. [2007] that extended to the beginning of the Holocene, which indicated significant fluctuations in the strengths of ISM. The paleoclimatic reconstructions from sediment proxies in the northern Arabian Sea suggest increase in ISM to be associated with the solar insolation maxima. By contrast, the study by Tiwari et al. [2010] in the eastern Arabian Sea suggests strengthening of the ISM during the late Holocene with no link to solar insolation maximum. In view of the above results we selected a core from the southeastern Arabian Sea that receives high monsoon precipitation and runoff during the ISM (Figure 1 A-F) to identify the changes in the strengths of ISM and the abrupt events during the past 16.7 ka BP. We present here surface hydrographic change from oxygen isotopes and Mg/Ca derived temperature. Further we have combined the new and previously published data to create a robust composite record of the timing and scale of surface hydroclimate change. Comparison is also made with published records of ITCZ linked East Asian monsoon and GISP2 core to explore the link within the Asian monsoon and the dependence of the low-latitude climate on that at high latitudes.

Study Area:

The ISM is a part of Asian summer monsoon (Figure 1G inset). During June-August the eastern Arabian Sea receives high precipitation and runoff (Figure 1A-F) that depresses salinity of surface waters along the southwest coast of India (southeastern Arabian Sea). These low salinity signatures are reflected in the lower $\delta^{18}\text{O}_w$ of seawater in this area [Dahl and Oppo, 2006]. Nutrient enrichment through upwelling supports high productivity off Arabia, Somalia and southeastern Arabian Sea during this period [Qasim, 1977]. Increased productivity is also observed during the northeast monsoon north of about 15°N latitude. The subsurface water renewal in the Indian Ocean mostly occurs through advection of waters from the south and to a smaller extent by outflows from the Persian Gulf and the Red Sea. As the rate of water renewal is moderate and productivity is high, a pronounced and perennial oxygen minimum zone (OMZ) with dissolved $\text{O}_2 < 0.5 \text{ ml l}^{-1}$ ($\sim 22 \mu\text{M}$) has developed in the Arabian Sea at intermediate water depths - between approximately 150 and 1200 m [Wyrki, 1971; Sen Gupta and Naqvi, 1984]. The Arabian Sea OMZ is one of the largest sites of water column denitrification in the world [Naqvi, 1987]. This process causes an enrichment of ^{15}N in dissolved nitrate [Brandes et al., 1998] and characteristically elevated $\delta^{15}\text{N}$ signatures are preserved in sediments. Sedimentary $\delta^{15}\text{N}$ records have therefore been used to infer changes in denitrification intensity in the past, with high $\delta^{15}\text{N}$ indicating more vigorous denitrification and probably stronger ISM and vice versa [Altabet et al., 1999, 2002; Ganeshram et al., 2000].

Methodology

For the present study we have used a gravity core AAS62/1 collected during the 62nd cruise of A.A. Sidorenko from southeastern Arabian Sea (11°30.456N, 74°37.388E) at 800 m water depth (Figure 1). The core length was 5.08 m and it was sub-sampled at 2 cm intervals in the upper 100 cm and at 5 cm intervals in the rest of the core. As the core was being cut the sediment colour was checked with the rock colour chart and was found to be grayish olive green (5 GY 3/2) uniformly throughout the core.

For measurement of the oxygen isotopic composition of carbonate ($\delta^{18}\text{O}_\text{C}$), shells of planktonic foraminifer *Globigerinoides ruber* were selected. *G. ruber* lives throughout the year in the Arabian Sea, and calcifies above 80 m depth [Dahl and Oppo, 2006]. These species were handpicked from the coarse fraction (250–350 μm size) using binocular microscope and cleaned ultrasonically. About ten *G. ruber* tests were selected from each of the 125 sediment intervals and analyzed for stable oxygen isotopes on GV Isoprime Mass Spectrometer. All the values are given in ‰ relative to PDB. Repetitive analysis of the NBS 18, 19 and internal laboratory standards revealed that analytical standard deviation was better than $\pm 0.07\text{‰}$.

Total inorganic carbon (TIC) was determined using a coulometer. CaCO_3 (%) was calculated by multiplying TIC with factor of 8.333, the reproducibility was found to be better than 5 %. In order to estimate total organic carbon (C_org), total carbon was measured by CNS elemental analyser (NCS 2500) and C_org was derived from the difference between total carbon and TIC, with a reproducibility of $\pm 1\%$.

For the analysis of the land derived (terrigenous) sediment grain size, 3 g of sediment sample was taken in a beaker. Sea-derived materials like organic matter and CaCO_3 were removed by treating with H_2O_2 and 1N HCL till no more effervescence was seen. The residue was suspended in distilled water and centrifuged; the process was repeated till all the acid was removed. The sample was then suspended in 30 ml of distilled water. 10 to 12 ml of the well-stirred slurry (depending on the obscuration range of the instrument) was analyzed using Malvern Laser particle size analyser (Master-Sizer 2000). The mud fraction of the sediment was divided into different size grades following Wentworth classification [Folk, 1968].

For the analyses of carbon and nitrogen isotopic composition of the sedimentary organic matter, 2 g of powdered sample was decalcified (treated with 1N HCl for the removal of CaCO_3). The residue was suspended in distilled water and centrifuged. The process was repeated till all the acid was

removed and the decalcified sediment was dried in an oven at 60°C. The decalcified powdered samples (~20 mg) were weighed in clean tin cups for analysis. A Delta V plus Stable Isotope Ratio Mass Spectrometer (Thermo®) coupled with a Euro-Vector Elemental Analyzer was used for isotopic measurements in a continuous flow mode. The $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ values are expressed as per mil (‰) deviations of the isotopic ratios from the Pee Dee Belemnite (PDB) and air standards, respectively. The analytical standard deviation based on replicate analyses of laboratory standards was about $\pm 0.2\text{‰}$ for $\delta^{15}\text{N}$ and $\pm 0.3\text{‰}$ for $\delta^{13}\text{C}$.

For Mg/Ca analysis 20 shells of *G. ruber* were selected under a binocular microscope from the 360-440 μm size fraction and gently crushed. Above core depth of 175-180 cm, the sensu stricto morphotype was strictly chosen, below this depth few sensu lato morphotype specimens needed to be included in order to have sufficient material for analysis. Cleaning was performed according to the protocol of Barker et al. [2003]. Mg/Ca ratios were measured on an Agilent Technologies 700 Series inductively coupled plasma-optical emission spectrometer (ICP-OES) equipped with an autosampler at the Center for Marine Environmental Sciences, Bremen. Instrumental precision during the analytical session was monitored by analysis of an in-house standard solution with an average Mg/Ca of 2.96 mmol/mol ($n=20$; $\sigma = 0.01$ mmol/mol) run after every five samples. For inter-laboratory comparison we measured the Mg/Ca of carbonate reference material ECRM 752-1 before and after the samples. Its Mg/Ca ratio was on average 3.83 mmol/mol ($n=4$; $\sigma = 0.01$ mmol/mol), in good agreement with the reported Mg/Ca ratio of 3.75 mmol/mol [Greaves et al., 2008]. Analytical precision based on three replicate measurements of each sample was on average 0.11% for Mg/Ca.

Mg/Ca temperatures were calculated using the calibration of Anand et al. [2003] for *G. ruber* (white): $\text{Mg/Ca} = 0.48 \exp(0.085 \cdot T)$. The $\delta^{18}\text{O}_\text{C}$ measured on *G. ruber* was corrected for continental ice volume using data set of Waelbroeck et al. [2002] to obtain the residual $\delta^{18}\text{O}_\text{C}$. Further $\delta^{18}\text{O}_\text{W}$ was obtained using equation of Bemis et al. [1998]: $\delta^{18}\text{O}_\text{W} = 0.27 + ((T - 16.5 + 4.8 \cdot \delta^{18}\text{C}) / 4.8)$ where T is Mg/Ca derived SST and $\delta^{18}\text{C}$ is residual $\delta^{18}\text{O}_\text{C}$. To obtain salinity from $\delta^{18}\text{O}_\text{W}$ we used the equation given by Dahl and Oppo [2006]: $S = (\delta^{18}\text{O}_\text{W} + 20) / 0.57$.

Chronological control on the core was achieved through 10 AMS ^{14}C dates on *G. ruber* at Woods Hole Oceanographic Institution (Table 1). The ages were calibrated using Calib 5.0.2 using ΔR 138 ± 64 years [Stuiver et al., 2005] and an age model was constructed through linear interpolation between the measured ages. The average age between two points was found to be 0.13 ka. The topmost age obtained at 6-8 cm interval was 1.18 ka BP; this was extrapolated to the surface of the core. Based on this extrapolation the core surface (0-2 cm) dated 0.64 ka BP. The computed

sedimentation rates between AMS dated tie points were found to vary between 6.3 and 98.7 cm/ka (Figure 2A). Higher sedimentation rates are associated with low sea level and decreased as the sea level rose except between 2.8 and 2.3 ka BP (85 cm) that could not be explained.

The $\delta^{18}\text{O}_\text{C}$ from the present study and published literature has been put together for different time slices in the southeastern Arabian Sea to distinguish influence of Bay of Bengal water in the study area (Figure 2B-E, Table 2). The $\delta^{18}\text{O}_\text{C}$, $\delta^{15}\text{N}$, SST, SSS, $\delta^{18}\text{O}_\text{w}$, CaCO_3 , C_org and grain size have been plotted against interpolated ages (Figure 3). We have compared our records with the other published data from the eastern Arabian Sea (Figure 4) to understand the timings of major ISM events and to synthesize region-wide paleo-monsoon sequences. We evaluate the link of ISM with changes in insolation and atmospheric CH_4 content as recorded in the GISP2 record (Figure 5). For a comparison with the Asian summer monsoon during B-A time, we have used $\delta^{18}\text{O}_\text{C}$ data from the Hulu and Timta caves plotted along with those from GISP2 (Figure 6).

Frequency analysis of the $\delta^{18}\text{O}_\text{C}$ record have been performed using the red-noise spectrum analysis of the program REDFIT which is based on Lomb periodogram and mainly appropriate for unequally spaced time series. The significance levels are presented to test reliability of periodicities as shown by spectral peaks (Schulz and Mudelsee, 2002).

Results

Variations in $\delta^{18}\text{O}$

The observed $\delta^{18}\text{O}_\text{C}$ values of *G. ruber* vary between -3.4 and -0.51‰ over the past 16.7 ka (Figure 3). The highest $\delta^{18}\text{O}_\text{C}$ values are expectedly observed in the oldest sediment (-0.6‰) and during the Younger Dryas (YD), whereas lower values characterise B-A event and the Holocene period. The B-A event is marked by lower values of $\delta^{18}\text{O}_\text{C}$, falling rapidly from $\sim -0.6\text{‰}$ at 15.3 ka BP to -1.7‰ at 14.4 ka BP and $\sim -1.5\text{‰}$ at 14 ka BP to -2.0‰ at 13.4 ka BP. The YD around 12.5 ka BP is marked by $\delta^{18}\text{O}_\text{C}$ increase to $\sim -1.3\text{‰}$. Thereafter, at the beginning of Holocene there is an abrupt $\delta^{18}\text{O}_\text{C}$ decrease by 1.9‰ within a short period of 200 years. Within the Holocene (since 10 ka BP) the $\delta^{18}\text{O}_\text{C}$ variations are quite large, ranging between -3.4 and -2.3‰ . Comparatively higher values are observed around 7.5 and 4.5 ka BP with uniformly low values ($\sim -3\text{‰}$) from 7 to 5.6 ka BP (Figure 3).

Variations in C_{org} , $\delta^{15}N$ and $\delta^{13}C$

The C_{org} content ranges between 0.6 and 5%, being lower at the beginning of the Holocene and increasing steadily thereafter towards the late Holocene (Figure 3). The $\delta^{15}N$ of organic matter ranges between 3.9 and 6.6‰ with lower values at ~16 ka BP and decreasing values during YD. The $\delta^{15}N$ and $\delta^{18}O_C$ follow similar trend during the B-A event with slightly elevated C_{org} content (Figure 3). The $\delta^{15}N$ values rose from 4‰ at 11.6 ka BP, just before the beginning of the Holocene to ~6.4‰ at the core top. The $\delta^{13}C$ of organic matter ranges between -22 and -19‰ indicating that the organic matter is dominantly marine.

Sea Surface Temperature (SST), Oxygen isotope ratios of surface waters ($\delta^{18}O_W$) and Sea Surface Salinity (SSS)

The core depth of AAS62/1 is well above lysocline and carbonate compensation depth [Naqvi and Naik, 1983], and Mg/Ca derived SST at the core top is 28.60°C that matches well with the modern SST of 28.58°C (Annual average 1960 to 2012, ICOADS data provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their Web site at <http://www.esrl.noaa.gov/psd/>) that rules out the bias due to dissolution. We are also aware about the limitation in SSS estimation [Rohling, 2000; Steinke et al., 2006] and therefore used combination of SST, $\delta^{18}O_W$, SSS, with other proxies shown in Figure 3 for the interpretation of results.

SST based on Mg/Ca ratio ranges between 25 and 30.5°C, and in general shows an increasing trend from 16.7 ka BP to the core top (Figure 3). The average temperature change for the deglacial and Holocene boundary is ~2.6°C. Between 16.7 and 15.4 ka BP the average SST was low, ~26.1°C, whereas during the B-A event there occurred an increase in SST ranging between 26.3 and 28°C (average 26.8°C). The YD exhibits a lower temperature (26.1 to 26.7°C) except for one point at 12.2 ka BP showing ~28°C. From the YD to Holocene boundary there occurred a temperature change of ~2°C within a period of 1ka. From 11.9 to 9 ka BP average SST was ~ 27.9°C, and from 9 to 0.6 ka BP temperature increased to an average value of 28.2°C.

The $\delta^{18}O_W$ ranges between -1.2 and 0.9‰ (Figure 3). Higher values are observed during 16-15.5, YD, 9.3, 7, 4.5 ka BP, whereas lower values are clearly exhibited during the B-A event and most of the Holocene. Extreme pulses of low values are seen at the beginning of Holocene reaching up to -1.5‰. The SSS at core location varies between 32.9 and 36.7. Higher salinity values are observed during 16.2 to 15.2 ka BP, YD and 4.5 ka BP (Figure 3). Lower salinities were associated with B-A event and most of the Holocene (Avg. 34.8), and during abrupt ISM event (AE).

Calcium carbonate content and Grain size

The CaCO_3 content in the core ranges between 23% and 46% (Figure 3). Relatively moderate increase in CaCO_3 content is observed during B-A event associated with slightly increased C_{org} (Figure 3). Clay ($<4 \mu\text{m}$), very fine silt (4 to 8 μm), and fine silt (8 to 16 μm) together, and medium silt (16 to 31 μm) and coarse silt (31 to 64 μm) together follow similar patterns (Figure 3). Grain size fractions <4 to 31 μm and 31 to 64 μm show inverse relationships. These medium and coarse silts contents reach the highest values between 9 and 7 ka BP, whereas clay and finer silt dominate during the YD, ~6.6 ka BP and 3.5-2 ka BP periods (Figure 3).

Discussion

The sea surface salinity (SSS) and $\delta^{18}\text{O}_w$ are known to be lower near the southwest margin of India than offshore in the Arabian Sea as a result of heavy rainfall and land runoff during the ISM (Figure 1) [Dahl and Oppo, 2006; Govil and Naidu, 2010]. Moreover, the Southeastern Arabian Sea also receives inflow of low-salinity surface water from the southeast (Bay of Bengal water) during the northeast monsoon. In the eastern Arabian Sea SSS during the northeast monsoon is low in the south and increases towards north as this low salinity water mixes with the high salinity water of the Arabian Sea. The *G. ruber* may incorporate $\delta^{18}\text{O}_c$ signature of the ISM and/or the northeast monsoon. Supporting the $\delta^{18}\text{O}_c$ signatures from the intensified northeast monsoon, Tiwari et al. [2005] based on a sediment core study suggested an intensified northeast monsoon that could have led to a more intense flow of low-salinity water from south with low $\delta^{18}\text{O}_c$ into the eastern Arabian Sea during the glacial times. Should this have been the case for the Holocene sediments, the $\delta^{18}\text{O}_c$ values at our core location were expected to be similar or higher than those from the southern core SK-20-185 analyzed by Tiwari et al. [2005] and other cores in this vicinity (Figure 2B-E) as salinity gradient increases from south to north. We have also put together $\delta^{18}\text{O}_c$ of *G. ruber* from the available literature to look for influence of the Bay of Bengal water into the Arabian Sea at different time slices (Figure 2B-E). The $\delta^{18}\text{O}_c$ values of these three slices at the present study location are much lower than the southern cores suggesting that the core site recording more influence of the ISM. Core SK157-4 studied by Saraswat et al., [2012] located south of the study area (Figure 2B), where Bay of Bengal water passes through equatorial region ($2^\circ 40' \text{N}$, $78^\circ 00' \text{E}$) before reaching the Arabian Sea yielded core top $\delta^{18}\text{O}_w$ of 0.3 ‰ (with SSS 34.7) that is much higher than our core top $\delta^{18}\text{O}_w$ of -0.5 ‰ (with SSS 34.7), suggesting that the present core must have incorporated more signatures of precipitation/runoff during ISM. The core AAS62/1 located in the southeastern Arabian Sea is closer to the Indian continental margin (Figure 2B) where there is precipitation and freshwater influx during the ISM (Figure 1A-F) that must have been

incorporated in the $\delta^{18}\text{O}_\text{C}$. Therefore, *G. ruber* from this region that lives throughout the year and calcifies above 80 m depth [Dahl and Oppo, 2006] is expected to record $\delta^{18}\text{O}_\text{C}$, $\delta^{18}\text{O}_\text{w}$ and SSS more signatures of ISM rainfall and runoff. The stable isotope and land-derived (sediment texture) records of AAS62/1 reveal well-resolved environmental variations during the last deglaciation and Holocene. Given the proximity of the core location to the region of highest rainfall along the Indian west coast (Figure 1), the planktonic $\delta^{18}\text{O}_\text{C}$ and $\delta^{18}\text{O}_\text{w}$ probably serve as proxy of precipitation and runoff (Figure 3).

The core AAS62/1 includes a short period of ~ 1.5 ka (bottom 80 cm) before the B-A event. The time series records of $\delta^{18}\text{O}_\text{C}$, Mg/Ca derived SST, $\delta^{18}\text{O}_\text{w}$, $\delta^{15}\text{N}$ and sediment texture during this period exhibit short-term oscillation in the water column characteristics and terrigenous influx. The C_org (1.3 to 2.1%) and CaCO_3 contents (~ 28 to 34%) during this period show minor variations. Between 16.7 and 15.1 ka BP the $\delta^{18}\text{O}_\text{C}$ values exhibit fluctuations (with increases up to 0.4‰) with periodicity of 0.15 to 0.3 ka. This period is marked with increases in $\delta^{18}\text{O}_\text{w}$ by 0.5‰ and SSS by 0.6 (Figure 3). As the study area at the present time is marked with higher freshwater inflow during the ISM (Figure 1), the increases in $\delta^{18}\text{O}_\text{w}$ and SSS during 16.7 to 15.1 ka BP suggest that there must have been reduced freshwater inflow to this region due to weaker ISM. Furthermore from 16.8 ka to 16 ka $\delta^{15}\text{N}$ decreased by ~ 1 ‰ (Figure 3) suggesting reduction in the denitrification intensity. The decrease in denitrification in the past has been related to reduced ISM induced upwelling and a consequent decrease in water column productivity, leading to lower oxygen demand at depth [Altabet et al., 1999]. The ISM induced denitrification ($\delta^{15}\text{N}$) signatures have also been recorded in the eastern Arabian Sea [Ganeshram et al., 2000; Agnihotri et al., 2008; Banaker et al., 2005; Kessarkar et al., 2010]. The decrease in denitrification observed in the present study may also be similarly attributed to a less intense ISM. To check, if this signature of weaker ISM has also been archived in other cores of the eastern Arabian Sea we have put together $\delta^{18}\text{O}_\text{C}$ data from the present study with other published records (Figure 4). It appears that major variations in the $\delta^{18}\text{O}_\text{C}$ did not occur during this period in the eastern Arabian Sea. Higher $\delta^{18}\text{O}_\text{C}$ values in the core SK 17 (Figures 1 and 4) during this time have been related to weak ISM and associated reduced runoff from the land [Anand et al., 2008]. Weaker ISM during similar period has been inferred from the wind proxy [Sirocko et al., 1993; 1996] and reduced denitrification rates in the northern Arabian Sea have been related to midwater ventilation by advection of more oxygenated Antarctic Intermediate Water during Heinrich events [Suthhof et al., 2001].

Studies on the western Arabian Sea based on geochemical and multi-tracer approach suggest that the ISM intensified in three succession events with first one at ~ 16 ka BP [Sirocko et al., 2000], whereas study in the eastern Arabian sea (off Goa) based on Mg/Ca ration suggest intensification of

ISM ~17 ka BP [Anand et al., 2008]. A recent study by Naidu and Govil [2010] suggests warming in the northern Indian Ocean around 19 ka which is synchronous with deglacial warming in Antarctica. As the warming occurred, the ISM was also intensifying. Our results suggest that the increase in ISM intensity was slow with minor fluctuations till 15.2 ka BP. Thamban et al. [2001], have reported fluctuations in $\delta^{18}\text{O}_\text{C}$ values around 15.2 ka BP that were attributed to increase in precipitation and warming of sea surface. Further the composite oxygen isotope section of cores AAS62/1 and AAS62/2 and other results from the Southeast Arabian Sea (Figure 4) suggest major intensification of ISM could have started after the LGM around 15.2 ka BP. This is consistent with the results of Naqvi and Fairbanks [1996] who reported a cessation of deep water outflow from the Red Sea presumably in response to a major intensification of the southwest monsoon around 15.5 ka B.P.

The B-A event is prominently seen in the proxy records with lower $\delta^{18}\text{O}_\text{C}$, $\delta^{18}\text{O}_\text{W}$ and SSS, and an increase in coarser grained terrigenous sediments. From 15.2 to 13.4 ka BP there is stepwise decline in $\delta^{18}\text{O}_\text{C}$ by $\sim 1.4\text{‰}$, $\delta^{18}\text{O}_\text{W}$ by 0.98‰ and SSS by 2.08. The declining $\delta^{18}\text{O}_\text{C}$, $\delta^{18}\text{O}_\text{W}$ and SSS clearly suggest increase in precipitation/runoff, as the study area is in the vicinity of high precipitation and runoff (Figure 1A-F). The lower values of $\delta^{18}\text{O}_\text{C}$ are very well correlated with increase in coarser grain size fraction (medium to coarse silt), supporting increase in land runoff hence increased terrigenous material. The $\delta^{18}\text{O}_\text{W}$ and SSS are similar to the present (surface sediments). Another core, AAS62/2, located in the area but away from the margin, also shows an increase in $\delta^{18}\text{O}_\text{C}$ by 1‰ during this period (Figure 4). The fluctuations in $\delta^{18}\text{O}_\text{C}$ are not well resolved in core AAS62/2 during this period (Figure 4), which could be due to low sedimentation rate. The C_org (1-2%) and CaCO_3 (24-31%) show slightly elevated values during this period. The $\delta^{15}\text{N}$ shows stepwise increases by $\sim 2.4\text{‰}$ from 16 ka to 13.4 ka suggesting increase in denitrification rate. This period is known to be the period of a more intensified monsoon. A stronger monsoon is expected to have been associated with higher productivity, in turn sustaining elevated mid-depth denitrification, similar to the present (surface sediments). The intensity of the summer monsoon seems to have been increased at the advent of B-A event. Increase in the ISM strength during 15.3 to 14.7 ka BP has been reported by Overpeck et al. [1996] who related it to the warming of the Tibetan Plateau and associated strengthening of the pressure gradient needed for a strong monsoon.

The $\delta^{18}\text{O}_\text{C}$ reaches the lowest values $\sim 2.0\text{‰}$ at about 13.4 ka BP (Figure 3) and starts increasing stepwise to 1.3‰ at 12 ka BP. Similarly there is increase in $\delta^{18}\text{O}_\text{W}$ by 0.88‰ and SSS by 1.55, suggesting decrease in freshwater influx (precipitation/runoff) into the study area. This indicates that there was an abrupt decrease in ISM intensity that coincided with the onset of the YD. This

increase in $\delta^{18}\text{O}_\text{C}$ is also associated with decrease in $\delta^{15}\text{N}$ suggesting suppressed denitrification during the YD.

The ISM system acquires energy from the subtropical Indian Ocean whereas the East Asian system does so from the western Pacific Warm Pool [Wang et al., 2005]. The Indian monsoon is often considered to be part of the Asian monsoon. A compilation of the Holocene paleo-studies by Staubwasser and Weiss [2006] revealed that the South Asian monsoon prevailed in the Himalayas and the equatorial Indian Ocean and not in the Indian Peninsula. These authors suggested that the south Asian summer monsoon rain was not uniformly enhanced during the early and mid Holocene. The B-A event has been well documented in marine cores from the northern [e.g. Altabet et al., 2002] and western [e.g. Ivanochko et al., 2005] Arabian Sea. Whereas very few cores from the southeastern Arabia Sea have preserved the B-A event signatures [Rao et al., 2010; Tiwari et al., 2005; Singh et al., 2006] with relatively low resolution. The present data set with moderately high sedimentation rate (53-99 cm/kyr) during the B-A event provides evidence for increase in the ISM. A comparison of $\delta^{18}\text{O}_\text{C}$ data from AAS62/1 is shown with very high resolution data from the Timta cave, Hulu cave, and GISP2 (Figure 6). The moisture to the AAS62/1 site is derived from the Arabian Sea, to Timta caves from the Bay of Bengal and Hulu caves from the Pacific Ocean. These three regions are connected by the ITCZ; the movement of the position of the ITCZ may have resulted in changes in monsoonal precipitation. Previous studies by Sinha et al. [2005] in the Himalayan region suggested coupling between the Asian monsoon and the climate of the North Atlantic. Present core under study may have effected by changes in sedimentation rates and dating error; higher resolution core from this area would bring out better picture of the relation between these three regions.

The increased ISM strength during the B-A event seems to have been interrupted by a short period of Younger Dryas, which is marked by high $\delta^{18}\text{O}_\text{C}$, $\delta^{18}\text{O}_\text{W}$, SSS and low SST, indicating reduced precipitation/runoff to the study area. The $\delta^{15}\text{N}$ values also decreased during this time suggesting less intense denitrification compared to the B-A period. Reduced denitrification during this time might be the result of more oxygenated intermediate waters due to advection of Antarctic intermediate waters during YD [Suthhof et al., 2001; Pichevin et al., 2007]. Further, reduced coarser grain-size (medium and coarse silt) suggests lower input from the land, which could also be linked to weaker ISM.

The record shows three abrupt shifts in $\delta^{18}\text{O}_\text{C}$ to very low values toward the end of the Younger Dryas. Low $\delta^{18}\text{O}_\text{C}$ values observed here could be the result of diagenetic alteration in *G. ruber*. We, however, did not observe any visible evidence of recrystallization or dissolution on *G. ruber* and the samples were well preserved with abundant pteropods. Planktonic foraminiferal population variability

could also give lower values. As we have chosen monospecific species of *G. ruber* of specific size range and about 10 specimens were taken per run to get the average values there are reduced chances of error. These low values for example, from 12 to 11.8 ka BP the $\delta^{18}\text{O}_\text{C}$ fell from -1.34‰ to -3.29‰ . A shift by -1.9‰ within a period of ~ 200 years could be caused largely by lower salinity arising from greater rainfall/land runoff that has been reflected in reduced $\delta^{18}\text{O}_\text{W}$ and SSS. Three such pulses of massive freshwater inputs are recorded in our core, centred around 11.7, 11.6 and 11.2 ka BP, suggesting high variability of monsoonal rainfall during this period (see AE Figure 3). A similar shift in $\delta^{18}\text{O}_\text{C}$ is also observed (as a single peak, perhaps due to poorer resolution) in the nearby core AAS62/2 (Figure 4). If we compare these data with the atmospheric methane content record from GISP2 (Figure 5), a high methane concentration seems to be associated with the intensified ISM. However, during this time denitrification rate appears to have been at its lowest that needs explanation. During increased ISM intensity, high precipitation/runoff in the study area may have resulted in stratification due to water column salinity gradient. During ISM there exist under current with more oxygenated water moving north in the study area. At the times of intensified ISM this current must have been stronger and may be keeping the subsurface water oxygenated and hence reducing denitrification [Kessarkar et al. 2010.]. Increase in the monsoon strength during 11.5-10.8 ka BP has been reported by Overpeck [1996], and more humid conditions around 11 ka BP have been recorded by Cancer et al. [2005] in the Nilgiris (western Peninsular India). Based on the sediment core from the Andaman Sea Naqvi et al. [1994] have reported shift in $\delta^{18}\text{O}_\text{C}$ and $\delta^{13}\text{C}$ of benthic foraminifera after the YD and related to the monsoonal changes. Similar oscillations at the end of the YD have also been reported from another core in the Andaman Sea and related to increase in the Irrawaddy River runoff and Indian Ocean Monsoon [Rashid et al., 2007]. Based on an evaluation of the published data Morrill et al. [2003] concluded that toward the end of the Pleistocene (~ 11.5 ka BP) Asian monsoon precipitation increased dramatically supporting a link between the North Atlantic climate and Asian monsoon [Wang and Fan, 1987; Overpeck, 1996]. Warming and an increase in Asian monsoon intensity at about the same time are also indicated by the $\delta^{18}\text{O}_\text{C}$ record from the South China Sea [Steinke et al., 2006]. Further, an increase in deepwater formation occurred around this period (within the limits of radiocarbon dating) in the North Atlantic suggesting atmospheric teleconnections between the two regions. Studies on stalagmites from the Oman point to monsoon maximum during the early Holocene [Neff et al; 2001; Fleitmann et al., 2003]. Comparing our oxygen isotope data with the 30°N insolation and methane data from GISP2, it would appear that intensification of Asian monsoon and enhanced methane production occurred in response to insolation maximum in the northern hemisphere. The YD interrupted this trend, however, and when the YD perturbation faded there was an abrupt intensification of the monsoons that peaked with the insolation maximum at 30°N (Figure 5). Superimposed on the insolation-forced increase in the

monsoon, there have been short term changes as evident from the abrupt freshwater discharge events recorded by us toward the end of the YD. Examination of such decadal to centennial scale changes and understanding of their causes require high sedimentation records during key time slices [e.g. Gupta et al., 2003; Kurian et al., 2009].

Presence of lower $\delta^{18}\text{O}_\text{C}$ suggests warmer and lower salinity conditions after 11 ka BP. The decrease in $\delta^{18}\text{O}_\text{C}$ between 11 and 10 ka BP along with the presence of higher coarser grain size during this time suggest increase in land derived material. The $\delta^{18}\text{O}_\text{C}$ has fluctuated by $< \pm 0.5 \text{ ‰}$ (Figure 3) since 10.5 ka B.P.; by contrast, the $\delta^{15}\text{N}$ values increased steadily up to ~ 3 ka B.P. suggesting intensification of denitrification through most of the Holocene following a minor minimum that occurred shortly after 10 ka B.P. Similar increase in $\delta^{15}\text{N}$ has been reported by Pichevin et al., [2007] and suggests that northeastern Arabian Sea influence Holocene climate. The C_{org} and CaCO_3 content also follow the increasing trend, like that of $\delta^{15}\text{N}$ and do not follow the insolation changes. Increase in ISM has been discussed by Tiwari et al. [2010] during late Holocene that does not follow insolation and suggest that the internal feedback processes to be equally important. The intensification of ISM has been reported to occur in two steps i.e. 9.4 and 9.1 ka BP based on data from the Arabian Sea [Gupta et al., 2003]. Analysis of a core off India at $17^\circ 45' \text{N}$ suggested that the main event occurred between 9.1 and 8.5 ka BP [Thamban et al., 2001]. The present data set derived from a region that comes under the direct influence of the southwest monsoon does not indicate very large changes, but between 9.1-8 ka BP $\delta^{18}\text{O}_\text{C}$ did decrease by 0.4 ‰ , associated with relatively invariable $\delta^{18}\text{O}_\text{W}$ and SSS. This was also accompanied by an increase in the amount of medium and coarser silt. Other workers [Neff et al., 2001; Gupta et al., 2003] reported abrupt changes in summer monsoon frequently occurring during the Holocene.

Between 7 and 5.6 ka BP $\delta^{18}\text{O}_\text{C}$ values are nearly constant ($\sim -3 \text{ ‰}$), with very small variation in $\delta^{18}\text{O}_\text{C}$ and SSS suggesting stable ISM, that is in agreement with other records from this region (Figure 4). The period from 7.2 to 6.5 ka BP is regarded as having being optimal for vegetation growth based on the pollen record, corresponding to the second humid phase [Van Campo et al., 1996]. Reduction in the input of the coarser fraction at this time can be explained by local changes in the sea level that was close to or above the present mean sea level [Mathur et al., 2004]. The increase in the $\delta^{18}\text{O}_\text{C}$, $\delta^{18}\text{O}_\text{W}$, SSS and decrease in SST ~ 5.5 ka BP suggests weakening of ISM with the minimum centred at 4.5 ka BP. This is the period of reduction in freshwater flux by the Indus River [Staubwasser et al., 2003]. After 4.5 ka BP the lower $\delta^{18}\text{O}_\text{C}$, $\delta^{18}\text{O}_\text{W}$ and SSS suggest the end of the dry period with intensification of ISM with a peak at about 3 ka BP. Observation of higher $\delta^{18}\text{O}_\text{C}$, $\delta^{18}\text{O}_\text{W}$ and SSS at 2.5

to 2.3 ka BP suggests a dry phase. This is also evident from other records from the Southwest Indian margin [Thamban et al., 2007]. Reduced precipitation during this period is consistent with data from Dongge cave in South China (Wang et al., 2005).

Spectral analysis of oxygen isotope data $\delta^{18}\text{O}_\text{C}$ was carried out to look for any cyclicity in the paleo monsoon strength and understand controlling factors using the Redfit program [Schulz and Mudelsee, 2002]. The $\delta^{18}\text{O}$ of *G. ruber* exhibits periodicities of 390, 500 and 620 years above the 95 % confidence level (Figure 7). The 500 year periodicity has also been reported by Bhushan et al. [2001] and attributed to the modulation of monsoon by the Atlantic deep water formation. Periodicity of 512 yrs has been reported by Wang et al. [2005]. Considering the ITCZ connection of the Indian monsoon with the East Asian monsoon, this periodicity of 500 could be related to the position of the ITCZ.

Conclusions

Post-glacial intensification of ISM with intensification of subsurface denitrification appears to have begun at about 15.2 ka BP coinciding with the onset of the B-A event. The ISM intensification was coupled with strengthening of the Asian summer monsoon and warming in the North Atlantic. The trend of intensifying monsoons, probably forced by the increase in northern hemisphere summer insolation, was however interrupted by the YD cooling that also led to a decrease in denitrification intensity. As the influence of the YD perturbation faded, a rapid intensification of ISM occurred with three very wet events of short duration occurring between 11.8 and 11.2 ka BP. Changes in ISM also occurred during the Holocene with relatively arid conditions prevailing around 4.5 ka BP, and humid conditions centred around 3 ka BP. The ISM was relatively stable between ~ 7 and 5.6 ka BP. Subsurface denitrification, in general, appears to have undergone an increase through the Holocene. The observed periodicity of 500 years is comparable to that of North Atlantic deep water formation and the East Asian monsoon.

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Caption for the Figures

Figure 1 (A-C) Monthly average of Rainfall, (D-F Monthly average freshwater influx (Evaporation-Precipitation) [Fennig et al., 2013]). Note that there is increase in freshwater flux on the southwestern margin of India during the southwest monsoon. (G) Thick circles are core locations showing the core AAS62/1, AAS62/2 [Rao et al., 2008], SK148/4 [Rao et al., 2010], SK17 [Singh et al., 2006] numbers in the brackets are water depth. Inset figure shows location of core AAS62/1, Timta and Hulu caves, contours of onset dates of Indian summer monsoon. Thick lines indicate Indian and Asian monsoon trajectories [modified after, Sinha et al., 2010].

Figure 2 (A) Calibrated radiocarbon ages and sedimentation rates (cm/ka). (B-E) Variations in $\delta^{18}\text{O}_\text{C}$ (*G. ruber*) in the southeastern Arabian Sea, (B) Core locations, (C) surface values, (D) between 5.6-7 ka BP when monsoon was stable (E) 11-12 ka BP there was abrupt event of ISM monsoon, data points were taken based on the available analysis during these time frames (see Table 2).

Figure 3 Temporal variations in the $\delta^{18}\text{O}_\text{C}$ record of *G. ruber*, $\delta^{15}\text{N}$ of C_{org} , Sea surface temperature (SST) in $^{\circ}\text{C}$, $\delta^{18}\text{O}_\text{W}$, Sea surface salinity (SSS) C_{org} , CaCO_3 content and grain size of land derived material. ISM-Indian summer monsoon, B-A- Bølling-Ållerød, YD- Younger Dryas, AE- abrupt ISM event, Grey bands indicate intensification of ISM. Vertical short bar denotes estimated error.

Figure 4 Comparison of $\delta^{18}\text{O}_\text{C}$ records from the southeastern Arabian sea SK148/4 [Rao et al., 2010]; AAS 62/1 (present study) vertical short lines are AMS ages of this core; AAS62/2 [Rao et al., 2008] vertical dotted lines are AMS ages of this core; SK17 [Singh et al., 2006],

Figure 5 Comparison of $\delta^{18}\text{O}_\text{C}$ of AAS 62/1 with the methane content of GISP2 core and June solar Insolation maximum at 30°N

Figure 6 Comparison of $\delta^{18}\text{O}_\text{C}$ of AAS 62/1, with $\delta^{18}\text{O}$ of GISP2 (Bidecadal Oxygen Isotope Data), Timta and Hulu caves [Stuiver et al., 1997; Sinha et al., 2005; Wang et al., 2001]. Grey bands indicate intensification of the ISM.

Figure 7 Power spectra of oxygen isotope time series of *G. ruber*. Horizontal line on the upper right shows 6db bandwidth, dotted line indicates 95 % confident line.

Table 1. Accelerator Mass Spectrometer (AMS) ages of sediment intervals dated in the core AAS-62/1

Depth interval in the core (cm)	Lab code at WHOI	Radiocarbon Age (years B.P.)	Calibrated Age Range (Years B.P.)
6-8	57869	1770 $\pm$ 35	1177
18-20	50064	2720 $\pm$ 40	2252
40-42	57870	3180 $\pm$ 35	2848
54-56	57871	4510 $\pm$ 25	4552
74-76	50065	6420 $\pm$ 45	7347
120-125	57872	9520 $\pm$ 45	10268
170-175	50066	10500 $\pm$ 60	11463
270-275	50067	11050 $\pm$ 55	12476
340-345	50068	12400 $\pm$ 75	13690
495-500	54961	14450 $\pm$ 65	16569

Table 2a. Location of cores and $\delta^{18}\text{O}_\text{C}$ (*G. ruber*) values at different time intervals that are used in Figure 2.
 *Values were extracted from the figures. #Average values for 11-12 ka BP in cores AAS62/1 and AAS62/2

	Core	Location		Age ka BP	$\delta^{18}\text{O}_\text{C}$ ‰ (<i>G. ruber</i>)
Saraswat et al., [2012]*	SK157/4	2°40'N	78°00'E	Surface	-1.9
				6	-2.0
				11.1	-2.18
Tiwari et al., [2005]*	SS3827G	3°42'N	75°54.5'E	Surface	-2.7
				6	-2.5
				12	-1.6
Rao et al., [2010]	SK148/4	8°12'	75°54'	Surface	-2.9
				5.6	-2.5
				12	-1.8
Pattan et al., [2003]*	SK129/CR-05	9°21'N	71°59'E	Surface	-2.5
				6	-2.5
				11.5	-0.95
Thamban et al., [2001]*	GC5	10°23'	75°34'	Surface	-3
				5.6	-2.5
				12	-1.5
Sarkar et al., 1990	SK-20-185	10°N	71°50'E	Surface	-1.85
Present study	AAS62/1	11°30'N	74°37'N	Surface	-3
				6	-3
				11.8	3.3
				11-12 [#]	(2.2) [#]

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7