Chapter 4 Discussion
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4.1 Terrigenous input in the southeastern Arabian Sea

The chronology of the sediment core SK-215/4 revealed that the core extends up to a time span of ~67 kyr BP. Linear interpolation between the $^{14}$C ages reveals variations in the sedimentation rate in the deep water core of SK-215/4 as 3.67 cm/kyr during the present interglacial and 5.26 cm/kyr for Last Glacial Period and MIS 3 and 4 record higher sedimentation rate of 7.46 and 7.67 respectively. These sedimentation rates correlate with early deep water sediment core studies from the southeastern Arabian Sea (Paropkari et al., 1991; Sarkar et al., 2000; Pattan et al., 2005). The core SK-215/5 extends up to the Late Glacial Period (13.5 kyr BP) and this core has 5 age control points which reveal different sedimentation rates: late Holocene - 15.4 cm/kyr, early Holocene -24.8 cm/kyr and late Glacial period records 31.16 cm/kyr. SK-215/5 record is comparable with sedimentation rate of shallow depth sediment core GC-5 (water depth: 280 m) studied by Thamban et al., (2001). The major difference between these two records is SK-215/5 shows an abrupt increase of sedimentation rate from late Glacial to early Holocene which is not evident in GC-5. This mismatch may be due to high intensity monsoon influenced high fresh water flow which might have increased the depositional rate of SK-215/5 when compared to the shallow depth. For the present study, I have used sediment core SK-215/5 to know the terrigenous influence in southeastern Arabian Sea because the terrigenous influence is more in shallow water area when compare to the cores from deeper waters. Further, this core has high sedimentation rate which helped me to reconstruct the past terrigenous input in this region with high resolution.

In general, unlike in the western Arabian Sea where lithogenic input is mainly through aeolian pathways from the Arabian Peninsula, such an input in the eastern Arabian Sea is largely fluvial from the Indian subcontinent (e.g., Paropkari, 1990; Shimmield et al., 1990). Reconstructions of high-resolution terrigenous input record of sediment geochemical data, in particular in the eastern Arabian Sea is, however, very limited (e.g., Thamban et al., 2007). Literature survey also reveals that existing paleoceanographic studies in the eastern Arabian Sea are largely dealing with past productivity, clay mineralogy and hydrographic changes (Naidu, 1991; Paropkari et
Indeed, less attention has been paid to monsoon-detrital input link in the past, although such information is essential in the eastern Arabian Sea, where large-scale variations in primary productivity may dilute terrigenous signals that are related to Indian summer monsoon fluctuations, at least during the Holocene interglacial.

Since the type and amount of terrigenous material depend on climatic conditions on the continent (Pattan et al., 2005), there must be a linear relationship between the Indian monsoon and terrigenous sediment deposition in the eastern Arabian Sea. A 420 cm long sediment core collected at a water depth of 460 m from the western continental margin, off Cochin, was therefore subjected for detailed investigation of sediment geochemistry. All the geochemical data discussed in this chapter are listed in the appendix. The most important aspects of the data set are presented and discussed here.

### 4.1.1 Sedimentary Bulk Composition Record

Depth profiles of major and trace elements analysed in the sediment core SK215/5 are shown in Figures 3.3, 3.4, and 3.5. The analysed elements are separated into two categories, detrital and biogenic, based on their prime characteristics. For example, elements contributed essentially by terrigenous input have pooled together in Figures 3.3 and 3.4 and biogenic elements in Figure 3.6. Terrigenous proxy elements such as Si, Al, Mg and K show, in general, an overall decreasing trend for the last ~13.5 kyr BP (Figure 3.1). All these elements show higher values between ~13.5 and 7 kyr BP and lower values are evident since ~7 kyr BP. Similar to this, detrital ratios Si/Al and Al/Ti exhibit higher ratios in the former interval of Late Glacial and early Holocene, while lower ratios are associated with sediments of mid- and late Holocene sediments (Figure 3.3). These all indicate higher terrigenous input during Last Glacial and early Holocene between ~13.5 and ~7 kyr BP and lower terrigenous input between ~7 kyr BP to the Present (Figure 3.3). Down-core profiles of all detrital elements and their ratios, especially Si/Al and Al/Ti, show distinct peaks at ~13.3, 11.9, 10.9, 9.2 and 7.4 kyr BP during the intervals of Late Glacial and early Holocene. Higher ratios of Al over Si and Ti during these times suggest intense chemical weathering in the Indian subcontinent and deposition of clay-rich sediments.
(Figure 3.1) most likely influenced by a strong summer monsoon. Invariably, all these proxies are decreasing towards late Holocene, suggestive of decreased summer monsoon during the course of Holocene. Such an interpretation is consistent with available monsoon reconstructions in the eastern and western Arabian Sea as well as other low-latitude summer monsoon reconstructions in south and east Asian regions (Agnihotri et al., 2002; Thamban et al., 2007; Gupta et al., 2004; Selvaraj et al., 2007). Consistent to the behaviors of above mentioned detrital proxies, other detrital elements, for example Ti and Ga also show higher values between ~13.5 and 7 kyr BP and lower values since ~7 kyr BP (Figure 3.5). Similar to detrital proxies, high concentrations of Fe and Mn, both elements in general used to detect diagentic changes in sedimentary column, in the lower part of the core, also revealing their association with detrital fraction. Depth profiles of Fe and Mn, however, mimic Ti and Ga rather than Si and Al (Figures 3.3 and 3.5). All four elements (Ti, Ga, Fe and Mn) show more or less constant values with less fluctuations compared to Si, Al, Mg and K in sediments deposited during the mid and late Holocene intervals suggesting stable monsoon climate and related silicate weathering for the last ~7 kyr BP. The close association among the former detrital group indicates their unique provenance either from lateritic rocks or Deccan basalts; both are major rock types which occupied the large areas of coastal plains and high lands in western part of India where the summer monsoon is a prime climate factor responsible for terrigenous supply to the eastern Arabian Sea.

In contrast to the detrital elements, depth profiles of biogenic elements such as Ca, P, and Sr exhibit an opposite patterns i.e., lower values are evident between ~13.5 and 7 kyr BP and higher values are seen since ~7 kyr BP (Figure 3.6). These patterns of biogenic elements reveal that these elements appear to be diluted by the input of detrital sediments during the early Holocene when the monsoon was intense. All these proxies show distinct shift from low to high values between ~8.1 and 6.3 kyr BP. This shift indicates a major change in sedimentary delivery to the study site in the eastern Arabian Sea due probably to a transition from strong to weak Indian summer monsoon during the mid-Holocene. The data further show that carbonate contents (CaCO$_3$) are generally lower (Figure 3.1) and terrigenous elements (Si, Al, K and Ti) are higher during the interval of intense Indian monsoon. A significant shift in carbonate and terrigenous elements occurs between ~7 and 6.3 kyr BP correlating with strong to weak summer monsoon shift during the mid-Holocene in India.
Figure 4.1: Diagrams show the correlation between the elements (a) Si with Al, (b) Ti with Al, (c) Mn with Al, (d) K with Al, (e) Fe with Al, (f) Na with Cl.
4.1.2 Silicate weathering record

Monsoon related weathering intensity can affect the relative abundances of major elements in the terrigenous detrital fraction of sediments (e.g., Wehausen et al., 2003). Since the analysed sediment core dominated by silt and clay size materials (Figure 3.1), the percentage of terrigenous input using the concentration of Al in sediments (terrigenous %-Al) has been calculated with respect to the concentration of Al in average shale (10%; Turekian and Wedepohl, 1961). Down-core variation of terrigenous %-Al shows in general a decreasing trend for the last ~13.5 kyr BP with higher terrigenous input during the late Glacial and early Holocene intervals (Figure 3.3). Lower terrigenous input is, however, evident during the last ~7 kyr BP. Importantly, clear millennial-scale peaks of terrigenous Al are evident between ~13.5 and 7 kyr BP. These terrigenous peaks are centered at ~13.3, 11.9, 10.9, 9.2, and 7.4 kyr BP. To be consistent, detrital elements such as Si, Al, Mg, and K, and detrital ratios, such as Al/Ti and Si/Al, show similar peaks in tandem with % of terrigenous input. These all indicate higher terrigenous input and thus higher content of clay in the study area, especially during the late Glacial and early Holocene periods. Although the core location is not proximal to any major river input, the millennial-scale peaks evident in the records of detrital proxies, suggesting a strong Indian monsoon and associated higher terrigenous input during these time periods. This inference is further substantiated by sedimentary K/Si ratio, a suitable proxy for silicate weathering of continental rocks, which shows higher K over Si in sediments (higher content of illite) deposited prior to ~7 kyr BP when compared to sediments deposited after ~7 kyr BP when Si dominates K (likely by lower content of illite clay) which is in well agreement with clay minerals record of Thamban et al., 2002 from the southeastern Arabian sea. This likely indicates the relative enrichment of K in the former time intervals attributed to intense silicate weathering when the Indian monsoon was strong (Gupta et al., 2004; Tamban et al., 2007).

Aluminum is believed to be an index element for the terrigenous source (e.g. Nath et al., 1989) and both Ti and Al are refractory elements which are extremely immobile in the marine environment (Bischoff et al., 1979). Si and Al display similar down-core patterns with a perfect positive correlation ($R^2=0.99$), which shows that there is no significant contribution of biogenic silica to the sediment. Moreover, Al
Figure 4.2: Diagrams show the correlation the elements (a) Co with Ti, (b) Ga with Ti, (c) Cr with Ti (d) Pb with Ti (e) Y with Ti (f) Cu with Ti
shows positive correlation with Ti (0.73), K (0.83), Fe (0.70), and Mn (0.69) indicating that most of the elements are associated with aluminosilicates of continental origin and thus a common source (Figure 3.6 and 4.1).

Na shows dual negative trends with Al i.e. high Al and low Na and vice versa. Such trends indicate that when the monsoon was high, Na leaching was also high and dissolved Na$^+$ ions may adsorbs on the clay particles in the sediments. Higher Al and lower Na and their negative correlation suggest that during the early Holocene, clay production in the continental region that was influenced by the Indian summer monsoon was likely to be high. Conversely, sediments deposited after 7 kyr BP show low Al and Na contents due to weak monsoon and associated low weathering of silicate rocks and thus low clay production. The strong correlation between Na and Cl ($R^2=0.94$) (Figure 4.1 and 3.7) indicate adsorption of sea salt with clay content of the sediments which is relatively high in sediments deposited in late Glacial and early Holocene intervals.

4.1.3 Trace elements record

Depth profiles of all the trace metals investigated are given in Figure 3.4 and 3.6. Ba/Al ratio shows less than $70 \times 10^{-4}$, a maximum value of background Ba/Al ratio calculated from the composition of the Earth’s upper crust (Taylor and McLennan, 1985) or average shale (Wedepohl, 1991) and other marine geochemical studies (Wehausen and Brumsack, 1999; Shimmield and Mowbray, 1991) throughout the core. This suggests that the entire Ba is derived from lithogenic input. Furthermore, Ba addition through water column productivity in the core site is either very minimum or nil. This inference also corroborates with Ba/Al values of studied sediments with carbonate, which do not show any relationship. Calcium, P, and Sr, exhibit positive correlation among themselves, suggesting a biogenic source and all seem to act as dilutants to the detrital elements.

It is apparent from Figure 3.5 that down-core variations of most of the trace metals Co, Cr, Cu, Ga, Rb, V, Y, and Zn, mimic down-core trend of Ti rather than Al. The peak values observed in the profiles of Al and terrigenous % of Al during late Glacial and early Holocene between ~13.5 and 7 kyr BP are almost absent in depth
Figure 4.3: Diagrams show the correlation of the elements (a) Ni with Ti, (b) V with Ti (c) Sr with Ti (d) Sr with CaCO₃ (e) Zr with Ti (f) Zr with CaCO₃
profiles of Ti as well as these trace elements. Bi-plots of all these trace elements, therefore, have been drawn against Ti (Figure 4.2, 4.3). Except Br and Zr, almost all the elements correlate well with Ti. Titanium shows a strong positive correlation with Co ($R^2=0.99$), Ga ($R^2=0.94$), Rb ($R^2=0.94$), and Y ($R^2=0.94$) indicating terrigenous source for these elements (Figure 4.2) Other trace elements Cr, Cu, Pb, Ni, and V exhibit dual correlations with Ti. In sediments deposited during the Late Glacial and early Holocene, these elements have correlation coefficients ($R^2$) of 0.89, 0.77, 0.68, 0.78, and 0.82, respectively, with Ti (Figure 4.2, 4.3) and thus indicate their derivation from intensively weathered provenance. Whereas, insignificant correlations of these elements with Ti evident in sediments deposited from ~7 kyr BP to the Present (Figure 3.7) suggesting less weathered nature of provenance rocks. These interpretations are consistent with the concentration of Ti which shows higher concentrations (>0.45%) in sediments deposited in the warm and wet early Holocene. But the same element shows <0.45%, equal to Ti concentration in the Earth’s upper crust (Taylor and McLennan, 1985), in sediments deposited from mid- through late Holocene. The higher values of Ti probably attributed by intense weathering of Deccan basalts rich in Ti (Turekian and Wedephol, 1961) in the former intervals when the monsoon was intense; an interpretation consistent with Sirocko et al. (2000). The distinct separation of trace elements during these intervals are further suggestive of Indian summer monsoon fluctuations which might have influenced the detrital delivery to the eastern Arabian Sea through chemical weathering of rocks and therefore, concentration of these elements in sediments is more. Furthermore, the above inferences are substantiated by Rb and Sr relationships with Ti. Rubidium essentially associates with lithogenic materials in the marine environment, showing a strong positive correlation with Ti ($R^2=0.94$). Strontium, however, negatively correlated with Ti ($R^2=0.96$) because of its biogenic origin and therefore, Sr shows extreme positive correlation with carbonate content of sediments ($R^2=0.99$) studied (Figure 4.3).

Zirconium, an immobile element during chemical weathering, attains relative enrichment in the warm and wet climate when other minerals are dissolved easily by hydrolysis. Moreover, Zr is found in silicate rocks, where it replaces Ti, rare earth elements, Fe, Ca and is enriched in heavy minerals such as ilmenite or rutile (Zr content: ~1000 ppm; Sirocko et al., 2000). Very surprisingly, depth profile of Zr
neither corresponds to Ti nor correlates with Al; two important elements of detrital origin in the studied core \( (R^2=0.74) \). The strong negative correlation coefficient of Zr with Ti \( (R^2=0.73) \) and low negative but significant correlation with Al \( (R^2=0.48) \) further support that Zr input is not from detrital source. Interestingly, Zr correlates positively with CaCO\(_3\) \( (R^2=0.88) \) and Sr \( (R^2=0.87) \) (Figure 4.3) and thus indicating a biogenic association. According to Sirocko et al. (2000), the high Zr content could be related to detrital zircon or inclusions of Zr in silicates and heavy minerals. Nonetheless, we do not know the possible mechanisms behind Zr-Sr-CaCO\(_3\) association in these sediments.

High concentration of terrigenous input between \( \sim \)13.5 and 7 kyr BP and low terrigenous input since \( \sim \)7.7 kyr BP inferred from the data set suggests high chemical weathering during late glacial to early Holocene section due to high intensity monsoon and the late Holocene low chemical weathering due to low intensity monsoon. All this study revealed that terrigenous input in the southeastern Arabian sea is contributed by the adjacent hinterland and the terrigenous input is largely controlled by the climatic variation in the study area.

4.2 Reconstruction of climatic/oceanographic conditions of SEAS

4.2.1 Reconstruction of Regional Climate

Climatic inferences of this study is mainly based on geochemical data of sediment core SK-215/5 because the chronology obtained for this core revealed that sediment accumulation rate is sufficient to resolve the climatic variation in the past. Geochemical data of the core SK 215/5 revealed that the most impressive, well-resolved climatic variations during late Glacial to present where that the climatic variation is very dynamic after the last glacial period.

Since I have discussed in detail about terrigenous elements proxy and its validity for climatic inferences in an earlier section (see section 4.1) here I highlight the climate record of this proxy to infer past climate changes. Down core profiles of terrigenous major elements such as Si, Al, Mg, and K show higher terrigenous input during late Glacial and early Holocene between \( \sim \)13.5 and 7 kyr BP and lower terrigenous input from 7 kyr BP to the Present (Figure 3.3). Profiles of all detrital
elements and elemental ratios Si/Al and Al/Ti also show distinct peaks at \(-13.3, 11.9, 10.9, 9.2, 7.4, 4.6, \text{ and } 1.8 \text{ kyr BP}\) during the Late Glacial to present, and its correlation with high intensity monsoon events record of other proxy suggest these terrigenous proxy has well recorded the climate shift since late glacial to present along with this down core variations of most of the trace metals, Co, Cr, Cu, Ga, Rb, V, Y, and Zn exhibit higher values between \(-13.5 \text{ and } 7 \text{ kyr BP}\) and lower values since 7 kyr BP. This pattern is very consistent with higher terrigenous input during the former period and lower terrigenous input in the latter interval in the study area.

Na shows negative correlation with Al i.e. high Al and low Na and vice versa, but perfect positive covariance with sedimentary Cl concentration. Such correlation indicates that when monsoon was high, Na leaching was also high and dissolved Na\(^+\) ions may adsorb on the clay particles in the sediments. Higher Al and lower Na and their negative correlation (Figure 4.4) suggest that during early Holocene clay production in the continental region was influenced by the Indian summer monsoon was likely to be high (Thamban et al., 2002) and in turn the fresh water input into the study area. Conversely, sediments deposited after \(-7 \text{ kyr BP}\) show low Al and Na contents due to weak monsoon and associated low weathering of silicate rocks and thus low clay production. Given that sea water is a dominant source for Na and Cl to the sediments, the strong correlation between them (\(R^2=0.94\)) (Figure 4.1) indicate adsorption of sea salt with clay content of the sediments, which is relatively high in sediments deposited in late Glacial and early Holocene intervals.

Figure 4.4 shows that the lower content of Na and Cl correlate with heavier values of \(\delta^{18}O\) and higher percentage of Na and Cl correlate with lighter values of \(\delta^{18}O\). These patterns strongly suggest that \(\delta^{18}O\) values have recorded the regional monsoon fluctuations which enable me to infer the climatic variations from late Glacial to Holocene along with the textural studies and productivity variation of the cores studied (see section 4.3). Moreover, these are in well agreement with the climatic shifts in southeastern Arabian Sea as well as global climate changes. After compiling all these proxy studies, I could demark the following climatic events from the data set, which is highlighted in Figure 4.4.
Proxy records reveal (Figure 4.4) high intensity monsoon events at ~13.3, 11.9, 10.9, 9.2 and 7.4 kyr BP during the Late Glacial and early Holocene as supported by various proxies such as $\delta^{18}$O and other terrigenous elements (significant finding from this study) (Figure 4.4). These events match well with previous paleomonsoonal studies from the eastern Arabian Sea, suggesting that intense precipitation had occurred between ~13 and 6 kyr BP on Indian subcontinent, which is coinciding with the Northern Hemisphere summer insolation maxima (Prell, 1984; Van Campo, 1986; Sirocko et al., 1993). Several workers have reported the major climatic, hydrographic and circulation change in the Indian monsoon regime immediately after ~16 kyr BP (Sirocko et al., 1993; Naqvi and Fairbanks, 1996; Overpeck et al., 1996; Thamban et al., 2002). The $\delta^{13}$C record of peat from the Nilgiri Hills of the Western Ghats (South India) revealed that following dry LGM, moist conditions were started at ~16 kyr BP (Sukumar et al., 1993). The same abrupt, multi-decadal event was also inferred in the monsoon-induced upwelling record from the NW Arabian Sea with an abrupt change in G. bulloides percentage at ~9 kyr BP (Gupta et al., 2003). The C/N ratios of organic matter varied between 8.6 and 10.7 with relatively higher values (10.4–10.7) during the early Holocene suggest that
organic matter is predominantly marine (C/N 7–9) whereas some mixing of nitrogen-poor terrestrial organic matter (C/N >15) seems to have occurred during the early Holocene due to the enhanced lithogenic input (Thamban et al., 2007). After that, frequent high intense monsoon of the early Holocene precipitation event has occurred in a less dramatic manner starting at ~ 8.5 kyr BP (Figure 4.4). This event is synchronous with the first aridification stage of the Sahara (Jung et al., 2004). A sudden decrease in the Indus River discharge at ~ 8.4 kyr BP in the NE Arabian Sea has also been suggested to indicate the cessation of the early Holocene summer monsoon event (Staubwasser et al., 2002). It is evident that abrupt changes in summer monsoons were common within the broad Holocene precipitation, as supported by the high-resolution speleothem record from the northern Oman as well as the Oman marginal sediments (Neff et al., 2001; Gupta et al., 2003). The present study also reveals a significant weakening of the Indian monsoon starting at ~7 kyr BP synchronous with the starvation of the Indus/Makran Fan system (Prins and Postma, 2000). A combination of archaeological and other land records in the Indian subcontinent also supports a substantial weakening of the summer monsoon at ~7 kyr BP (Gupta et al., 2006). During the mid-Holocene period, one of the most significant weak monsoon periods is recorded in our study is at ~5.4 kyr BP. This event was well preserved in all terrigenous proxies records like Si, Al, Mg, and K (Figure 3.3). Such an event was also well demonstrated in the planktonic foraminiferal oxygen isotope and clay mineral records of sediment cores from the southwest coast of India which have been interpreted as a substantial decrease in precipitation related to Indian summer monsoon (Sarkar et al., 2000; Thamban et al., 2001; Thamban et al., 2002). The reduced precipitation at ~5.5 kyr BP is also noted in the Dongge Cave stalagmite record of southern China, one of the most well-resolved terrestrial monsoon records currently available (Wang et al., 2005). The above period also matches with the decreased biological productivity in the Oman margin (Gupta et al., 2005). After this dry phase a sharply increased high monsoon intensity between ~ 4.9 and 4.6 kyr BP was recorded in terrigenous proxy. After this event, proxy shows a long dry phase but $^{18}$O (G.rubber) record show reduced value at ~ 4.2 Kyr BP. This reduced value may be due to increased Indus river discharge this interpretation correlating with the Indus river increased discharge events at ~ 4 kyr BP, revealed from the varve sediments record (von Rad et al., 1999a). The terrigenous long dry phase record between ~ 4.5 -
2.1 kyr BP is matches with the records of Sarkar et al. 2000 suggests such a change in Indian summer monsoon precipitation during this period. The East Asian monsoon study from Retreat lake at Tiwan suggest that dry phase extended between ~ 4.5 - 2.1 kyr BP (Selvaraj et al., 2007). Late Holocene experienced a high monsoon event at ~2 kyr BP. Precipitation records from the Indian subcontinent did not show any significant increase in rainfall around this period but this event is recorded in an earlier study from the present study area (Thamban et al., 2001). After ~2 kyr BP event, there is weak monsoon event at ~ 1.3 kyr BP and this event correlate with the severe weakening of the summer monsoon activity around ~1.5 kyr BP that was evident in the Arabian Sea upwelling (Gupta et al., 2003) as well as historical records from India (Pandey et al., 2003). All these climate events which is recorded in the proxy records of core SK 215/4 are comparable with regional as well as high-resolution global climatic records. Such compatibility of present records suggests that the eastern Arabian Sea is climatically well connected globally.

A summary of the major, regionally consistent, rapid climate events thus identified in this study as well as in comparison with other records from the Arabian Sea is as follows:

1. The early Holocene intensification of summer monsoon occurred in two steps at 9.5 and 9.1 kyr BP. This is also clearly evident in the records of Gupta et al. (2003) and Ivanochko et al. (2005).

2. A significant decrease in monsoon intensity is noted at 8.5 – 8 kyr BP and this was strongly manifested in the dolomite (aridity proxy) record of Sirocko et al. (1993) as well as upwelling records of Gupta et al. (2003, 2005).

3. A mid-Holocene arid event around 6–5.5 kyr BP, wherein both precipitation and oceanic productivity remained low. This is well demonstrated in the records of Sarkar et al. (2000), Thamban et al. (2001, 2002), and Gupta et al. (2003, 2005).

4. During the late Holocene, monsoon conditions again deteriorated around 3.5 kyr BP. This is evident in the present study as well as that of Sarkar et al. (2000) and Ivanochko et al. (2005).

5. Around 1.6 kyr BP, a deterioration of the summer monsoon intensity is recorded, followed by a substantial enhancement around 1 kyr BP. These
events are also evident in the records of Sarkar et al. (2000), Gupta et al. (2003), and Ivanochko et al. (2005). However, there is a conspicuous mismatch in the records of freshwater input by the Indus River during the same period, possibly due to the fact that these records represent the winter/spring rainfall (Staubwasser et al., 2003).

The major monsoon events described by the present synthesis of marine records from the southeastern Arabian Sea basin appears to be synchronous with the Holocene “Rapid Climate Change” (RCC) events extracted from several globally distributed, multi-archive, multi-proxy records (Mayewski et al., 2004). Assuming that the dating uncertainties are not very significant, it is suggested that the differences in proxy responses could be an artifact of the different response time of proxies to the same forcing mechanism.

4.2.2 Sea surface temperature (SST) Reconstruction

The extent of SST variation in tropical regions during the glacial periods is still a matter of debate (e.g., Rind, 1990). The glacial-interglacial difference in δ¹⁸O recorded by planktonic foraminifers is a measure of the glacial difference in temperature of the surface seawater (Broecker, 1986). Based on planktonic and benthic oxygen isotopic results, Broecker (1986) concluded that tropical ocean temperatures remained within ± 2 °C of their present value. Manabe and Hahn (1977) suggested that the increased albedo over Asia, rather than a lower SST of the Indian Ocean, was responsible for a weak SW monsoon during the glacial periods. Quantification of tropical temperatures of the East African highlands based on the pollen studies (Bonnefille et al., 1990, 1992) indicate that continental temperature during the glacial periods were lower by 3-4°C than today. Recently, Thamban (2001) ruled out a deglacial warming of up to 3°C during the Last Glacial-Interglacial transition in the EAS and thus paleo-temperature reconstructions are in disagreement with one another and also with previous reconstructions based on foraminiferal transfer function (CLIMAP, 1981). Several other studies of the National Institute of Oceanography (NIO) also showed that SST has changed marginally (~2°C) or has remained the same in the Arabian Sea during the last 20 kyr BP (Duplessy, 1982:
The amplitude of SST change between the glacial and interglacial modes therefore remains to be determined.

Based on the alkenone studies, Rostek et al. (1994) reconstructed the paleo-SST of the southern Arabian Sea for the last 170 kyr. Comparison of alkenone-based paleo-SST data with δ¹⁸O record of western and southeastern Arabian Sea showed that the temperature and δ¹⁸O fluctuation are coherent. A higher SST during Marine Isotope Stage (MIS) 5 and lower SST during MIS 4 and 3 were documented: a maximum SST of 27.5°C during MIS 5.5 and 26 °C during 5.4 and 5.1 and broad SST minimum of ~ 23-24 °C was documented for MIS 4 and 3. Rostek et al. (1994) concluded that the SST difference between the Holocene and LGM was 2°C and similar results were also obtained from two alkenone-based SST records off Oman in the western Arabian Sea (Emeis et al., 1995).

As mentioned earlier, δ¹⁸O variation could be due to the regional changes in both SST and the evaporation–precipitation (E-P) balance. In order to estimate the variations in SST and salinity, it is necessary to separate the δ¹⁸O component resulting from changes in global ice volume SST and salinity from the measured δ¹⁸O values. The amplitude of variation in δ¹⁸O, which is a measure of the δ¹⁸O changes due to the component of global ice volume, SST and salinity variations. The δ¹⁸O component of global ice volume change was removed by subtracting 1.2‰ (Fairbanks, 1998) from Δδ¹⁸O. The residual amplitude of δ¹⁸O, after correcting for the global ice volume change, is an independent estimate of SST and/or salinity variations. However, if the residual amplitude of Δδ¹⁸O was zero, then the δ¹⁸O variation is only due to the global ice volume effect.

The amplitude of δ¹⁸O between the LGM and Holocene estimated from core SK-215/5, is -2.1‰ which is comparable with the value -1.9‰ estimated by earlier workers (Broecker, 1986, Sonzogini et al., 1998) as well as in recent study (-2.1‰; Thamban et al., 2001). The δ¹⁸O value that attributed to global ice volume change is -1.2‰. Therefore, these data suggest that the EAS did express variation in SST and/or salinity during the last ~67 kyr BP in addition to recording the global ice volume signal. Broecker (1986) show a decreasing δ¹⁸O by 0.2‰ for each degree of
temperature rise. Therefore, the residual δ¹⁸O value of −0.9 ‰ apparently suggest that the SST in the EAS was lower by 3–4°C during glacial period, but SST variation of up to 3–4°C is not valid for EAS, because several studies in this region showed that salinity also varied from LGM to Holocene (Duplessy 1982, Sarkar 2000); therefore, a residual δ¹⁸O value of only -0.4‰ (~2°C temperature variation) is valid for SST of southeastern Arabian Sea (Emieis et al., 1995; Rostek et al., 1997; Bord 1999).

### 4.2.3 Salinity Reconstruction

Variation of surface salinity in the tropical ocean is strongly related to E-P balance and hence, the latter is a sensitive climatic indicator. The E-P balance in an ocean basin leaves its signature on the δ¹⁸O signal of the surface sea water is, in turn, imprinted in the δ¹⁸O of planktonic foraminifers (Rostek et al., 1993).

After detecting -0.4 ‰ for SST variation (see earlier section), the remaining Δδ¹⁸O values are −0.5‰ for core SK-215/5. These values suggest an increase in salinity of up to 1-1.5‰ p.s.u (0.28‰ per p.s.u. salinity increase: Rostek et al., 1993). Therefore, the data suggest that fresh water input was reduced and evaporation enhanced during the glacial periods.

Duplessy (1982) showed that δ¹⁸O of G. ruber in the eastern Arabian Sea is lower by 0.5‰ compared to values from the same latitude in central and western Arabian Sea. These lower values are due to the low salinity waters. This could be because of the heavy rainfall over the west coast of India during the SW monsoon. Duplessy (1982) related the δ¹⁸O of surface waters with the salinity pattern in the Arabian Sea and demonstrated the strong evaporation prevalent in the basin. Therefore, the planktonic δ¹⁸O in this region must have varied in the past as a function of local precipitation also. The data obtained in this study show enrichment in δ¹⁸O during the last glacial period. This enrichment could be the result of stronger evaporation linked to strong, dry NE monsoon winds (Sarkar et al., 2000) and reduced fresh water input due to a weak SW monsoon (Duplessy, 1982: Van Campo et al., 1982; Sarkar et al., 1990) during the glacial periods. Sarkar et al. (2000) also interpreted the down core variations of planktonic δ¹⁸O in the eastern Arabian Sea sediments in terms of the E-P balance.
Late quaternary $\delta^{18}O$ studies of the Arabian Sea sediments indicated a low salinity during MIS 5 but high salinity during glacial period, especially during MIS 6 and 2 (Rostek et al., 1993). These data suggest that during MIS 5, the SW monsoon was strong. As the SW monsoon intensity was reduced and the dry NW winds dominate during glacial periods, there were changes in the E-P balance. Therefore, the weak SW monsoon resulted in a decreased fresh water input to the eastern Arabian Sea and hence, an increase in the salinity of this region.

There is a sharp north-south gradient in the summer monsoon rainfall along the west coast of India. It is reflected in the surface salinity of the eastern Arabian Sea (Duplessy, 1982: Sarkar et al., 2000). This difference may be explained in terms of lower salinity due to higher inputs of fresh water from the southwestern coast of India when compared to the northern region which receives much less rainfall. Thamban et al. (2001) suggested that the large $\delta^{18}O$ fluctuation at ~15.2 kyr BP is mainly due to the increased influx of the freshwater related to intensification of the summer monsoon and partly related to the deglacial warming of the sea surface. The present study recorded the high intensity monsoon related fresh water influence in the $\delta^{18}O$ data (migration of values towards lighter side) since ~13.5 Kyr BP (Figure 3.8). It is proved that after 15 kyr BP, the intensity of the southwest monsoon had increased which has reduced the salinity of the region since 15 kyr BP to early Holocene. Sarkar et al. (2000) also documented such a correlation between surface salinity (influenced by fresh water input) and $\delta^{18}O$ of planktonic foraminifers. Hence, the difference in residual $\Delta\delta^{18}O$ values of the two cores studied is a reflection of the salinity change due to the SW monsoon intensity.

4.3 Productivity of the southeastern Arabian Sea

In marine sediments, the organic carbon and CaCO$_3$ contents reflect the productivity, detrital contribution, or both. For example, the direct covariance between CaCO$_3$ and organic carbon can be attributed to productivity (Sheu and Presley 1986; Sheu and Huang 1989). The accumulation of biogenic calcium carbonate on the seafloor mainly depends on the surface water productivity, dissolution through the water column and dilution by the non-carbonate fraction such
as terrigenous matter. The carbonate-flux (carbonate mass-accumulation rate) on the seafloor has been extensively used as a reconnaissance for marine productivity. But its application is valid only if the preservation is complete. It has been shown by several workers that past climate induced variations in the preservation of carbonate are complicated approach. Detailed studies have been carried out in the Arabian Sea to understand the biological productivity during the late Quaternary (Siroko and Sarnthein, 1989; Shimmield et al., 1990; Clemens et al., 1991; Murray and Prell, 1992; Naidu, 1991; Bhusan et al., 2001, Thamban et al., 2001, Guptha et al., 2005).

4.3.1 Productivity record of Core Sk-215/4

The present studied deep-water core (SK-215/4) CaCO₃ and Corg record (Figure 3.2) shows a systematic trend in relation with glacial and interglacial stages, it shows that these proxy can be used to study climate-induced productivity of the study area. The increased CaCO₃, Corg, sand %, and reduced silt and clay % during the late Holocene, and subsequent decreased CaCO₃, Corg, sand %, and increased silt and clay % during glacial to early Holocene suggest increased productivity during late Holocene, and reduced productivity between last glacial to early Holocene. During MIS 3 increased CaCO₃, Corg, sand % and reduced silt and clay % suggest increased productivity. MIS 4 shows the decreasing trend of CaCO₃ and Corg and increased percentage of silt and clay which suggest reduced productivity when compare to MIS 3. Overall observation of all these proxy trend suggest increased productivity during the interglacial period and reduced productivity at glacial periods, this interpretation from this study is well in agreement with the interglacial high productivity record of Pattan et al., 2003 from southeastern Arabian Sea.

4.3.2 Productivity record of core sk-215/5

Down core profiles of CaCO₃, Corg and textural variation show (Figure 3.1) that the core has a well preserved record of productivity changes in the past. Down core profiles show reduced % of CaCO₃, Corg and reduced sand content and subsequent increased % of silt and clay between ~13.5 and ~ 8 kyr BP. This indicates reduced productivity during the late Glacial to early Holocene which is also recorded in the geochemical data. Depth profiles of geochemical data show that biogenic elements such as Ca, P, and Sr exhibit lower values between ~13.5 and ~ 7 kyr BP
and higher values are seen since ~7 kyr BP (Figure 3.6). These patterns of biogenic elements reveal that these elements appear to be diluted by the input of detrital sediments during the early Holocene when the monsoon was intense. All these proxies show distinct shift from low to high values between ~8.1 and 6.3 kyr BP. This shift indicates a major change in sedimentary delivery to the study site in the southeastern Arabian Sea due probably to a transition from strong to weak Indian summer monsoon during mid-Holocene. The data further shows that carbonate (CaCO₃) % are generally lower (Figure 3.2) and terrigenous elements (Si, Al, K and Ti) are higher (Figure 3.3) during the interval of intense Indian monsoon. A significant shift in carbonate and terrigenous elements occurs between ~7 and 6.3 kyr BP, correlating with strong to weak summer monsoon transition during the mid-Holocene in India and also supports the late Glacial to early Holocene reduced and productivity. During the mid- to late Holocene ~6.3 kyr BP to present the increased %, CaCO₃, Corg and sand % subsequent reduced % of silt and clay show an increased productivity during the entire late Holocene period. The interesting observation from this late Holocene CaCO₃ % and biogenic elements such as Ca, P and Sr is that reduced % of these proxies at ~2 kyr BP and 4.6 kyr BP, these periods are correlating with high intensity monsoon events in the Arabian Sea (see Figure 4.4 high intensity monsoon events). This reduced productivity is due to fresh water dilution effect to the core site. The recorded late glacial to early Holocene reduced and increased productivity records are in well agreement with the earlier studies from the Arabian Sea (Thamban et al., 2001; Guptha et al., 2005).

4.3.3 δ¹³C productivity record

In recent years, δ¹³C of panktonic foraminifers (mainly G. ruber and G.trilobus) was used as a proxy for the reconstruction of past pCO₂ (Wefer et al., 1999; Ganssen and Kroon, 2000). In general, the δ¹³C of CO₂ is inversely correlated with nutrient concentration (Ganssen and Kroon, 2000). Wefer et al. (1999) also suggested that the δ¹³C of carbonate-secreting organisms is a marker for the fertility of surface waters. The reason for δ¹³C enrichment in surface waters is that during photosynthesis ¹²C is preferentially used in the formation of organic matter. Thus, the export flux is enriched in this isotope and ¹³C tends to be left behind enriched surface

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water. This suggests that enrichment of $\delta^{13}$C is controlled by the concentration of nutrients in the surface waters.

### 4.3.3.1 $\delta^{13}$C productivity record of core SK-215/4

Down core profile of $\delta^{13}$C data shows (Figure 3.8a), though the data is not a high resolution one, an increasing trend from the bottom of the core and reaching maximum value at Marine Isotope Stage (MIS) 3 and a subsequent reduced value during Last glacial to early Holocene and the increased value at late Holocene suggests increased productivity during interglacial when compare to glacial period, which is well in agreement with carbonate (CaCO$_3$) productivity record of the same core.

### 4.3.3.2 $\delta^{13}$C productivity record of Core SK-215/5

The down core profile of $\delta^{13}$C data shows (Figure 3.8) late glacial and early Holocene reduced values and increasing trend from ~8 kyr BP suggest decreased productivity during late glacial to early Holocene and increased productivity during late Holocene which is correlates with the productivity proxy (CaCO$_3$, C$_{org}$) record of the same core. The interesting observation from the $\delta^{13}$C down core profile of this core is the reduced values at ~2 and ~4.6 kyr BP. This correlates with the reduced productivity record of CaCO$_3$, C$_{org}$ due to high intensity monsoon related fresh water input dilution effect.

### 4.4 Possible mechanisms

As we know the Earth’s climate is controlled by the sun through its solar radiation which was also proved by long-term climate variation i.e., glacial and interglacial climate changes (Milankovitch, 1939; Imbrie et al., 1989) short-term changes e.g., Holocene climate variation. The present study from the southeastern Arabian Sea has revealed long term climate change since MIS 4 and short term changes since Last glacial period. Earlier studies revealed that, in glacial period, after ~16 kyr BP Indian monsoon intensity has increased and has continued till early Holocene (Sirocko et al., 1993; Neff et al., 2001; Thamban et al., 2002). The terrigenous records of present study also revealed increased monsoon intensity since ~13.5 to 6 kyr BP. Earlier studies from the Arabian Sea have shown this increased
precipitation due to glacial boundary force (Sirocko et al., 1993; Overpeck et al., 1996). Within this increased monsoon in early Holocene, high intensity monsoon events recorded in this study indicate a rapid northward migration of the ITCZ. After ~8 kyr BP the gradual long term decrease in monsoon precipitation (as inferred low terrigenous input in this study) due to the continuous southward migration of ITCZ. On a millennial scale, the mechanism for switches in monsoon state is directly related to the relative heating of the Tibetan plateau and mid-latitude atmospheric moisture conditions (Higginson et al., 2004). The persistent oscillations observed within the various proxy records throughout the Holocene suggest that cyclicity is inherent to the low latitude climate systems. Although periodicities by themselves do not necessarily suggest a mechanistic link, they do indicate a possible relation between the monsoon and external forcing factor. The influence of solar activity variations in the Indian monsoon has already been explored, based on the presence of some periodicities and coherence at certain bandwidths with respect to the Δ¹⁴C data (Neff et al., 2001; Agnihotri et al., 2002; Staubwasser et al., 2002; Fleitmann et al., 2003). However, the exact mechanisms by which solar variability controls the climate system, (especially the low latitude monsoon systems are only recently being understood (Kodera, 2004; Gupta et al., 2005; Hameed and Lee, 2005; Wang et al., 2005).

Conventional climate modelling output suggests that small variations in solar activity may not lead to significant changes in major forcing factors. However, recent studies report that all these models overlook the various climate amplifiers that are available, especially in the low latitudes (Foukal, 2003). Typically, changes in the total solar irradiance were considered to be responsible for the decadal to centennial climate changes. However, it is extremely difficult to explain the observed regional differences in climate change based on radiative forcing alone. Using modern meteorological datasets and based on available model data, Kodera (2004) suggested that the solar influence on monsoon activity originates from the stratosphere through a modulation of the upwelling in the equatorial troposphere, which produces a north-south seesaw of convective activity in the Indian sector during summer. Increased solar activity would thus increase the convective activity in the equatorial region and would bring higher precipitation over Arabia and India (Kodera, 2004). Such a direct link between solar output and summer monsoon activity was also attributed to the observed close relation between reduced solar input and periods of weakened Indian
monsoons during the Holocene (Gupta et al., 2005). Therefore, the direct dynamic response of the troposphere, manifested as a seesaw of convective activity towards an external stimulation of solar activity, seems to be capable of explaining the direct solar modulation of the Holocene monsoon records as well as its regional disparities. It is suggested that throughout the Holocene, externally, small changes in solar activity controlled the Indian monsoon to large extent whereas internally, non-solar causes could have influenced the amplitude of changes.

4.5 Global teleconnection

Figure 4.5: Proxies records showing reduced southwest monsoon events in core SK-215/5, (a) G. ruber δ18O, (b) sedimentary Na content, (c) Cl concentration, and (d) Al values. Gray bars highlight the dry phase events.

Present study has revealed nine major intervals of summer monsoon minima over the past ~13.5 kyr BP (Figure 4.5), which are aligned within the radiocarbon age uncertainties to intervals of cold spells in the north Atlantic Ocean region (Bond et al., 2001). The well documented 8.2 kyr cold event (Alley et al., 1997) is also visible in our monsoon record. Our record shows that the summer monsoon, in general, was strongest in the early Holocene marked by high amplitude shifts between dry and wet phases and also summer monsoon shows a gradual weakening over the past 8 kyr with a more or less stable dry phase beginning ~5 kyr B.P that coincides with the onset of
an arid phase in India (Sharma et al., 2004) and termination of the Indus Valley civilization (Staubwasser et al., 2003; Gupta, 2004; Gupta 2005).

The present studied monsoon record from southeastern Arabian Sea is very similar to the Holocene changes in Arolik Lake, North Pacific in which increases in temperature and moisture corresponded to intervals of increased solar output, and vice versa (Hu et al., 2003). Previous studies correlated reduced solar irradiance to Holocene glacial advances in Scandinavia (Denton and Karlen, 1973), an early Holocene cooling event in lacustrine records from the Faroe Islands (Bjorck et al., 2001), drift ice record from the North Atlantic (Bond et al., 2001) and lake records from the North Pacific (Hu et al., 2003). Neff et al., 2001 found five episodes of reduced monsoon rainfall at the times of intense solar minima in the ~10 – 6 kyr BP records of cave deposits from Oman. Present study also recorded similar kind of dry phase within the early Holocene period (Figure 4.5). The low temperature cycles in the Sargasso Sea (Keigwin, 1996) increases in upwelling in the Cariaco Basin (Black et al., 1999), distinct episodes of drought in Yucatan Peninsula (Hodell et al., 2001) all have been linked to episodes of reduced solar irradiance, indicating that the footprints of solar impact on climate can be seen across tropics to poles.

The present works has significantly demonstrated the SW monsoon variability during the past ~13.5 kyr and its climate teleconnection with other part of the world