Environment and Climate Changes during the Late Quaternary: Inferences from Sedimentary Records of Southeastern Arabian Sea

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Shinu N

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Cochin University of Science and Technology
Cochin-682 016, India.

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DECLARATION

I hereby declare that the thesis entitled **"Environment and Climate Changes During the Late Quaternary: Inferences from Sedimentary Records of Southeastern Arabian Sea"** is an authentic record of research work carried out by me under the supervision and guidance of Prof. A.C. Narayana, Department of Marine Geology and Geophysics, School of Marine Sciences, Cochin University of Science and Technology, in the partial fulfillment of the requirement of the Ph.D degree in the Faculty of Marine Science and no part thereof has been presented for the award of any other degree in any University/Institute.

Cochin-682016
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Shinu N

CERTIFICATE

I certify that the thesis entitled, **"Environment and Climate changes during the Late Quaternary: Inferences from Sedimentary Records of Southeastern Arabian Sea"** is an authentic record of research work carried out by Ms. Shinu N. under my supervision and guidance at the Department of Marine Geology and Geophysics, Cochin University of Science and Technology, in the Faculty of Marine Sciences in partial fulfillment of the requirements for the degree of Doctor of Philosophy and no part thereof has been presented for the award of any degree in any University/Institute.

Cochin – 682 016
May, 2008

Dr. A. C. Narayana
Professor (on lien)
Dept. of Marine Geology and Geophysics
Cochin University of Science and Technology
Cochin-682 016

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Shinu N

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CHAPTER-I

Introduction

1.1. General introduction

The Indian monsoon is an important feature of the tropical climate, marked by seasonal reversal in the wind system – the summer or south west (SW) and winter or northeast (NE) monsoons. The intense upwelling during summer monsoon causes high surface production as well as organic carbon flux to the seafloor, where as during winter monsoon the upwelling activity is low, resulting in a low flux of organic carbon (Krey, 1973; Schulz et al., 1998). Thus the Indian monsoon is critical for understanding past regional and global climate variability as well as hydrological and carbon budget of the tropics (Goodbred and Kuehl, 2000; Fleitmann et al., 2003; Gupta et al., 2003; Sharma et al., 2004; Wang et al., 2005). Sensible or direct heating of the Asian landmass and the latent heat released from precipitation drive the SW monsoon circulation (Webster, 1987). The Indian monsoon has a profound impact on the economy of South Asian region and is crucial to the well-being of large population of Asia. Hence long time series of climate and weather parameters are crucial to developing predictive climate/ weather models.

The nature and timing of these climatic changes and their relationship with the intensity of monsoonal precipitation are the foci of paleoclimatological research the world over. The Indian monsoon precipitations affect the lives of two-thirds of the Indian population that directly or indirectly depend on agriculture and related industries. The knowledge of past climate variability is of great relevance for understanding the dynamics of the climate system. A comprehensive knowledge of the monsoon behaviour during different climate periods can provide deeper insights into possible future circulation patterns. Therefore we need to be able to reconstruct past climate variability using multi-proxy data to evaluate the reliability of predicted climate scenarios.

The Quaternary was a period of major environmental changes and has been the most eventful among all other geologic periods. The Late Quaternary period (<130 ka BP) has witnessed large changes in global climate. Eustatic sea level fluctuated significantly in accordance with the waxing and waning of continental ice sheets. There is a necessity for an understanding of climatic variation and changes during the Quaternary period, in order to comprehend fully our present climate. A paleoclimatic record, spanning the Late Quaternary period is, therefore, fundamental to comprehension of modern climate and the causes of climatic variations and changes (Kutzbach, 1978). Unless the natural variability of climate is understood, it will be extremely difficult to identify with confidence any anthropogenic effects on climate.

Marine sediments have the potential to provide high resolution records of paleoclimatic and paleoceanographic change. Such records are essential for knowing about the climatic variability on a range of time scales, for the Holocene as well as over glacial-interglacial intervals. The recovery of such core sediments of Holocene period has provided an opportunity to construct very high resolution records of climatic and oceanographic variability during the transition between glacial and modern conditions, and to study the relationship between water column productions, sediment supply and bottom water ventilation in a coastal basin. Bottom water oxygen levels in coastal basins respond not only to the rate of renewal of the deep water, but also to the settling flux of organic carbon through the water column.

The continental shelf off southwestern India (Mangalore – Cochin) is 60 km wide and shelf break occurs at ~60 m water depth. The inner shelf is smooth / even with gently sloping topography where as middle shelf shows uneven topography with small topographic highs. The study of marine sedimentary records has contributed enormously for the understanding of the Palaeoceanography and Quaternary climate. The climatic and sea level changes recorded in the sediments during the Late Quaternary can be distinguished using several proxies which bear clear signals of the past climate and oceanography. Biological, bio-geochemical, geochemical and sedimentological proxies are the most commonly employed ones for this purpose. The oceanic sedimentary records provide a better understanding on the past variations, as they are derived from the inputs as well as from coastal processes such as upwelling and biological productivity. Moreover oceanic regimes are sites of burial for sediments and organic matter derived from continental and marine sources.

1.2. Reconstruction of paleoclimate

The oceans cover nearly 70 % of the Earth's surface, which drive and regulate its total climate system. The climate is having very complex interactions with atmosphere, hydrosphere, cryosphere, biosphere and lithosphere. Such large scale interactions may produce distinct changes in climate. Hutton proposed principles of uniformitarianism considering *'present as key to the past'*; whereas the paleoclimatologists endeavour to *'predict the future climate by looking in to the past variations'*. Only when the causes of the past climate fluctuations are understood, then only it will be possible to anticipate or forecast climatic variations in the future (Bradley and Eddy, 1991). Based on present knowledge our past Earth has undergone several alternating warm and cold periods particularly in the Pleistocene with amazing rhythm. Therefore the Pleistocene period is known as the Great Ice Age. Since the ocean has a major role in generating such climatic variations, its response to the Earth's climate change although very complex has been distinct and measurable. Therefore, oceans provide most important and easily accessible repository for tracing the paleoclimate records. Almost all changes in oceanic environment in response to climatic variation could be traced within the seafloor itself. The seafloor sediments record the changes in water column chemistry, biological activity, air-sea interaction; inter oceanic overturning, land-ocean interaction, deep water circulation etc.

Earth's radiation budget largely controls the climate. A steady state climate unchanged over a period should ideally represent well-balanced radiation budget of the Earth i.e., incoming solar radiation equals the out going solar radiations. This balance has fluctuated several times in the past resulting in cooling and warming of the climate, which are commonly known as glacial and inter glacial periods respectively. Earth has experienced such dramatic climatic conditions during the last 2.7 million years. With in the late Quaternary itself there are 10 major cold (glacial) events separated by warm (interglacial) events known as climatic cycles. These climatic cycles are further punctuated by shorter time span, moderately cold and warm events called stadial and interstadial respectively. The main effect these glacial-interglacial climate cycles was extensive waxing and waning of the continental ice sheets resulting in fall and rise of the global sea level, which have further modified the continental and oceanic climate set up. For instance, during the last glacial maximum (LGM) the global sea level was lowered by about 120 m as a result of global oceans loosing large amount of fresh water, which was locked on the continents in the form of ice (Fairbank, 1982; Shackleton and Opdyke, 1973). This type of climate changes occur on global scale since the time Earth has come into existence as a part of its natural dynamics. The information related to those changes has been stored in different forms (proxies) on continents and oceans. The time –scale of changes, which one would resolve depends upon the process involved in the genesis of particular proxy and the location of its formation. The shallow water corals may provide the information on yearly time-scale, but the deep abyssal oceanic sediment on thousands of years time-scale. The resolution also depends up on the sampling and measurement techniques. It is possible to isolate anthropogenic or local effects from a composite record, if we subtract the natural effect with reasonable confidence. Therefore, it is important to learn about the variability in the past climate to understand or predict the possible future changes. This highly intriguing, fascinating, and complex branch of earth sciences is termed as 'Paleoclimatology'.

It is essential to study long term, high resolution marine or continental paleoclimate records to gain a better understanding of the present monsoonal climate or to predict future climate changes. As the Instrumental records are available only for the last two centuries, estimates of global climate variability during the past can, therefore, be drawn as indirect inferences from the "proxy" indicators.

A variety of marine and multi proxy records have been used in recent years to expand our understanding of fluctuations in the intensity of the Indian summer monsoon, climate and environment during the past. Numerous proxies from marine and continental records have been used to study the changes of monsoon variability and to understand the forcing mechanisms of the Asian summer monsoon covering a time span of thousands of years (Clemens et al., 1991; Prell et al., 1992; Sirocko et al., 1993; Naidu and Malmgren, 1996; Schulz et al., 1998; Gupta et al., 2003). Proxy indicators by virtue of their physical, biological and chemical origins are indirectly related to paleo oceanographic and paleo climatic variables of interest. Therefore no single proxy has a unique and direct association with monsoon strength. All proxies have the potential to be influenced by processes other than changes in the monsoon or climatic and/ or oceanographic variables of direct interest (Clemens and Prell, 2003).

Paleomonsoon reconstructions over South Asia since the Last Glacial Maximum (LGM) have relied on sediment cores from the western Arabian Sea (Sirocko et al. 1996; Overpeck et al., 1996; Zonneveld et al., 1997; Gupta et al., 2003; Jung et al., 2004). Several studies on the past variability of the Indian monsoon have made use of marine sediments of eastern Arabian Sea (Sarkar et al., 2000; Anderson et al., 2002; Prabhu et al., 2004; Prabhu and Shankar, 2005). Areas with high sedimentation rates (that can provide high resolution data) are now the focus of major coring efforts (IMAGES Planning Committee, 1994). Sediments from such areas can document changes on the ~ 10^2 years timescale or even at the decadal scale (e.g., Hughen et al., 1996).

Evidence for the climate history of the Asian summer monsoon have been recorded for the Holocene period based on various geological archives such as marine sediments (Sirocko et al., 1996; Overpeck et al., 1996; von Rad et al., 1999b) lake sediments (Wei and Gasse, 1999; Zhu et al., 2003) and pollen and isotope records from peat bogs (Frenzel, 1994; Schulz, 1999; Tang et al., 2000; Hong et al., 2000) or lakes (Gasse et al., 1991; Jarvis, 1993; Van Campo et al., 1996). A few studies of the Arabian Sea sediments have also provided deep insights into orbitally induced monsoon variability (Clemens et al., 1991; Anderson and Prell, 1993; Naidu and Malmgren, 1996; Reichart et al., 1998).

1.3. Scientific rationale

Arabian Sea, a semi-enclosed basin, is surrounded by arid landmasses to the west and north and by coastal highlands of western India to the east. A large amount of fresh water input through Indus River brings in a large amount of terrigenous sediments during the SW monsoon period and major influx of dust plumes from the Arabian Peninsula. The Arabian Sea experiences extremes in both atmospheric forcing and oceanic circulation that lead to the seasonal variability in surface water properties (Lal, 1994). High biological productivity and subsequent decay of organic matter lead to extremely low dissolved oxygen concentration and related intense water column denitrification along the continental margin of the Arabian Sea (Naqvi, 1987). The proximity to the Sahyadri hill ranges in the east and high rainfall in the hinterland make the biology and chemistry of the Arabian Sea quite susceptible to continental influences both through fluvial and atmospheric inputs of trace metals, nutrients and other elements that can affect ocean productivity. In view of its unique oceanographic and physiographic set ups, the Arabian Sea is considered as a natural laboratory to study past climate and possible future climatic changes (Smith et al., 1991).

1.4. Objectives of the present study

The present study envisages the following objectives:

- To evaluate the sedimentation rates in the southeastern Arabian Sea.
- To assess the type and distribution patterns of clay minerals.
- To understand the climatic changes during Late Quaternary based on mineralogy and geochemistry of marine sediments.
- To infer the paleoenvironmental conditions in the southeastern Arabian Sea.

1.5. Previous studies

The Indian monsoon system is one of the major atmospheric components of the tropical climate and is a complex system. Previous studies have suggested that the strength of the summer monsoon, upwelling and productivity in the Arabian Sea are related together. Majority investigations were mainly concentrated in the western Arabian Sea, especially in the Oman margin. The eastern Arabian Sea has been focused since last decade, as it is an intriguing region with respect to its response to the past climate driven monsoon variations. However, it is necessary to explore this region in more detail in order to understand the effects of past climate on relative strength of the monsoons and associated responses such as sedimentation history, productivity, fluvial input of detritus etc. whether these parameters in the eastern Arabian Sea responded in concert with varying monsoon regime or were they showing difference for a given climate scenario.

Sediment accumulation rates have been estimated by dating marker sediment layers in sediment cores and various dating techniques have been employed. Along the western continental margin of India, the dating and accumulation rates have been carried out by employing Pb-210 dating technique, and either conventional C-14 dating technique or accelerator mass spectrometry C-14 dating. Pb-210 dating techniques give the modern sedimentation rates i.e., for last 100 years, whereas C-14 dating techniques provide the accumulation rate for the late Quaternary period. C-14 dating technique provides the information with regard to long term time frame paleoenvironmental , paleomonsoonal and paleoproductivity scenario.

Studies on sediment accumulation rates on the southwestern continental margin of India are limited and clustered at a few places and most of the studies were of shallow regions (Nigam and Nair, 1989; Karbassi, 1989; Manjunatha and Shankar, 1992; Nambiar et al., 1991; Caratani et al., 1994; Nambiar and Rajagopalan, 1995; Pandarinath et al., 1998, 2001, 2004). Information on sedimentation rates on the continental slope is limited (Shankar and Manjunatha, 1995; Somayajulu et al., 1999; Thamban et al., 2001; Agnihotri et al., 2003; Pandarinath et al., 2004, 2007). Earlier studies by various workers along the eastern Arabian Sea have presented contrast results with regard to sedimentation accumulation rates. Sedimentation rates on the continental slope off Saurashtra show a relatively small variation (0.77 -1.03 mm/yr) over the last 1300 yr at 280 m water depth. But at deeper water levels (480 m), the sedimentation rates widely vary from 0.05 to 1.34 mm/yr during the Holocene (Somayajulu et al., 1999). The sedimentation rate on the upper continental slope region off Cochin was higher (0.52 mm/yr) during 15-16 ka BP (Thamban et al., 2001). The sedimentation rate was relatively low (0.16-0.19 mm/yr) during the past 19000 years off Taingapatnam (Calicut) (Nambiar and Rajagopalan, 1995). Futher low sedimentation rates (0.01-0.04 mm/yr) during the Holocene were reported by Agnihotri et al. (2003).

Arabain Sea has been extensively investigated for organic carbon production and preservation. High content of organic carbon in surface sediments in eastern Arabian Sea, especially western continental margin of India suggest high biological production and better preservation (Calvert et al., 1996; Paropakari et al., 1992; Slater and Kroopnick, 1984; Rao and Veerayya, 2000). The primary cause of carbon enrichment in sediment has been strongly debated (Paropakari et al., 1987, 1992 and references there in). The preexisting deep water anoxic condition (Calvert et al., 1993; Demaison and Moore, 1980; Paropakari et al., 1992), high productivity and favourable sediment texture (Calvert et al., 1995; Pedersen and Calvert, 1990; Pedersen et al., 1992; Rao and Veerayya, 2000) have been invoked as the causes for good preservation of the organic matter. Interestingly, there have been contrasting views with past productivity in the Arabian Sea, and it is worth mentioning that past productivity changes in the basin have exhibited region specific (western vs eastern) responses to the climate change. On the one hand the western Arabian Sea has been shown to record interglacial high productivity (Emeis et al., 1995; Naidu and

Malmgren, 1996; Shimmield et al., 1990; Spaulding and Oba, 1992), while eastern Arabian Sea has indicated enhanced glacial productivity (Cayre and Bard, 1999; Rostek et al., 1997; Schulte et al., 1999; Thamban et al., 2001). Duplessy (1982) has observed that stronger winder monsoons during the LGM than the Holocene and is attributed to the glacial high productivity.

Earlier studies have mostly concentrated on the northwestern Arabian Sea (Clemens et al., 1991; Sirocko et al., 1993; von Rad et al., 1999a,b; Naidu and Malmgren, 1996; Naqvi and Fairbanks, 1996; Overpeck et al., 1996; Zonneveld et al., 1997; Reichart et al., 1998; Schulz et al., 1998). Since well-studied records are few compared to the other parts of the Arabian Sea (Indian margin), sediments from northwestern Arabian Sea region are particularly suitable to monsoon related paleoenvironment, paleoclimate and paleoproductivity changes since the terrigenous supply in this region is low, where the variations follow the paleomonsoon patterns. Large volume of data exists on the study of paleoclimate from the Arabian Sea and in particular from the western part of Arabian Sea (Naidu, 2004; Naidu and Niitsuma, 2003; Sirocko et al., 2000; Gupta et al., 2003) and north western Arabian Sea (Reichert et al., 1997, 2002; von Rad, 1999 a,b). These studies concentrate largely on the variability of upwelling strength and its associated biological productivity induced by summer monsoon. Most of the researchers have employed organic carbon, calcium carbonate, stable isotopes, clay minerals or sedimentological analysis and elemental geochemistry of sediments as proxies to understand and reconstruct the upwelling intensity in the Arabian Sea. The studies have shown that the upwelling and its associated biological productivity reflect the intensity of southwest monsoons. It is also found that the productivity is greater during Holocene than the last glacial times. Studies from the western and northwestern Arabian Sea shows organic carbon, conventional proxy for biogenic productivity (Ganeshram et al., 1999; Rixen et al., 2000) closely relates the surface productivity in accordance with the strength of summer monsoon unlike eastern Arabian Sea where its distribution is affected by sedimentation (Agnihotri et al., 2003) and bottom water anoxia (Sarkar et al., 1993).

Paropakari et al. (1987, 1992, 1993), based on organic carbon data from the recent sediments of the Eastern Arabian Sea, invokes an argument that the organic carbon does not reflect the surface productivity. It is further argued that the concentration of organic carbon in sediments principally depends on the intensity of preservation due to strong oxygen minimum condition (OMZ), which lies on the continental slope of western continental margin of India (Paropakari et al., 1987, 1992, 1993; Calvert et al., 1995). Several other factors are also responsible for organic carbon preservation. Prakash Babu et al. (1999) found that the organic carbon distribution is very well correlated with the water column productivity. Pattan et al. (2003) have reported higher concentration of organic carbon in southeastern Arabian Sea during stadials. These observations suggest that the organic carbon distribution is influenced by various oceanographic, geological processes rather than surface productivity alone.

Calcium carbonate is also being used widely as proxy for biological productivity (Nair et al., 1989; and various other researchers). Gehlen et al. (2005) has done the reassessment of the dissolution of marine carbonates and concluded that the evolution of the bulk composition of the carbonate fraction is not paralleled by a significant change in its stochiometric concentration product and it reflects ongoing differential dissolution due to kinematic effects.

Naidu and Malmgren (1999) studied the Quaternary carbonate records from the equatorial Indian Ocean and its relationship with productivity changes using stable isotopes, Ba/Al ratios, percentages and fluxes of $CaCO_3$ in the total sediment and suggests that the percentages and fluxes of total $CaCO_3$ do not exhibit any apparent relationship to the glacial/interglacial cycles and fluctuations do not correspond to either an Atlantic or a Pacific pattern. CaCO3 fluctuations mainly reflect productivity changes of $CaCO_3$ secreting organism in the water column.

The existence of interglacial/ glacial $CaCO_3$ cycles in the equatorial Pacific was first observed during the Swedish Deep Sea Expedition (Arrhenius, 1952). Many studies have been devoted to unraveling $CaCO_3$ cyclicity in the Pacific and Atlantic oceans. Arrhenius (1952, 1988), Olausson (1971, 1985), Emerson and Bender (1981) and Archer (1991) argued that productivity has the most important influence on the CaCO3 content in deep sea sediments, where as Olausson (1965), Gardner (1975), Berger (1973, 1992) were of the opinion that it is mainly controlled by dissolution.

The variations due to terrigenous dilution and/or dissolution of $CaCO_3$ have been thoroughly debated by various researchers (Ruddimann, 1971; Damuth, 1975; Gardner, 1975), Pacific (Arrhenius, 1952; Farrell and Prell, 1989; Berger, 1992) and

Indian Ocean (Olausson, 1965; Naidu, 1991; Naidu et al., 1993; Naidu and Malmgren, 1999) and they suggest that dissolution and dilution are the most important factors for greater amplitude fluctuations in $CaCO_3$ records, whereas productivity might only be ranked third in importance.

Previous studies show concrete evidence that dissolution is stronger during inter glacials in the Pacific and during glacials in the Atlantic (Volat et al., 1980). This dichotomous pattern of $CaCO_3$ preservation between Atlantic and Pacific oceans has spurred the interest of many paleoceanographers. This basin to basin fractionation model has been suggested to explain this difference in $CaCO_3$ preservation patterns between Atlantic and Pacific Oceans (Berger, 1973). This model involves changes in the circulation patterns of the deep oceans, and resulting redistribution or fractionation of CO_2, $CaCO_3$, and nutrients, to explain the out-of-phase dissolution patterns between the Atlantic and Pacific oceans (Olausson, 1971, 1985; Berger, 1973; Luz and Shackleton, 1975; Crowley, 1984).

The Quaternary $CaCO_3$ patterns in the Indian Ocean have not been analyzed in such detail as in the Pacific and Atlantic Oceans. In the Indian Ocean some sites exhibit a Pacific pattern (Olausson, 1965; Oba, 1969; Naidu, 1991), whereas others show both Pacific and Atlantic patterns (Peterson and Prell, 1985; Naidu et al., 1993). In the equatorial Indian Ocean, cores marked by greater dissolution, like IC-5 (Oba, 1969), V34-53 (Peterson and Prell, 1985), and Swedish Deep Sea Expedition core 154 (Volat et al., 1980), generally exhibit a Pacific type of dissolution pattern in major portion of the records. The composite coarse fraction index (CCFI) also reveals that greater values during glacial episodes and lower values during interglacial episodes (Bassinot et al., 1994), which also confirms greater dissolution during interglacials (Murray and Prell, 1992; Naidu et al., 1993).

Although the organic and inorganic constituents of marine sediment are the most widely used proxies in climate research, the terrigenous fractions of the sediments are very useful especially in reconstructing the terrestrial environments (Chamley, 1989; Chauhan, 1999: Thamban et al., 2002). Clay minerals have been used as a paleoclimate proxy by various researchers (e.g. Biscaye, 1965; Chamley, 1989; Singer, 1984; Rao and Rao, 1995; Thamban et al., 2002, Kessarkar et al., 2003 etc). Biscaye (1965) argue that clay minerals in marine environment are found to be largely detrital and can be successfully used as paleoenvironment indicator. Chamley (1989) suggests that clay minerals of marine sediments reflect the prevailing climatic conditions, hydrography, geology and topography of the continental source area and its composition basically indicates the intensity of weathering especially the degree of hydrolysis at source region and can be used as paleoclimatic proxy.

Regional studies on clay minerals along the western margin of India are quite a few (Stewart et al., 1965; Mattait et al., 1973; Nair et al, 1982 a, b; Rao, 1991; Rao and Rao, 1995; Chauhan and Gujar, 1996; Thamban et al., 2002). Mattait et al. (1973) reported the clay mineralogy of samples from some widely spaced traverses of the continental shelf and Nair et al., (1982b) made a detailed study of the clayey sediments of the inner shelf. Subramanian (1980) and Naidu et al. (1985) reported the clay mineralogy of fluvial and bed load sediments from the west coast of India and Ramaswamy et al. (1991) the mineralogy of a particulate flux accumulated in traps moored at different locations of the Arabian Sea. Regional studies of the clay mineralogy of marine sediments from the northern Indian Ocean (Griffin et al., 1968; Rateev et al., 1969; Goldberg and Griffin, 1970; Kolla and Biscaye, 1977; Kolla et al., 1976, 1981) have shown the influence of the major rivers of the southwest coast of India on deep sea sedimentation. Clay mineral studies on surficial sediments in the Arabian Sea have so far been used mainly to charecterise the provenance and transport pathways of fine grained sediments (Kolla et al., 1976, 1981; Nair et al., 1982 a,b; Chauhan, 1994; Rao and Rao, 1995, Gingele, 1996; Thamban et al., 2002). Besides there are other few studies on sediment cores from the western Arabian Sea, reporting the past variations in clay mineral distribution in relation to changes in atmospheric circulation and continental climate during the Late Quaternary (Sirocko et al., 1991; Sirocko and Lange, 1991). Only very few studies were conducted on the eastern Arabian sea, to deduce the paleoclimatic conditions using clay mineral proxies.

Previous studies of clay mineral composition of marine sediments from the continental margin of southeastern Arabian Sea were mainly based on surficial sediments and were used to understand the provenance and transport pathways of fine grained terrigenous sediments (e.g., Biscaye, 1965; Kolla et al., 1981; Nair et al., 1982 a,b, Chauhan, 1994; Rao and Rao, 1995; Rao and Wagle, 1997). Only very few studies were conducted on the eastern Arabian Sea continental margins to infer the paleoclimatic conditions using clay mineral proxies.

The type and amount of terrigenous material in the marine environment depend on climatic conditions on the continent. About 95% of terrigenous material in the ocean reaches from the rivers and is deposited on continental margins. The elements and their transport pathways in the Arabian Sea suggest a number of sources, viz. detrital input from Somalia, aeolian from Arabia, detrital riverine input from Indus, Tapti and Narmada rivers and weathering of Deccan trap, gneissic rocks, laterites and submarine weathering of Carlsberg Ridge. The lithogenic input in the western Arabian Sea in largely eolian and is fluvial in the southeastern Arabian Sea with a meagre aeolian fraction. Existing palaeoceanographic studies in the southeastern Arabian Sea are confined to productivity variation, clay mineralogy and hydrography changes and variations of terrigenous input through geochemistry is not well recorded (Pattan et al., 2005).

The recent development of a suite of geochemical (Calvert and Pedersen, 1993; Piper and Isaacs, 1995; Rosenthal et al., 1995; Calvert and Fontugne, 2001) and isotopic (Calvert et al., 1992; Altabet and Francois, 1994; Ganeshram et al., 1995; Farrell et al., 1995) indicators or tracers of paleooxygen and/or paleoproductivity changes. High resolution continuous and quantitative chemical analyses of sedimentary records provide useful information on paleoclimatic changes (Peterson et al., 2000). Biogenic and terrigenous material can be estimated from major element composition and examination of temporal or spatial variations in the contents of these components is useful for reconstructing paleoclimatic and paleooceanographic changes (Sirocko et al., 1991; Tada, 1991; Irino and Tada, 2000, 2002). Several recent studies of marine sediments from the Arabian Sea have demonstrated their utility for reconstructing abrupt climate change during the last glacial period. In particular, indicators of marine paleoproductivity, sea surface temperature (SST) and dust flux, record intensity fluctuations in the dominant monsoon climate with similar regularity to climatic changes over Greenland (Reichart et al., 1997, 1998, 2002; Schulz et al., 1998; Sirocko et al., 1993, 1996; Schulte et al., 1999, 2000). Sediment trap studies have demonstrated that productivity and SST variations in the northern Arabian Sea can be attributed to a combination of SW monsoon-induced upwelling and convective mixing associated with winter cooling (Nair et al., 1989; Madhupratap et al., 1996).

Al and Ti content in the marine sediment mainly derive from continental origin supplied through eolian and fluvial pathways. Therefore, Al and Ti have been commonly used as geochemical indicators to discover the content of aluminosilicate detritus of continental origin (Govil et al., 2004). Terrigenous sedimentation in the marine environment is mainly through fluvial or eolian pathways, which provides information about conditions on the continent and mechanisms of transport from continent to marine environment.

In the present study, sedimentation rates, organic carbon, calcium carbonate, clay minerals, major and trace elements are employed to understand paleooceanographic, paleoclimate and paleoenvironment during the Late Quaternary in the southeastern Arabian Sea. The data generated in this study can be utilized as an archive, and adds to the existing knowledge of the southeastern Arabian Sea.

1.6. Study area

The Arabian Sea is a unique basin composed of complex sea floor, seasonally changing hydrography, and isolation from the Arctic. It covers an area of about 3,863,000 km^2 and is located between 7 and 25°N latitudes and 55°E and 75°E longitudes forming the northwest water body of the Indian Ocean. The Arabian Sea is surrounded by arid land masses to the west and north and by coastal high lands of western India to the east.

The southeastern Arabian Sea comprises a dynamic water body receiving intense summer monsoon overhead precipitation with well defined precipitation gradient from ~4000 mm in southern Konkan coast to ~300 mm in northern Saurashtra coast, decreasing northward at rate of ~350 mm per degree latitude (Sarkar et al., 2000). This region supports moderately high productivity due to seasonal upwelling (Haake et al., 1993) and strong oxygen minimum zone (OMZ) due to which high export production is apparent (Rao and Veerayya, 2000 and references therein). In contrast to the large input of of the dust due to the western Arabian Sea from Arabia (Sirocko et al., 1991; Shimmield et al., 1990). The south eastern Arabian Sea receives very low dust input and is dominated by the terrigenous load delivered by the large network of rivers quickly draining the Deccan traps during the summer monsoon rains and to certain extent the Indus river input derived from the Himalayas particularly in the abyssal depths.

The surface salinity structure of the Arabian Sea is peculiar. The north to south decreasing, west to east trending isohalines of the basin are punctuated in the eastern region by the inflow of low salinity Bay of Bengal (BOB) water, which forms a low salinity tongue along the western margin of India. The biogeochemical responses of the Arabian Sea in general and southeastern region in particular to the past climate and their past linkage with monsoon variations have been traced in the sediment. Therefore, the sedimentary records from the southeastern Arabian Sea are well suited repository to explore the fluctuations in closely interlinked past monsoon strength, salinity structure, productivity and fluvial erosion in Deccan trap region. As the southeastern Arabian Sea is the least studied region of the Arabian Sea in regard to paleoclimate reconstruction, may be holding interesting information with respect to the biogeochemical responses associated with the climate change.

1.7. Physiography

The geomorphology of the hinterland is heterogenous and marked by the presence of numerous small and medium size rivers viz., Bharatapuzha, Chalakudy, Periyar, Muvattupuzha, Pamba and Achankovil flowing from the Western Ghats. Annually these rivers discharge suspended sediment load of sbout $11.7\text{x } 10^5$ metric tons. The width of the coastal plain generally ranges between 5 and 10 km but attains a maximum width of about 30 km at some places. The coast is charecterised by various land forms such as lagoons, barrier islands, beach ridges, paleo-strand lines, alluvial plains, marshy lands and flood plains. The shore line is generally straight trending NNW–SSE with minor variations and lies as a narrow and low lying land. A series of sand dunes/strand lines oriented parallel to the general direction of the coast line hinders the flow of the rivers, there by trapping sediments and enlarging the alluvial plains. Physiographically, the southwestern part of India is very diverse. From the coastal plains in the west, the landscape changes steadily through mid lands to high lands in the east. Coastal alluviums though prominent in central Kerala, are nearly absent in the northern and southern parts (Narayana & Priju, 2006). Midlands are characterized by lateritic plateaus that extend upto the coast. Four physiographic divisions have been identified: (i) the 600-1800 m highland terrain of the rocky mountain (ii) the 300-600 m midland characterized by thick lateritic cover, (iii) the <10 m high coastal plains of raised beaches and sand dunes (Karunakaran and Sinha-Roy, 1981). Sahyadri (Western Ghats) has a significant influence on the intensity and distribution of rainfall over peninsular India. The Sahyadris, a major physiographic feature covering an area of about 54,000 km^2 along the west coast of India, is composed of Precambrian granulites, gneisses and schists in southern part. Recent alluvium underlain by Mio-Pliocene Warkalli formation (ferrugenised sandstones with clay intercalations) is present in Kerala at places along the coast. Geological formations in the hinterland are largely gneissic rocks, which are in some places covered by thick laterites (Krishnan, 1968). Rivers of the southwest coast of India are seasonal and carry maximum sediment load during the Southwest monsoon (June-September). During south west monsoon period, intense coastal upwelling occurs south of 15^{o}N. Coastal currents around India reverse seasonally. During the southwest monsoon, the largest transport of west India coastal current is observed off the southwest coast of India (Fig.1.1).

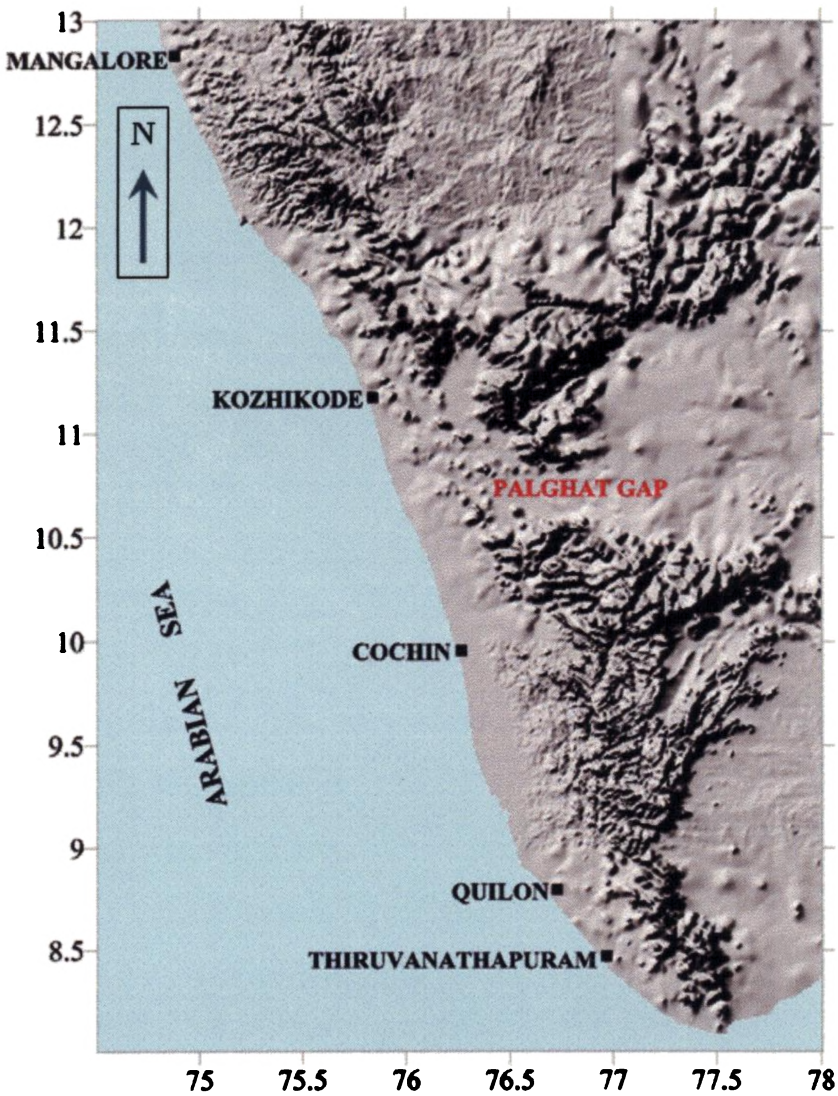

Figure. 1.1. Physiography map of the southwest coast of India.

1.8. Geology of hinterland

The Deccan trap basalts are the dominant rocks in the north western coasts of India. Precambrian granulites, granites, schists and charnockites are the predominant rock types in the region. The khondalite belt with intercalations of minor charnockite and leptynite and calc silicate rocks occurs in southern Kerala. The central part comprises dominant charnockites with minor granite gneisses and meta-sedimentary rocks. Granulites, schists and gneisses, intruded by acid and alkaline plutons, are present in northern Karala. The oldest rocks are charnockites, which also are the most wide spread rock unit in Kerala. In the coastal region, these rocks are extensively lateritised at places. The coastal laterites are developed up on the Deccan lavas in the north western coast and continue extensively southward upon Archean and Proterozoic lithologies as well as up on rare out crops of Mio-Pliocene sediments (Widdowson and Gunnel, 1999). The Miocene Warkalli and Quilon formations (ferrugenous sand stones with clay intercalations) are exposed at isolated locations mainly at southern tip of India. Recent soil alluvium occurs through out the coastal tracts of south west India. Kaolin clay deposits of residual and sedimentary types occur as 10-20 m thick layers at several along the south west coast of India and the resources are estimated at 127.48 million tons (Indian Minerals Year Book, 2001). Laterites are inter-bedded with layers of clay, which is of kaolinitic in felsic rocks, and lithomarge in iron-rich rocks (Soman, 2002) (Fig.1.2).

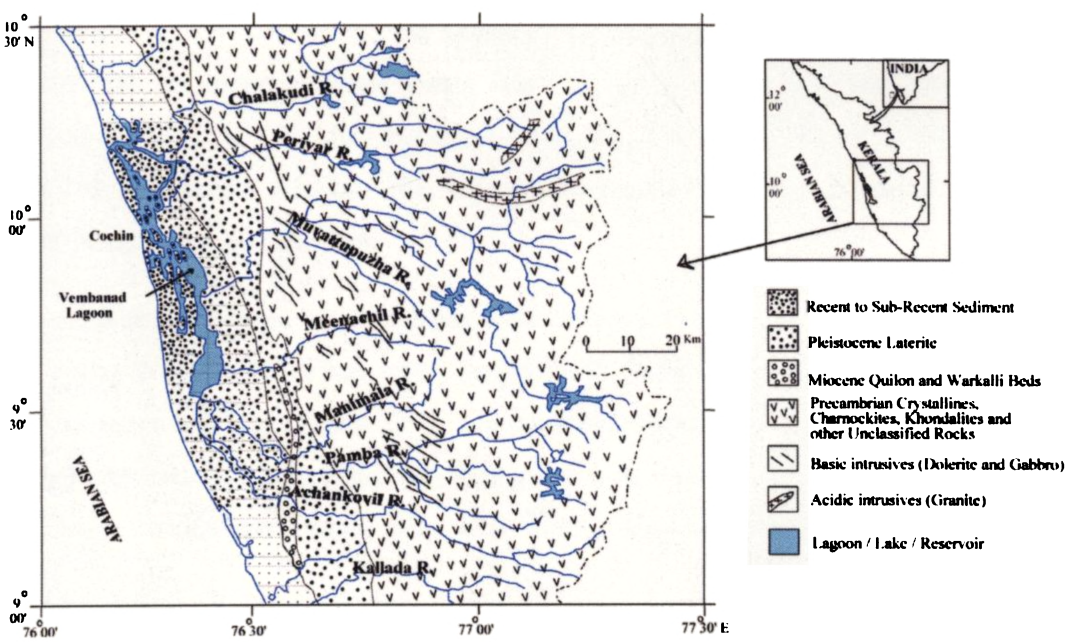

Figure 1.2 Geology, drainage and watershed areas of the Central Kerala (CESS, 1984, GSI, 1995),

1.9. Monsoon and Climate

Monsoon is considered as one of the most important air-sea interaction process of tropical climate system. Half of the tropics of the entire globe are under monsoonal climate. The monsoons play a significant role in modulating global climate through various forcing mechanisms. It has been debated by various paleoclimatologists that during the last glacial period, changes in both summer and winter monsoons (Schulz et al., 1998; Burns et al., 2003; Wang etal., 2001 and references therein) have occurred and varied synchronously with millennial scales. During the Holocene Indian monsoon (Sirocko et al., 1996; Gupta et al., 2003; Neff et al., 2001., Fleitmann et al., 2003) exhibited millenial scale variability and can be correlated with Atlantic climate variability.

Monsoons are the seasonally reversing winds which bring rain to the Indian subcontinent and cause upwelling along the continental margins. The seasonal reversal of the wind direction between summer and winter drives the southwest (SW) and northeast (NE) monsoons in the Indian Ocean and precipitation over south Asia. In summer, differential heating of the continental and oceanic regions leads to low atmospheric pressure above the Asian Plateau and high atmospheric pressure over the relatively cool southern Indian Ocean. This results in a strong low level jet stream, the Findlater Jet (Smith et al., 1991). Much of the intensity of this jet derives from the direct heating of the troposphere above Asia and through latent heat collected over southern subtropical ocean which is transported across the equator and released by precipitation over south Asia (Clemens et al., 1991). The winter monsoon (NE monsoon) is charecterised by low seasonal insolation over Asia and relatively high albedo due to seasonal snow cover. These boundary conditions produce a high pressure cell in the low level atmosphere over Asia which results in a northeasterly wind flow over the Arabian Sea. The winter monsoon is relatively weak and less significant in the Arabian Sea.

The south west coast of India experiences a tropical climate with seasonally reversing wind patterns and large variations in precipitation. Temperatures in the region range between 23 and 37°C., the southwest (SW) monsoonal wind of oceanic origin are established by mid May along the west coast of India. During the SW monsoon, winds blow from southwest during May-September, but change to a northeasterly direction during the northeast (NE) monsoon season. These winds continue to grow strong until June, when there is a sudden burst or strengthening of the southwest winds. The winds are the strongest during July and August, but become weak in September, ahead of the NE monsoon season, which lasts through October and November. Summer monsoon (June-September) account for the major part of the annual rainfall (>300 cm), whereas winter monsoon (October-January) account for about 50- 60 cm rainfall. Western Ghats plays an important role in the intensity and distribution of rainfall over the west coast of India (Narayana, 2006). The area is experiencing wet and dry climate.

1.10. Hydrography of the Arabian Sea

The climate is found to influence the hydrography of the northern Indian Ocean. Climate in the Arabian Sea region is characterized by strong monsoonal winds invoking large seasonal changes in hydrography and particle fluxes. During May to September the winds of the wet and hot SW monsoon in the region are directed towards the hot, low-pressure zone of the High Himalayas and Tibet (the "SW monsoon"). During the northern hemisphere winter, the circulation changes drastically: a high pressure zone over Siberia produces south- and eastward directed cool and dry continental air masses which flow from the Asian continent towards the Indian and west Pacific Oceans (the "NE monsoon").

In the Arabian Sea, seasonal reversal of the monsoonal winds during summer and winter produces spatial changes in surface circulation, hydrography and biological productivity. The surface water productivity is at its maximum in the western Arabian Sea during summer, when intensified coastal and open ocean upwelling occurs off Somalia and Oman (Banse, 1987; Prell et al., 1990). In the northeastern Arabian Sea, high biological productivity is recorded during winter which is linked to convective mixing because of surface water cooling, resulting into the injection of nutrients to the surface waters (Madhupratap et al., 1996). Biogeochemical cycling processes in the Arabian Sea are controlled by both the southwest (SW) and northeast (NE) monsoon wind systems of the summer and winter seasons respectively. The Arabian Sea is highly productive with increased vertical fluxes of particulate organic carbon and suspended particulate matter (Krey, 1973).

The peculiar monsoon wind regime produces dramatic seasonal changes in physical and biogeochemical processes in the upper water column. Wind induced upwelling of nutrient rich water and related high primary productivity is one of the characteristics of this basin (Kobanova, 1968; Nair et al., 1989). High productivity and non availability of well ventilated water to the intermediate depth result in the development of extremely low oxygen mid-depth layer causing intense denitrification (Naqvi, 1987), thus making the basin one of the world's largest nitrogen sink (Codispoti, 1995). The formation of high salinity water in the northern Arabian Sea (Rochford, 1964; Morrison, 1997) due to excess evaporation-precipitation (E-P), and its seasonal spreading south ward along the eastern region (Prasannakumar and Prasad, 1999) are unique seasonal hydrographic features of this basin. Therefore, the Arabian Sea is a complex but interesting natural laboratory to study the past climate.

Seasonal reversal of the monsoon winds over the Arabian Sea influences not only the surface circulation, productivity and biogenic and lithogenic fluxes; but also CO_2 uptake and heat budget in the region. The SW monsoon s accompanied by strong winds, cloudy skies and moist air. In addition, the ocean gains heat (an average of 89.5 Wm^{-2}) but SST decreases by 5.5°C and the mixed layer deepens to almost 80 m in the central Arabian Sea and becomes shallow at the Oman margin compared to the inter mansoon seasons (Schott and McCreary, 2001). The salinity in the upper water column during the SW mansoon is reasonably uniform, but decreases from about 36.5 to 36.0 ‰ during the course of the SW monsoon season. The SW monsoon wind field causes upwelling along the Arabian coast, leading to persistent jets of cold up welled water extending from the Oman coast and often advecting laterally about 60 km offshore (Brink et al., 1998: Fischer et al, 2002). Outside of the regions of coastal upwelling during the SW monsoon, SSTs are generally high, varying from 27 to 29°C. The NE monsoon is characterized by moderate winds, clear skies and dry air. During the NE monsoon the ocean loses heat to the atmosphere with a 3°C SST decrease. The two inter monsoon seasons are substantially different from either of the monsoon seasons, and are characterized by weak winds and strong sea-surface heating. The winds are generally weak during the inter-monsoonal periods, with the fall inter monsoon being significantly shorter than the spring inter monsoon (Weller et al. 2002). The SW monsoon determines the climate of the northern Indian Ocean during the northern hemisphere winter.

The surface circulation in the Arabian Sea is modulated by the seasonal variation of the monsoon wind system. The seasonal reversals of the surface wind field over the tropical Indian Ocean are far more dramatic than in other regions of low latitudes, and these reversals have profound impact on the surface current system (Wyrtki, 1973; Hasternath and Greischar, 1991). During the summer monsoon period the low level southeasterly trade winds of the southern hemisphere extend across the equator to become southerly or south westerly in the northern hemisphere. On the other hand, during the winter monsoon period, the oceanic circulation is relatively weak and is characterized by the north equatorial current, and the east ward flowing equatorial counter current. Thus, the direction of the wind–flow from the continent towards the Arabian Sea (in south west direction) causes the circulation in the Arabian Sea to reverse. Circulation of water at the ocean surface is largely a response to the overlying atmosphere circulation which exerts drag at the surface. However, circulation of deeper water in the oceans of the world is a consequence of density variations, which result from differences in temperature and salinity brought about by sensible and latent heat fluxes, precipitation, and runoff at the ocean surface, this is termed the thermohaline circulation. It is now well documented that numerous rapid changes in ocean circulation and atmospheric conditions took place through out the last glacial periodic studies of well dated high resolution sediments and uplifted coral reefs provide particularly valuable insights into such events at the very end of the last glaciation.

1.11. Geological and Oceanographic Settings

Rivers are the principal agents of transport of detrital sediments into the eastern Arabian Sea. It experiences a conspicuously wet and dry climate. The Deccan Traps cover an area of about 500 000 km^2 with a maximum thickness of 1200-1700 m in central and NW and are overlain by Quaternary sediments. The Western Ghats mountain range is to be found composed of two major rock types: the Deccan Trap basalts between Tapti and Morumugao and Precambrian gneisses and schists between Mormugao and Cape Comorin. The Western Ghats experience a humid tropical climate with a maximum rainfall of about 3000 mm. Sahyadri is drained by several minor rivers that bring sediment load into the coastal Arabian Sea. The Ghats stretch along the shore between Mormugao and Bhatkal as well as between Quilon and Cape Comorin. Here the coastal region is occupied by recent alluvium and the Warkalli beds of Tertiary age (ferrugenised sandstones with intercalated clays). Extensive lateritisation of parent rocks is a characteristic feature in western India. The Archean and Proterozoic rocks display a more complex history of lateritisation (Fig.1.3).

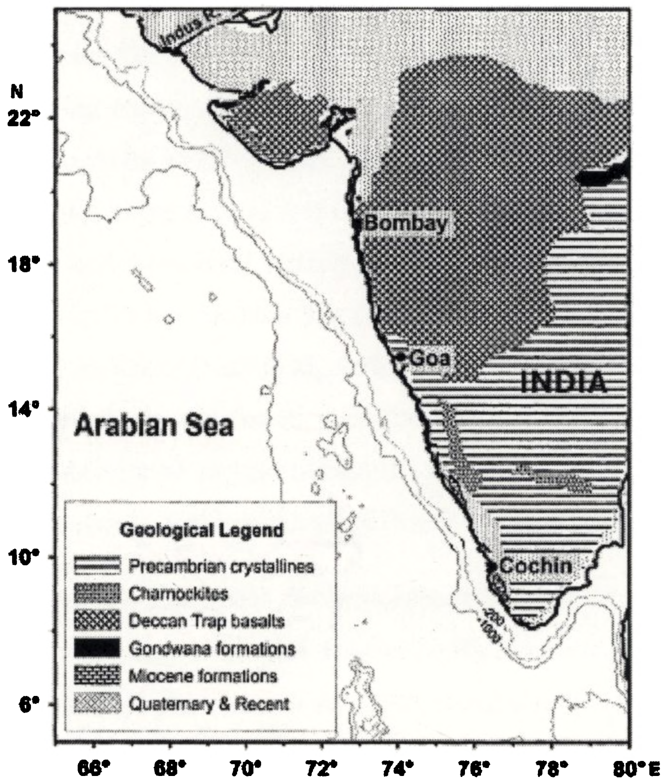

Figure. 1.3. Geologic map of the western coast of India.

The width of the continental shelf is widest (345 km) off Surat–Tarapur and narrows down to 60 km off Cochin. The shelf break occurs at about 120m in the northern part and at about 80 m in the southern part.

Two major sediment types occur on the continental shelf: modern clastic clays on the inner shelf and relict sandy sediments on the outer shelf (Rao and Wagle, 1997). The continental slope is covered by silty clays which are an admixture of dominant terrigenous and biogenic components.

Circulation in the Arabian Sea is controlled by the seasonal reversal of the winds caused by the alternate heating and cooling of the Tibetan Plateau and the resulting pressure changes. During summer (June through September) strong southwesterly winds blow across the Arabian Sea, which cause offshore Ekman transport and intense upwelling along the Oman and Somalia margins and the southwest coast of India (Wyrtki, 1973).

The upwelling process brings cold, nutrient-rich waters from a few hundred meters depth into the surface and fuels the biological productivity in the euphotic zone. Thus, the Arabian Sea is one of the most productive regions of the world oceans (Qasim, 1977). During winter, northeasterly winds suppress the upwelling and cause low biological productivity in the western Arabian Sea. In contrast, the northeasterly winds cause upwelling in the eastern Arabian Sea. Therefore, the seasonal reversal of the wind direction and associated circulation pattern has a direct bearing on the biological productivity in the Arabian Sea (Kobanova, 1968) and the lithogenic and biogenic flux to the sediment (Nair et al., 1989). In addition to the strong monsoonal character of the atmospheric and oceanic circulation, the Arabian Sea is unique by the presence of a well developed oxygen minimum zone (OMZ) with one of the largest marine volumes of suboxic ocean water today (Morrison et al., 1998).

Summer monsoon circulation is charecterised by atmospheric heating and the development of low pressure over Asia relative to higher pressure over the southern sub tropical Indian ocean. The summer monsoon is marked by strong south westerly winds over the Arabian Sea while cyclonic circulation about the Asian low creates north westerly winds over the Arabian Peninsula. The northwest flow over the northwest Arabian Sea drives strong coastal and open-ocean upwelling systems with consequent responses in a large array of oceanographic variables including SST, mixed layer depth (MLD), nutrient content and productivity (Dickey et al., 1998; Honjo et al., 1999, Smith et al., 1998, 1999; Weller et al., 1998).

Asian subsystems are driven by different boundary thermal conditions associated with land ocean configuration and topography, investigation of their different characteristics over annual cycle may enhance our understanding how tectonic and solar orbital forcing affects paleomonsoon variation. Likewise study of different responses and feed backs of these monsoon sub-systems may shed light on how internal dynamics of the coupled- atmosphere-ocean system influences the monsoon inter-annual variability.

CHAPTER-II

Methodology

To achieve the scientific objectives discussed in chapter I, sediment cores recovered from the continental slope region of the southeastern Arabian Sea were studied in detail for C–14 ages, sedimentation rates, sediment grain size, organic carbon, calcium carbonate, clay mineralogy, and major and trace elements.

2.1. Sediment cores

Three gravity cores were collected during the 38th cruise of M/V A. A. Siderenko, a Russian research vessel chartered by the Department of Ocean Development, New Delhi and during the 145th cruise of O.R.V. Sagar Kanya, in the year 2001 and 1999 respectively. The details of core locations, water depth and length of sediment cores are given in Table 2.1 and Figure. 2.1. All the core samples used in this study are from the oxygen minimum zone.

Table 2.1. Details of the cores recovered from southeastern Arabian Sea.

Cruise No.	Core No.	Latitude °N	Longitude °E	Water Depth (m)	Length of core (m)
AAS-38	4	10° 15'N	75° 39'E	239	2.70
AAS-38	5	10° 16'N	75° 32'E	875	2.50
SK-145	B/C-8	9° 54'N	75° 32'E	745	2.70

A careful visual examination indicated that there was no distinct lamination or turbidites in the cores. As X-radiography was not carried out, we could not establish the possible role of reworking of fine clays derived from the shelf during lowered sea levels. Colour and lithology of the sediments were noted shortly after the recovery of the core onboard. For high-resolution analysis, sub sampling of cores was done on board at 1 cm interval for the top 100 cm and at 2 cm interval for the rest of the core. The sub samples were properly labeled and stored in polyethylene bags and immediately stacked in a freezer.

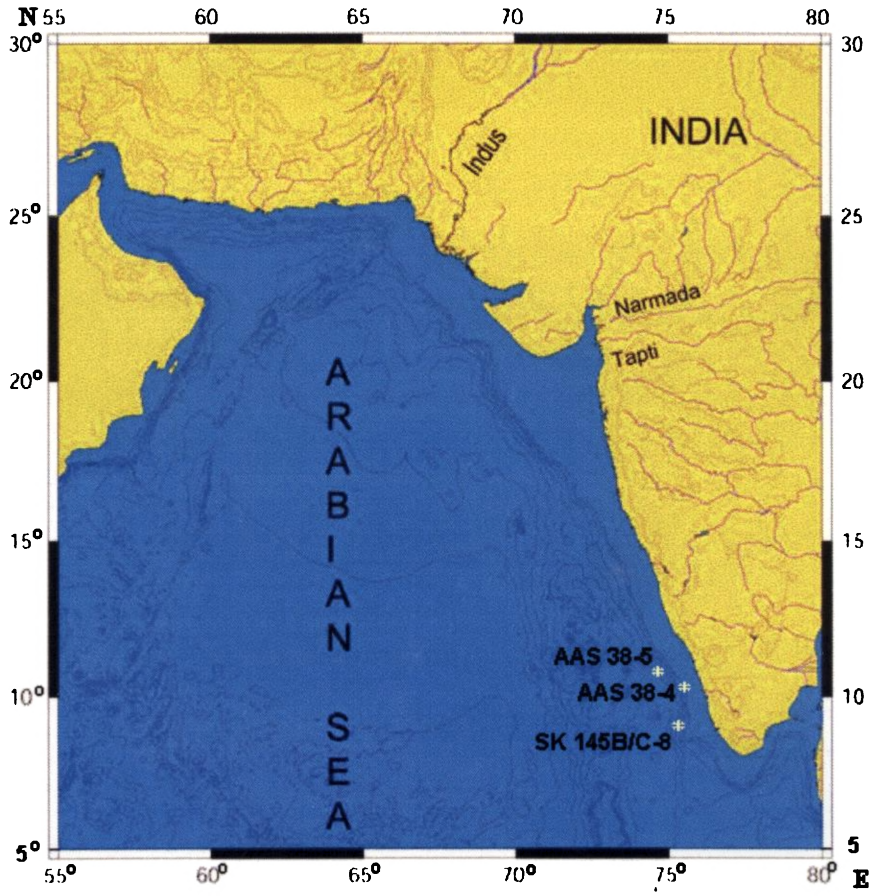

Figure 2.1. Location map of the study area showing core sample locations

The samples were brought to the lab and dried at 50°C temperature in a hot air oven, before carrying out further studies. The techniques and steps followed in the analysis of core samples are shown in Figure 2.2.

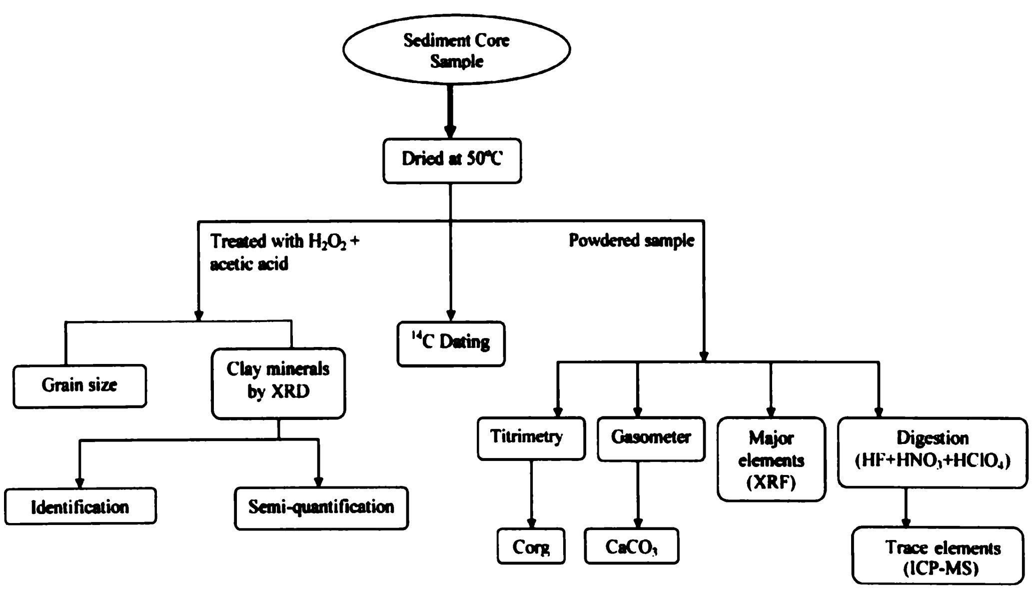

Figure.2.2. Flow chart shows the steps/ techniques used in this study

2.2. Radiocarbon (^{14}C) dating

During the course of geological time, equilibrium has been achieved between the rate of ^{14}C production in the upper atmosphere and the rate of decay of ^{14}C in the global carbon reservoir. This means that the weight of ^{14}C estimated to be produced each year in the upper atmosphere is approximately equal to the weight of ^{14}C lost through out the world by the radioactive decay of ^{14}C to nitrogen with the release of β-particles ($_6C^{14} \rightarrow {}_7N^{14}$ + β+ neutrino). The total weight of global ^{14}C thus remains constant. Further, plants and animals assimilate a certain amount of ^{14}C into their tissues through photosynthesis and respiration. The ^{14}C content of these tissues is in equilibrium with that of the atmosphere because there is a constant exchange of new ^{14}C once the old cells die. As soon as the organism dies, this exchange and replacement of ^{14}C radioactive clock will be activated (Bradley, 1999).

For studies of Late Quaternary sediments, ^{14}C (radiocarbon) dating has proved to be an important input, because of the ubiquitous distribution of ^{14}C. The technique can be used to date the samples of peat, wood, bone, shell, paleosols, old sea water, marine and lacustrine sediments, and atmospheric CO_2 trapped in glacial ice. The useful time frame for radiocarbon dating spans a period of major global environmental change that would be virtually impossible to decipher in detail without accurate dating control.

In most of radiocarbon laboratories, conventional methods -either proportional gas counters or liquid scintillation techniques – are employed. In gas counters, carbon is converted into a gas (methane, CO_2 or acetylene), when it is put into a "proportional counter" capable of detecting β particles. In liquid scintillation counters, the carbon is converted into benzene or some other organic liquid and detects scintillations produced by the interaction of β particle and a phosphor added to the organic liquid. In the present study, radiocarbon dating of sediment core samples was carried out by Gas Proportional Counting method.

The sediment core sample is acidified using mild orthophosphoric acid to extract carbon-dioxide under vacuum conditions (Yadava and Ramesh, 1999). The extracted CO_2 is purified by trapping moisture and CO_2 in different traps cooled by various coolants. The purified is reacted with tritium free hydrogen gas in the presence of a catalyst in a stainless steel reaction vessel maintained at 450-500°C. The reaction between CO_2 and hydrogen takes for about 12-15 hours to form methane gas (CH_4). The synthesised methane gas is taken in to a glass vacuum system and purified. The synthesised methane gas is filled in to a gas proportional counter kept in a lead and iron shield. The gas proportional counter is given DC power to develop a potential gradient inside the counter. The charge developed inside the counter gets accelerated by the potential gradient. Finally, the negative charge is collected by the collector electrode. The collected charge gets converted to voltage pulses by the counting electronics. The beta radiation emitted by the carbon-14 atoms is measured in anti-coincidence channel. The net beta count rate is obtained by subtracting the total count rate from the background count rate. The net count rate of the given sample is converted to percent modern carbon upon comparison with the modern standard net count rate. The percent modern carbon value is used to calculate age of the sample based on 5730 year as the half life of carbon-14.

The bulk sediments of core samples were dated for radio carbon ages using gas proportional counter at National Geophysical Research Institute, Hyderabad. The ^{14}C dating was carried out on sediment layers representing different sections of the core. In total, seven sub-samples covering the entire length of the core AAS 38-4 have been dated; whereas only four sub-samples covering the top 100 cm of the core AAS 38-5 could be dated. However, the age obtained for the sub-sample at 100 cm of core length has been discarded because of erroneous result. Assuming the linear sedimentation rate, the radiocarbon ages were computed using linear regression equation for the entire core length. The core SK 145 B/C–8 has not been subjected to radiocarbon ages. However, the ages obtained for AAS 38-5 were used for interpretation of the data of the core SK 145 B/C-8, as both the core sites are close to each other and have similar physiographic conditions.

2.3. Grain size analysis

In general the term texture represents the size, shape, roundness, grain surface features and fabric of grains. Because of differential erosion, transportation and deposition sediments laid down in different depositional environments may possess distinct particle size distributions. By determining the particle size distributions it is possible to talk about the environment of deposition and so to utilize this technique as a tool for environmental reconstruction (Lario et al., 2002). Textural parameters comprising sand, silt and clay ratios of the sediment core samples were carried out using standard methods (Carver, 1971; Folk, 1980).

Grain Size analysis of core samples was carried out at 4 cm interval following the standard sieve and pipette techniques and a total of 150 samples were analysed. About 15 to 20 g of dried sediment sample was weighed accurately and transferred into a thoroughly cleaned 1000ml beaker. The samples were made salt free by repeated washings using distilled water and subsequently treated with H_2O_2 and glacial acetic acid in order to remove the organic matter and carbonates respectively. The samples were wet sieved through a 63μm sieve to separate sand and mud fraction. The sand fraction (>63μm) retained in the sieve was dried and weighed, while the mud fraction was collected in a 1000 ml measuring glass cylinder and subjected to pipette analysis to estimate the silt and clay contents following the method of Folk (1980). Percentage distribution of sand, silt and clay fraction in each sample was determined. Size classification adopted is sand >63μm, silt 63 - 4μm and clay < 4μm. Based on the relative abundance of sand, silt and clay in each sub sample in each size class, sediment texture was determined according to the classification of Folk (1980). The sand-silt-clay ratios were plotted on trilinear diagram for textural classification / nomenclature based on Folk's (1980) textural classification.

2.4. Organic carbon (C_{org}) analysis

Since the oceanic CO_2 concentration is supposed to be in equilibrium with the CO_2 concentration of the atmosphere, to understand the global carbon cycle it is important to know how much organic carbon is buried in marine sediments, and how much has been withdrawn from the carbon cycle. Further the information on organic carbon helps to understand biological productivity in the aquatic regions. Organic carbon in marine sediments is primarily derived from settling biogenic debris from the water column and their abundances serve as indicators of primary productivity (Muller and Suess, 1979). The organic carbon concentrations in sediments, especially in the continental margin region, depend up on the depositional environment, sedimentation rate, bottom water redox condition etc. (Canfield, 1994).

The relationships between monsoon, primary productivity, preservation and accumulation of organic matter in the Arabian Sea as well as their variations with time were investigated using different methods. There are still many uncertainties and controversies concerning the factors controlling organic matter accumulation and its linkage to monsoon and primary productivity. In order to understand the distribution pattern, its relation to productivity and paleoenvironment and paleoclimate of the region, organic carbon in sediment cores was estimated.

Samples were oven dried at 50°C and finely ground in agate mortar for the determination of organic carbon content. The organic carbon content (C_{org}) of sediment samples were determined by using a modified Elwakeel and Riley method (Gaudette et al., 1974) which is based on the exothermic heating and oxidation of organic matter. The sample was treated with potassium dichromate and concentrated sulphuric acid, followed by back titration with ferrous ammonium sulphate using diphenyl amine as an indicator. The principle behind this method is based on the oxidation of organic carbon with chromic acid and titrimetric determination of the oxidant consumed.

2.5. Calcium carbonate ($CaCO_3$) determination

The accumulation of calcium carbonate on the sea floor is mainly controlled by the surface water biological productivity, rate of dissolution during its journey through the water column as well as on the sea floor and dilution by the non-carbonate fraction and terrigenous matter (Naidu and Malmgren, 1996). The carbonate contents in sediment column are mainly used to understand the dilution effects influenced by terrigenous supply, which in turn has a bearing of monsoon.

The carbonate content in sediment cores was determined by rapid gasometric technique, following Mueller and Gastner (1971) method. This method is based on the volumetric determination of CO_2 released by acidification of the powdered sample with IN HCI solution.

An amount of 0.5g of the 99.5% standard $CaCO_3$ was taken in a 50ml conical flask. About 5ml of concentrated HCI was taken in a small glass vial and carefully kept inside the conical flask. The flask was then fitted to a simple apparatus for producing and trapping the CO_2 evolved from $CaCO_3$. The $CaCO_3$ was allowed to react with HCI by shaking the flask vigorously. The CO_2 produced by the reaction was then allowed to pass through a distilled water column, where it displaces certain amount of distilled water. The amount of distilled water displaced was collected in a standard measuring cylinder, which is directly proportional to the CO_2 produced during the reaction with carbonate and in turn to the percentage of $CaCO_3$ available. The experiment on the similar line was carried out for the sediment samples (i.e., 0.5 g of powdered bulk sediment sample was added to 5 ml HCl in to the conical flask and it was allowed to react with HCl). The volume of the displaced liquid after the reaction was noted each time. Then, the percentage of $CaCO_3$ available in the sediment sample is calculated, using the equation,

$$CaCO_3\,\%\text{ in the sample} = \frac{([\text{standard } CaCO_3] \times \text{vol. of liquid displaced for the sediment})}{\text{Vol. of liquid displaced for standard } CaCO_3}$$

A total 60 samples (core AAS 38-4 & 38-5) were analysed for $CaCO_3$ measurements. We could not do the $CaCO_3$ estimation for the other core (SK 145B/C-8) because the facility available was limited. Replicate analyses of both samples and carbonate standards show that the analytical precision is $\pm$ 1%.

2.6. Clay mineral analysis

At present, the palaeoclimatic interpretations of marine clay assemblages are yielding, at best, rather broad palaeoclimatic information. This approach cannot achieve the same degree of resolution as other techniques such as isotope or microfossil studies. Clay mineral assemblages may provide integrated records of overall climatic impacts, whereas other techniques are more likely to reveal local or temporary climates.

Clay minerals can be used as a tool a powerful source for the interpretation of marine depositional processes and their study also reflects weathering conditions imparted on the source rock. Smectite and kaolinite were used to elucidate depositional origin, where as chlorite and kaolinite have been used as valuable indicators of provenance, paleoenvironment and paleoclimate (Chamley, 1989). The clay minerals in the core samples were analysed employing the following steps.

2.6.1. Sample preparation:

Approximately 15-20 g of the dried sample was taken and dissolved salts were removed by washing thoroughly with distilled water. This process has been continued till the sample was deflocculated. The presence of organic matter in the sample can produce broad X-ray diffraction peaks, increase the background, and inhibit dispersal of other minerals if present in significant amount. Therefore, the organic matter from the sediment samples was expelled by treating with 20-25 ml of 30% H_2O_2. The sample was again washed with distilled water. The sample was also made free of carbonates by treating with 10 ml glacial acetic acid. Excess acid was removed by washing with distilled water.

After the sample is disaggregated and deflocculated, >63μm size fraction sediments were separated by wet sieving and the supernatant was collected in a 1000 ml cylinder. The 2μm size fraction of sediment was separated based on the settling velocity (Stoke's law) principle. The resulted suspension was transferred to a settling column and allowed to settle for 3 hrs 27 minutes i.e., until the required size fraction was obtained. The clay water suspension was used for making oriented clay slides of almost equal size and thickness by pipetting equal volume (1ml) on to glass slides. The glass slides were allowed for air-drying at room temperature, while drying, care was taken to avoid contamination by dust and other means (Biscaye, 1964; Gibbs, 1965).

2.6.2. X-Ray diffractogram

X-ray diffraction methods are commonly used for identification of mineral components in soils, sediments and rocks, particularly for clay minerals in sediments and soils. X- ray diffraction studies were carried out on the air dried samples (Gibbs, 1965). The slides were run on X-ray Diffractometer (XRD) from 2 to 30° 2θ at 1° 2θ / minute using Ni filtered CuK∞ radiation. The XRD facility at National Institute of Oceanography, Goa (Philips-1840 Model), and at Ocean Science and Technology Cell, Mangalore University (Model-D8 Advance) was availed. The samples were then glycolated by exposing the slides to ethylene glycol vapours at 100°C for one hour and rescanned the slides under same instrumental settings for the confirmation of montmorillonite. The peaks for different clay mineral groups like kaolinite, chlorite, illite and smectite were identified. In order to differentiate the kaolinite and chlorite peaks, the samples were also scanned from 24 to 26° at 0.5° 2θ/minute (Biscaye, 1964).

2.6.3. Semi-quantification of clay minerals

The relative contents of clay minerals were determined using peak areas of smectite (15-17 Å), illite (10 Å), chlorite (7 Å) and kaolinite (7 Å). The peak areas of the spectra of these clay minerals were calculated by using the glycolated X-ray diffractograms and relative weight percentages were calculated following the semi-quantitative method of Biscaye (1965).

The relative clay mineral contents of smectite, illite, kaolinite and chlorite were determined using ratios of integrated peak areas of their basal reflections, weighted by empirically estimated factors. Accordingly, 17° Å peak area of the smectite is multiplied by 1, the 10 Å peak area of illite by 4, and both the kaolinite and chlorite proportions positioned at 7 Å peak were multiplied by 2. As the 001 plane of kaolinite and 002 plane of chlorite partly overlap each other at 7 Å, the relative proportions of both minerals were first deduced from the areas of 002 kaolinite and 004 chlorite reflections at 3.54 Å and 3.58 Å respectively. The clay minerals quantified in this study are smectite, kaolinite, illite and chlorite.

2.7. Major elements

High resolution, continuous, and quantitative chemical analysis of sedimentary records provides useful information on paleoclimatic changes (Peterson et al., 2000; Haug et al., 2001). The contents of biogenic silica, biogenic carbonate, and terrigenous material can be estimated from major element composition and combination, and examination of temporal and spatial variations in the contents of these components is useful for reconstructing paleoclimate and paleooceanographic changes.

In general, sediment is derived to the marine environment from more than one source. One of the challenges facing sedimentologists and geochemists is to unravel these sources, both because of the paleoenvironmental information they carry, and because mobile elements will interact differently with the different sediment components. Elements are supplied from land to the sea in two different states (dissolved and particulate) as the result of weathering of crustal rocks. Particulate elements are transported by rivers as suspended matter and through the atmosphere as airborne dusts (Delany et al., 1967; Tsunogai et al., 1985; Uematsu, 1987), and dissolved elements by rivers (Martin and Whitfield, 1983). The proportion of dissolved to particulate differs with each element (Chester and Messiha-Hanna, 1970). Airborne dusts and suspended matter in river water generally have the chemical composition of average shale rather than of the crustal rocks (Li, 1981). The major elements were analysed in order to understand the source and dilution effect on $CaCO_3$.

2.7.1. X-ray fluorescence spectrometer (XRF) method

Major element concentrations were determined by X-ray fluorescence spectrometry as per the procedures given in Calvert (1990). Selected dried samples were powdered, and packed in polyethylene bottles and used for major element analysis by conventional XRF method. The XRF intensities measured from elements in dry samples may increase or decrease due to the XRF absorption and enhancement effects by other elements, called the matrix effect (Tertian and Claisse, 1982). In order to evaluate the matrix effect on the XRF intensities, plates of aluminium, titanium, iron and quartz and powdered potassium nitrate and calcium carbonate were prepared as the standard samples. Pressed pellets are prepared by using collapsible aluminium cups having 12 mm diameter (Govil, 1985). These cups are filled with boric acid and about 1 g of the finely powdered sediment sample is put on the top of the boric acid and pressed under a hydraulic press at 20 tons pressure to get a pellet. A Philips MagiX PRO model PW 2440 X-ray fluorescence spectrometer coupled with automatic sample changer PW 2540 is used for the major elemental composition determination. The MagiX PRO is a sequential instrument with a single goniometer based measuring channel covering the complete measurement range from few ppm to 100% concentration. The rhodium (Rh) anode with end window is normally used in X-ray tube.

The concentrations of major elements in 60 samples at different core depths were analysed employing the Philips MagiX PRO model PW 2440 X-ray fluorescence spectrometer at National Geophysical Research Institute, Hyderabad. The major elements determined are: Si, Al, Fe, Ti, Ca, Na, K, Mg, Mn and P. The precisions were of + 3% for the major elements. The major elements in sediment cores AAS 38-4 and AAS 38-5 were determined, and samples of SK 145B/C-8 could not be subjected to major element analysis as the XRF facility availability was limited.

2.8. Trace elements

The varieties of processes result in trace element enrichments that mirror the specific conditions prevailing at the time of deposition and early diagenetic stages. Consequently trace element abundances in sediments allow us to reconstruct paleodepositional conditions (Werne et al., 2003; Lyons et al., 2003; Riboulleau et al., 2003; Rimmer, 2004; Rimmer et al., 2004; Algeo and Maynard, 2004; Algeo, 2004; Nameroff et al., 2004; Tribovillard et al., 2004a). While using trace element concentrations to reconstruct paleoenvironmental conditions, one must assess whether they are relatively enriched or depleted. In general, the degree of enrichment or depletion of a trace element in a sample is evaluated relative to its concentration in a reference that is commonly the average crustal rocks or average shale (Wedepohl, 1971, 1991; Mclennan, 2001). In general, trace elements in fine grained sediments, which are relatively rich in organic matter, are used for paleoenvironmental reconstruction. Rare earth elements are important because their geochemical properties enable them to be powerful tracers of chemical processes. Trace elements are analysed using the following procedure:

2.8.1. Dissolution of sediment samples:

To a 50 mg sample, 10 ml acid mixture of 6 parts HF, 3 parts of HNO_3 and 1 part of HCl were added along with 0.5 ml of 10 mg/ml Rh in microwave vessel to act as internal standard. The vessels were then closed and mounted in teflon beaker and heated. This procedure is repeated once again to ensure total dissolution of the total samples. After completion of the heating, the beakers were cooled to room temperature and carefully kept in a fume-hood and the then 1ml of $HClO_4$ is added to each beaker. The solution was evaporated to incipient dryness. The residue was dissolved in 20 ml of 1:1 HNO_3 and brought to a final volume of 250 ml. Clear solutions were obtained in all cases. A procedural blank solution was also prepared (Balaram and Rao, 2003).

Electronic grade HF, analytical reagent (AR) grade $HClO_4$ and distilled HNO_3 and HCl were used in the preparation of samples and standards. Distilled / deionised water was used for all analytical purposes.

2.8.2. Inductively Coupled Plasma-Mass Spectrometer (ICP-MS)

Trace element analyses were carried out on an Inductively Coupled Plasma – Mass Spectrometer (ICP-MS-Perkin Elmer SCIEX, Model ELAN R DRCII) at National Geophysical Research Institute, Hyderabad. The important optimisation criteria include adjustment of nebular gas flow, setting of detector and lens voltages, radiofrequency forward power settings and performance of calibration.

The principles of ICP-MS are based on the assumption that, ions may be generated in a suitable ionising source such as an ICP for all elements in the sediment. Ions are physically extracted from the plasma into mass spectrometer and measured using an ion detector. Sample introduction for plasma spectrometry is generally accomplished using solution nebulisation. Sample solutions are aspirated by a nebuliser which shatters the liquid into fine droplets using an Ar stream of ~1L/minute. These droplets are directed into a spray chamber which removes the unsuitable larger material, and allows only the finest spray to pass into the plasma (Balaram, 1997). The ICP-MS offers very low detection levels. The analytical accuracy was checked by analysing international standards (Marine mud / MAG-1).

A total of 100 samples (35 from AAS 38-4; 30 from AAS 38-5; 35 from SK 145B/C-8) representing different sedimentary column of each core were analysed. The trace elements determined are V, Cr, Co, Ni, Cu, Sr and Ba.

CHAPTER-III

Sedimentation, Organic Carbon and Carbonate Records: Paleoclimate and Paleoenvironmental Controls

3.1. Introduction

Weathering and erosion processes in different climatic zones may result in characteristic inorganic products. The bulk of sediments deposited on continental margins were carried by rivers and about 6-11 billion metric tons of sediment accumulate in the ocean basins annually, and thus provide an archive of climatic conditions, near the ocean surface or on the adjacent continents, and the oceanic/atmospheric circulation prevailed at the time of deposition (McManus, 1970; Kolla et al., 1979). These sediments provide a record of past climate and oceanic circulation, in terms of surface water temperature and salinity, dissolved oxygen in deep water, nutrient and trace element concentrations.

Although the paleotopography of the late Pleistocene appears as basic factor in controlling the sedimentation records on the continental shelf and slope region, other factors such as sea level fluctuations and climate play an important role. Climate can affect sediment input, which in turn affects sediment records. Variations in sediment texture and sedimentation rates suggest varied hydrolysis and monsoon conditions in the hinterland region. The changes in marine sedimentation rates in the proximity of continents also provide preliminary clues about the past variations in aeolian dust, fluvial erosion input, marine productivity and the past climatic conditions (Prins et al., 2000; Sirocko et al., 1993). The sedimentation rates are essential to estimate the mass accumulation rates of the components of the seafloor, which eliminates the bias (due to dilution effect), associated with interpreting direct weight concentrations and with varying physical properties of sediments.

Sediments, be they continental or marine, are the best proxies since it is easily available for paleoclimatic studies (Ramesh and Somayajulu, 1992). As far as geochronology is concerned, marine sediments are more suitable for radiometric dating than their continental counterparts and, as well, offer continuous records. Several recent studies of marine sediments from the Arabian Sea have demonstrated their utility for reconstructing abrupt climate change during the last glacial period (Kolla et al., 1981; Rao and Rao, 1995; Thamban et al., 2001; Pratima et al., 2007).

Sediment distribution and deposition on continental shelves result from a complex interplay of tectonics, sea level change, climate, and oceanographic processes. Coarse grained particles of marine sediment cores provide evidence of former ice-rafting episodes. In the late Quaternary, there were quasi-periodic episodes of major ice rafting; and these are now termed Henrich events (Henrich, 1988).

Sediments of southwestern continental shelf and slope of India consist predominantly of river borne clays, and the aeolian fraction is meager (Naidu et al., 1985; Rao and Rao, 1995). Studies on sediment accumulation rates on the south western continental margin of India are limited and clustered at few places (see Pandarinath et al., 2004 and references there in). Most of the studies were on sediment cores of shallower region which can be grouped in to two categories - (i) modern or present day sedimentation rates obtained from top few centimeters of sediments by ^{210}Pb dating techniques (Borole, 1988): and (ii) slightly longer sedimentation records obtained by dating deeper sediment layers by ^{14}C method (Nambiar et al., 1991; Caratini et al., 1994; Nambiar and Rajagopalan, 1995; Pandarinath et al., 1999; Thamban et al., 2001).

The organic matter that escapes decomposition is buried and preserved in marine sediments. However, it has been debated for long as to whether the amount depends on bottom-water O_2 concentration or there is no influence of O_2 on preservation (Canfield, 1994). It is argued that processes other than water-column anoxia can lead to high organic carbon concentrations (Pedersen and Calvert, 1990; Calvert and Pedersen, 1992). It is suggested that high productivity in the photic zone leads to a high burial rate of organic carbon on continental margins (Pedersen et al., 1992; Calvert et al., 1995; von Rad et al., 1999a). Demaison and Moore (1980) advocate that the presence of oxygen-depleted waters at intermediate depths, which slows down the decomposition of organic matter, enhances preservation. The organic carbon concentration of sediment may be highly influenced by the relative rain rate of organic matter versus diluting clastic debris. Therefore, organic matter serves as a reliable proxy in paleoproductivity and paleoenvironmental studies. Therefore quite a few studies on organic carbon variations along the eastern Arabian Sea (Paropakari et al., 1987, 1993; Calvert et al., 1995; Thamban et al., 2001; Pratima et al., 2007).

Since the first documentation of glacial/interglacial cyclicity of $CaCO_3$ in the equatorial Pacific by Arrhenius (1952), it has been debated extensively whether the productivity of $CaCO_3$ secreting organisms or dissolution of $CaCO_3$ exert an influence on the $CaCO_3$ cycles. In addition, the influence of dilution of $CaCO_3$ cycles by terrigenous input is also debated. In general, the amplitude of carbonate cycles increases and the average values of $CaCO_3$ decrease in areas influenced by terrigenous dilution and/or dissolution in the Indian Ocean (Olausson, 1965; Naidu, 1991; Naidu et al., 1993). Thus it is considered that dissolution and dilution are important factors in governing the fluctuations in $CaCO_3$ records.

Monsoons play a dominant role in controlling the regional climate, biological productivity and particulate flux in the northern Indian Ocean. Sediment trap experiments have demonstrated that biological productivity and terrigenous supply in the Arabian Sea is strongly linked to the intensity of the monsoons. Detailed studies have been carried out in the Arabian Sea to understand the monsoonal influence on the biological productivity and terrigeneous supply during the Late Quaternary (Sirocko and Sarnthein, 1989; Shimmield et al., 1990; Clemens et al., 1991; Murray and Prell, 1992; Naidu, 1991; Naidu et al., 1993; Naidu and Shankar, 1999; Bhushan et al., 2001). It is generally understood that the summer monsoon was stronger during interglacials than glacials (Prell et al., 1992 and references there in). Nevertheless, differential reaction of productivity proxies in different regions of the Arabian Sea leads to contradictory conclusions on the relationship between productivity and monsoonal strength (Clemens et al., 1991; Naidu and Shankar, 1999; Agnihotri et al., 2003; Ivanova et al., 2003).

The input of terrigenous material to the Arabian Sea is related to monsoon intensity and varies on a regional scale across the Arabian Sea (Sirocko, 1989). Increased sedimentation rate enhances the organic carbon preservation in sediments (Heath et al., 1977; Sarnthein et al., 1988).

In this chapter, sedimentation rates, grain size characteristics, organic carbon and calcium carbonate are evaluated in terms of paleomonsoonal intensity, paleoproductivity and paleoenvironment in the eastern Arabian Sea.

3.2. Radiocarbon ages of sediments

Radiocarbon dating remains the best tool for evaluating the ages and sediment accumulation rates in the time scale of several tens of thousands of years. Marine sediments are important components of the Quaternary stratigraphic record. The radiocarbon (C-14) ages of sediment layers at different depths of sediment cores have been successfully used by various researchers (Bhushan et al., 2001; Pandarinath et al., 2001, 2003; Thamban et al., 2002; Agnihotri et al., 2003) to understand the long–term time frame (Late Quaternary) paleoenvironmental, paleomonsoonal and paleoproductivity aspects along western continental margin.

C-14 dates were obtained for seven sedimentary layers at different layers of the core AAS 38-4 (Fig. 3.1a). The top layer (10-13 cm) of the core exhibits a radiocarbon age of 3030 $\pm$ 135 yr BP. The sediment layer at 48-50 cm gives the age of 4390$\pm$ 160 yr BP; 99-100 cm layer has recorded 6760 $\pm$ 180 yr BP; the layer at 148-150 cm depth shows the age of 10120 $\pm$ 230 yr BP. The radiocarbon ages of sediment layers at 198-200cm and 248-250 cm are 15605$\pm$ 320 yr BP and 18415 $\pm$ 405 yr BP respectively. The bottom layer (268-270 cm) of the core gives a radiocarbon age of 19445$\pm$ 465 yr BP. The three radiocarbon dates that were obtained for the core AAS 38-5 reveal similar sedimentation patterns as that of the core AAS 38-4 (Fig. 3.1b). The top layer (5-7 cm) of the core AAS 38-5 gives an age of 2100 $\pm$100 yr BP; the sediment column at 49-50 cm has the age of 4360 $\pm$ 110 yr BP and the layer at 74-75 cm is dated as 5990 $\pm$130 yr BP.

Based on these radiocarbon ages, geochronology of the sedimentary column and linear sedimentation rates at the core sites were computed.

3.3. Linear sedimentation rates (LSR)

The sedimentation rates are essential to estimate the mass accumulation rates of the components of the seafloor, which eliminates the bias (due to dilution effect), associated with interpreting direct weight concentrations of sediments with varying physical properties (Sirocko, 1989). The linear sedimentation rates were computed from the radiocarbon ages of sediment layers to understand temporal variations and depositional and paleomonsoonal conditions in the region.

The linear sedimentation rates computed from the radio carbon ages of the studied cores in the present work are shown in Figure 3.1. It is observed that the sedimentation rates in the shallow water core are consistently higher than in the deeper water core.

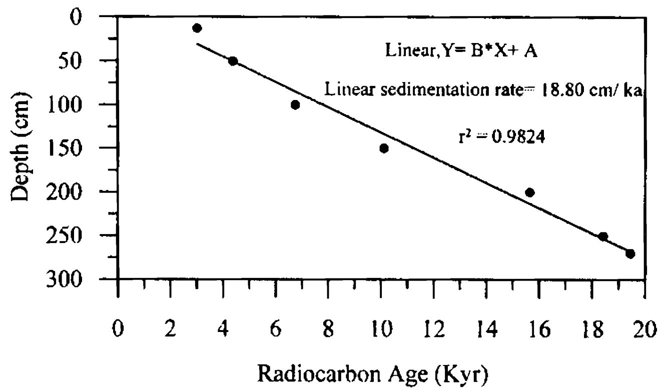

Figure 3.1a. Radiocarbon ages and the linear sedimentation rates at the core location of AAS 38-4.

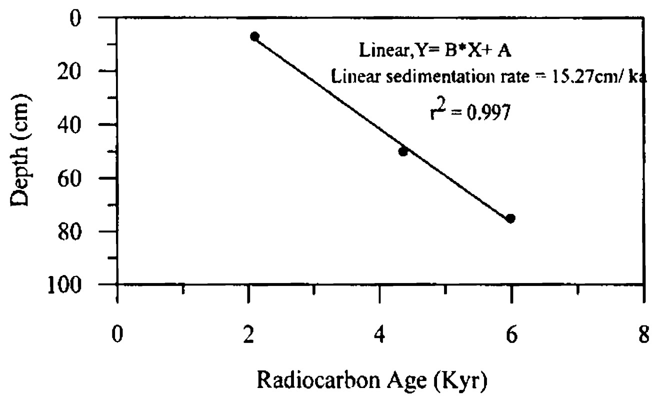

Figure 3.1b. Radiocarbon ages and the linear sedimentation rates at the core location of AAS 38-5

Sedimentation rate at the core site AAS 38-4 (239 m water depth) varies from 4.29-27.21cm/ka during the past 19.5 ka BP with an average rate of 18.8cm/ka (Fig. 3.2a). Sedimentation rate is very low (4.29cm/ka) during 0 - 2.3 ka BP From 3.1 – 4.3 ka BP, LSR increased drastically reaching up to 27.21cm/ka, where as during 4.6–6.76 ka BP LSR shows a slightly decreasing pattern and reaches up to 21.10cm/ka. During 7.3-10.1 ka BP it further decreases to 14.88 cm/ka. From 10.1-15.6 cm/ka BP it further decreases reaching up to 9.04cm/ka. Then LSR shows a considerable increase in sedimentation rate upto 18.08 cm/ka during 16-19.4 ka BP (Fig. 3.2a).

At the core site AAS 38-5 (875m water depth) the sedimentation rate varies from 3.33 to 19.03cm/ka with an average of 15.27 cm/ka. Low sedimentation rate was observed during 0-2 ka BP (3.33 cm/ka). During 2.5-5.3 ka BP very high (19.03

cm/ka) sedimentation rate was recorded. Beyond 5.3 ka BP slightly low (15.34 cm/ka) is recorded (Fig 3.2b).

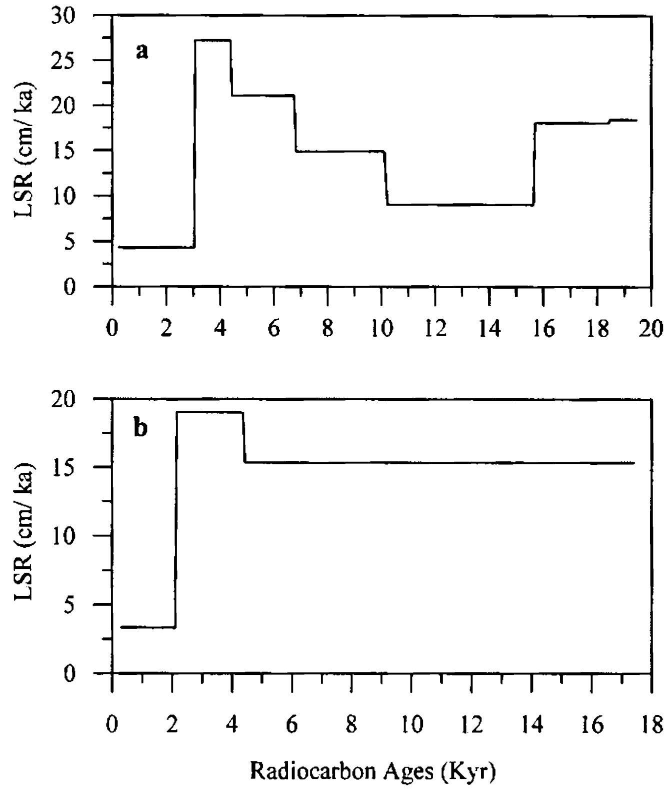

Figure 3.2. Temporal variation of LSR in the cores
(a) AAS 38-4 (b) AAS38-5

Transport and deposition of sediments on the sea floor depend on the flocculation of sediment particles, bottom topography, and prevailing hydrographic conditions on the shelf and slope regions. Understanding the rates at which sediment accumulation takes place in the marine environment is useful in a wide spectrum of fields, such as - to quantify the rate of detrital influx from the land, to trace the sediment pathways, to provide the required geochronological time frame in paleoclimatic and paleoproductivity studies, and in paleomonsoonal and paleoclimatic studies of the adjacent hinterland.

Pandarinath et al. (2004) have brought out an excellent review on the sedimentation rates of western continental margin of India. It is observed that sedimentation rates, in general, decrease with increasing water depths, suggesting the reduced influence of the terrestrial input with depth. However, there are exceptions to this generalities as observed by various researchers: (i) sediment accumulation rates are practically the same (~0.02mm/yr) at 370, 600 and 1680 m water depths on the continental slope off Mangalore (Somayajulu et al., 1999; Agnihotri et al., 2003); and (ii) the sediment accumulation rate in the slope region off Saurashtra (~3.8mm/yr) is greater than in the shallower shelf region (~ 1.8-2.5mm/yr; Borole, 1988). Similarly, sediment accumulation rates on the deeper slope (0.19 mm/yr; 776 m water depth) are considerably greater than those along the shallower slope (0.01mm/yr; 332 m water depth) south off Cochin (Pandarinath et al., 2003).

Sedimentation rates generally decrease from north to the south along the western continental margin of India. Probably such variations are influenced by the major rivers such as Indus, Sabarmati, Narmada, Mahi, and Tapti with high discharge and suspended particulate matter (SPM) drain into the northern part, whereas minor rivers such as the Netravati, Sita, Kollur-Haladi, Bharatipuzha, Periyar etc., with lesser bed load drain into the southern part. . The Indus river alone contributes 0.44x 10^9 metric tons (Holeman, 1968) and the Narmada and Tapti together discharge about 0.06x 10^9 tons (Borole et al., 1982) annually.

Another reason for such variation is that the rivers of the northern part drain through volcanic rocks (Deccan basalts), which are more susceptible to weathering than the granitic gneisses, charnockites and other mixed igneous and metamorphic rocks that form the catchment terrain of the southern rivers. The gradual decrease in sedimentation rates from northeastern Arabian Sea to its southeastern continental margin is also reflected in the width of the continental margin, where width is greater in the northern part and narrower in the southern part (Pandarinath et al., 2004).

Comparatively high sedimentation rates were observed on the continental slope (3.8mm/yr) than the outer shelf (1.8-2.5 mm/yr) off the Gulf of Khambat (Borole, 1988). Sedimentation rates on the continental slope off Saurashtra (Somayajulu et al., 1999) show a relatively small variation (~0.77-1.03mm/yr) over the last 1300 yr at a 280 m water depth, but at deeper levels (480m), rates vary widely from ~1.34mm/yr during ~ 8010-7480 yr BP to ~ 0.05 mm/yr during ~ 7480mm/yr BP to present (Table 3.1)

Sedimentation rates recorded during the Pleistocene and Holocene periods, at 1724m water depth in the Laccadive Trough region off Mangalore, were of ~0.04 and ~ 0.03mm/yr. The sedimentation rate on the upper continental slope region (water

Table 3.1. Late Quaternary sedimentation rates along the Southeastern Arabian Sea

Area	Core No	Location (water depth)	Core depth	Dating technique	Age in (yr BP)	LSR (mm/yr) (period)	Reference
Off Mangalore	RVG/207	41 m	0.50-0.60	C-14	1330 ± 80	0.45 (1330 yr BP- present)	Pandarinath et al., 1998
			0.90-1.00	C-14	2090 ± 80	0.53 (2090-1330 yr BP)	
	3268 G	12.53°N; 74.17°E (600 m)	0.00-0.01	AMS C-14	566 ± 25	0.02 (566 yr BP- present)	Somayajulu et al., 1999
			0.25-0.26	AMS C-14	3202 ± 60	0.09 (3202-566 yr BP)	
			0.50-0.51	AMS C-14	7176 ± 93	0.06 (7176-3202 yr BP)	
			0.71-0.72	AMS C-14	9876 ± 60	0.08 (9876-7176 yr BP)	
	3268 G2	12.52°N; 74.16°E (370 m)	0.03-0.04	AMS C-14	2056 ± 52	0.02 (2056 yr BP- present)	Somayajulu et al., 1999
			0.26-0.27	AMS C-14	6906 ± 132	0.05 (6906-2056 yr BP)	
			0.50-0.52	AMS C-14	6066 ± 116	-	
			0.61-0.64	AMS C-14	9485 ± 80	0.14 (9485-6906 yr BP)	
	CA3800	1724 m	0.10-0.15	C-14	4810 ± 110	0.03 (4810 yr BP- present)	Shankar and Manjunatha, 1995
			0.40-0.45	C-14	12910 ± 450	0.04 (12910-4810 yr BP)	
			0.65-0.70	C-14	11960 ± 260	-	
			0.70-0.75	C-14	17630 ± 360	0.06 (17630-12910 yr BP)	
			0. 95-1.00	C-14	26120±1230	0.03 (26120-17630 yr BP)	
			1.15-1.20	C-14	29990 ± 2680	0.05 (29990-26120 yr BP)	

	3104 G	12.83°N; 71.76°E (1680 m)	0.02-0.03	AMS C-14	1357 ± 90	0.02 (1357 yr BP- present)	Agnihotri et al., 2003
			0.09-0.10	AMS C-14	3311 ± 164	0.04 (3311-1357 yr BP)	
			0.19-0.20	AMS C-14	6469 ± 137	0.03 (6469-3311 yr BP)	
			0.34-0.35	AMS C-14	14630 ± 388	0.02 (14630-6469 yr BP)	
			0.52-0.53	AMS C-14	19586 ± 500	0.04 (19586-14630 yr BP)	
			0.68-0.69	AMS C-14	20817 ± 394	0.13 (20817-19586 yr BP)	
			0.84-0.85	AMS C-14	26128	0.03 (26128-20817 yr BP)	
			0.99-1.00	AMS C-14	28428	0.07 (28428-26128 yr BP)	
			1.20-1.22	AMS C-14	35581	0.03 (35581-28428 yr BP)	
			1.42-1.44	AMS C-14	41087	0.04 (41087-35581 yr BP)	
Off Cochin	GC-5	10.38°N; 75.57°E (280m)	045 - 0.50	AMS C-14	2330 ± 30	0.22 (2330 yr BP- present)	Thamban et al., 2001
			1.00 -1.05	AMS C-14	5620 ± 50	0.17 (5620-2330 yr BP)	
			1.45-1.50	AMS C-14	12860 ± 60	0.06 (12860-5620 yr BP)	
			1.95-2.00	AMS C-14	15230 ± 70	0.21 (15230-12860 yr BP)	
			2.45-2.50	AMS C-14	16200 ± 70	0.52 (16200-15230 yr BP)	
	AAS38-4	10°15.59'N, 75°39.5'E (239 m)	10-13	C-14	3030	0.0429 (3030 yr BP- present)	Present Study; Narayana et al., 2008
			48-50	C-14	4390	0.2721 (4390-3030 yr BP)	

			99-100	C-14	6760	0.2110 (6760-4390 yr BP)	
			148-150	C-14	10120	0.1488 (10120-6760 yr BP)	
			198-200	C-14	15650	0.0904 (15650-10120 yr BP)	
			248-250	C-14	18415	0.1808 (18415-15650 yr BP)	
			268-270	C-14	19445	0.1942 (19445-18415 yr BP)	
	AAS38-5	10°16.18'N, 75°32.2'E (875 m)	5-7	C-14	2100	0.033 (2100 yr BP- present)	Present Study; Narayana et al., 2008
			49-50	C-14	4360	0.1903 (4360-2100 yr BP)	
			74-75	C-14	5990	0.153 (5990-4360 yr BP)	
Off south of Cochin	SK177/12	8.73°N; 76.03°E	0.00-0.05	C-14	8690 ± 100	0.01 (8690 yr BP- present)	Pandarinath et al., 2003
		(332 m)	0.35-0.375	C-14	30500±1700	0.02 (30500-8690 yr BP)	
			0.50-0.525	C-14	41470±3400	0.01 (41470-30500 yr BP)	
			0.625-0.65	C-14	> 48000	-	
	VC-3125	8.24°N; 77.15°E (21 m)	1.07-1.17	C-14	8850 ± 140	0.12 (8850 yr BP- present)	Nambiar and Rajagopalan, 1995
	VC-3126	8.24°N; 77.16°E (22 m)	3.12-3.20	C-14	8420 ± 160	0.37 (8420 yr BP- present)	Nambiar and Rajagopalan, 1995
			3.40-3.50	C-14	8750 ± 130	0.88 (8750-8420 yr BP)	
	VC-3066	8.21°N; 77.17°E (21 m)	2.24-2.40	C-14	9390 ± 150	0.24 (9390 yr BP- present)	Nambiar and Rajagopalan, 1995

	SK177/11	8.2°N; 76.47°E (776m)	0.54-0.56	C-14	2885 ± 80	0.19 (2885 yr BP- present)	Pandarinath et al., 2003
			1.00-1.05	C-14	5950 ± 90	0.16 (5950-2885 yr BP)	
			1.50-1.55	C-14	8690 ± 100	0.18 (8690-5950 yr BP)	
			2.00-2.05	C-14	16040 ± 150	0.06 (16040-8690 yr BP)	
			2.50-2.55	C-14	18990 ± 200	0.17 (18990-16040 yr BP)	
			3.00-3.05	C-14	22900 ± 300	0.13 (22900-18990 yr BP)	
			3.60-3.65	C-14	28900 ± 300	0.10 (28900-22900 yr BP)	
Off south western tip of India	3101 G	~6°N; 74°E? (2680 m)	0.01-0.02	AMS C-14	1930 ± 100	0.01 (1930 yr BP - present)	Agnihotri et al., 2003
			0.11-0.12	AMS C-14	4685 ± 145	0.04 (4685-1930 yr BP)	
			0.26-0.27	AMS C-14	10614 ± 389	0.03 (10614-4685 yr BP)	
			0.38-0.39	AMS C-14	13491 ± 205	0.42 (13491-10614 yr BP)	
			0.53-0.54	AMS C-14	18320 ± 342	0.04 (18320-13491 yr BP)	
			0.71-0.72	AMS C-14	20564 ± 418	0.08 (20564-18320 yr BP)	
			0.87-0.88	AMS C-14	21715 ± 397	0.14 (21715-20564 yr BP)	
			1.02-1.04	AMS C-14	24515	0.06 (24515-21715 yr BP)	
			1.32-1.34	AMS C-14	28976	0.07 (28976-24515yr BP)	

depth of 280m) off Cochin (Thamban et al., 2001) was higher (~0.52mm/yr) during 16200-15200 yr BP when compared to ~0.06-0.21mm/yr during 15200 yr BP to present. South of Cochin, at 332m water depth, the bottom layer (0.625-0.650m) of a short core was found to be older (> 48000 yr BP) than the dating limit of the 14-C technique (Pandarinath et al., 2003). Consequently, the sedimentation rates at this core location are very low (~ 0.01-0.02mm/yr for the past 41500 yr BP) (Table 3.1).

Further south, on the inner continental shelf off Taingapatnam (south of Cochin), sedimentation rates of ~0.12mm/yr (~ 8850 yr BP to present), 0.37mm/yr (~ 8420 yr BP to present) and ~0.24 mm/yr (~9390 yr BP to present) were reported at water depths of 21, 22 and 21 m, respectively (Nambiar and Rajagopalan, 1995). At a deeper water depth (776m) off Taingapatnam, the sedimentation rate was relatively low and uniform (0.16-0.19 mm/yr) during the past 18990 years, except during the 16040-8690 yr BP where sedimentation was very low (~ 0.06mm/ yr) (Pandarinath et al., 2003).

Agnihotri et al. (2003), have recorded lower sedimentation rates (~0.01 to 0.04 mm/yr) during the Holocene (~ 10600 yr BP to present) than the pre-Holocene (~0.04 to 0.42mm/yr; ~29000-10600 yr BP) off the southern tip of India at a 2680m water depth. The sedimentation rate during the Pleistocene-Holocene transition (~13490-10610 yr BP) was significantly higher (~0.42mm/yr) than during ~28980-13490 yr BP (~0.04 -0.14mm/yr) and ~10610 yr BP to present (~ 0.01-0.03mm/yr) (Table 3.1).

Because of their small size, the sediment load of the rivers of southwestern India is small; and a part of this sediment load is being trapped/sequestered in coastal plains and extensive lagoons and estuaries. This is also perhaps one of the reasons for relatively low sediment accumulation on the adjacent shelf during Holocene.

3.4. Sediment texture

Distribution pattern of sediments in the continental shelf and slope regions mainly depend on the nature of sediment flux, hydrography and topography of the ocean floor. On the western continental shelf, the sediments are generally clastic siltyclays /clayey silts on the inner shelf (modern) and relict sandy sediment (deposited during early Holocene) on the outer shelf (Rao and Wagle, 1997).

Sediments on the continental slope are siltyclay in nature and are an admixture of dominant terrigenous and biogenic material (Rao and Rao, 1995).

Core AAS38-4

The sediment texture varies significantly at different layers of the core AAS 38-4. The core exhibits sandy mud, muddy sand and muddy texture (Fig. 3.3A).

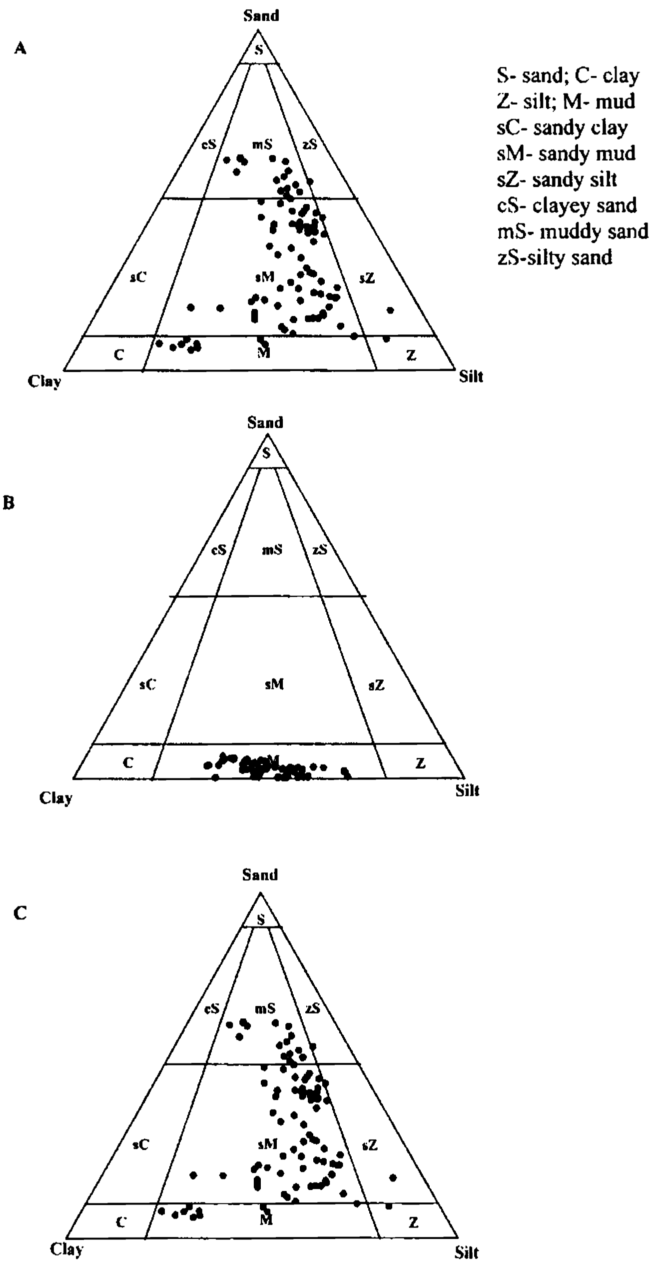

Figure. 3.3. Ternary diagrams showing the textural nomenclature of sediments in the cores (A) AAS 38-4 (B) AAS 38-5 (C) SK 145B/C-8

At this core site the sand content shows a large variation ranging from 6–62%with an average of32%. Sand content is comparatively low (<10 %) during 0-3.3 ka BP; from 3.4 ka BP onwards sand content shows an increasing trend, and a higher content (24-62 %) of sand is recorded during 3.6-11.9 ka BP. The maximum (~60 %) sand content has been recorded during the early Holocene. During 12.5-16.6 ka BP the sand content decreases considerably, ranging from 10 -15 %, whereas it increases up to 20% during 17-19.4 ka BP (Fig 3.4).

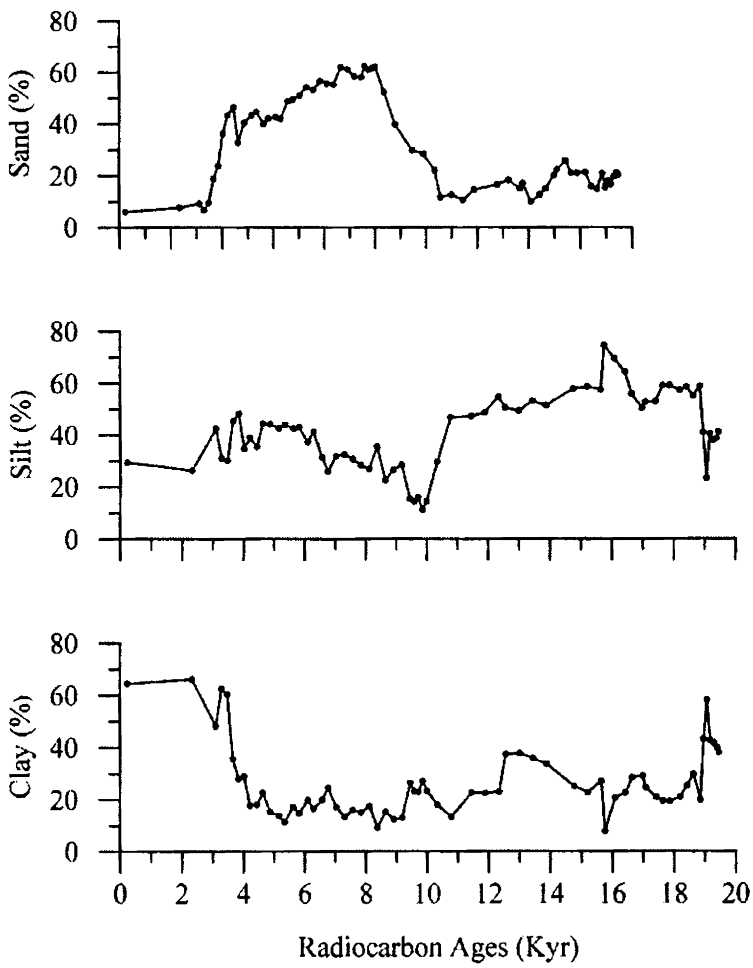

Figure.3.4.Temporal distribution of sand, silt and clay fractions in the core AAS 38-4

Silt content varies from 11-75 % with an average of 41 %. Silt is significantly higher through out the core length, except during 9.4–10 ka BP where the silt content is 11-16 %. During 0.2 – 9.1 ka BP. the silt content ranges from 26- 44 %; during 10.8 - 18.8 ka BP it generally ranges between 50-60%, except a few lower values of 40 % (Fig 3.4).

The clay content recorded in core AAS 38-4 varies from 10-63 %. In the top portion of the core i.e., 0.2-3.5 ka BP the clay content is about 60 %. Clay content ranges from 18-36 % during the period 3.6–4.6 ka BP., and 10-20 % during 4.8–9.1 ka BP. Clay content of 18- 27 % is recorded during 9.4–10.3 ka BP., and 23 – 38 % during 11.4–15.6 ka BP., with high (>35 %) clay content during 12.5-13.8 ka BP. The recorded clay fraction ranges from 20-30 % during 14.7–18.8 ka BP. The bottom of the core, representing 18.9 -19.4 ka BP period, records higher clay content ranging from 38-43 % (Fig 3.4).

Core AAS38-5

This core exhibits muddy texture sediments (Fig. 3.3B). Sand content is low (0-7 %) through out the core length. Sand content is more or less uniform and varies from 4-6% during 1–3.9 ka BP; it varies from 2-4 % during 4–7.7 ka BP and 0–2 % during 8–16.6 ka BP. Sediments of the core AAS 38-5 are mostly constituted by silt, followed by clay fraction. Silt content varies from 32–70% with an average of 49% (Fig 3.5).

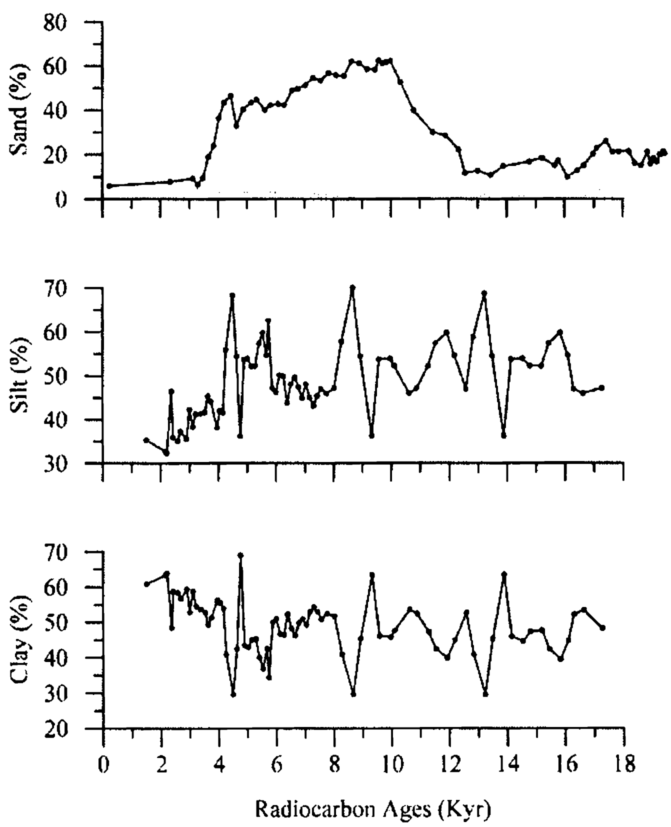

Figure.3.5.Temporal distribution of sand, silt and clay fractions in the core AAS 38-5

The core top, representing 0.2–2 ka BP, consists of less silt content (<35 %) with a decreasing trend during 2.3–5.7 ka BP it increases up to 68 %; and during 5.8–8 ka BP the silt content exhibits a decreasing trend. During 8.3–10 ka BP the silt content increases to 70%, whereas during 10.6–10.9 ka BP it again decreases to 46%. From11–16 ka BP silt content increases to 60%, where as during 16.3–17.3 ka BP silt content is slightly lesser, i.e., 46% (Fig 3.5)..

Clay content is also very high at the core site and varies from 30–64% with an average of 49%. The core top representing 0–2.2 ka BP period consists of high clay content (>61%). During 2.4–4.3 ka BP clay content is low (41%) and during 4.3 – 5.9 ka BP it is further low (30%), except at a few core intervals. High clay content (54%) is recorded during 5.9–8.01 ka BP, whereas during 8.3-10.1 ka BP it is low (30%). Clay content increases up to 54% during 10.6–0.8 ka BP. It varies considerably from 30–64% and exhibits increasing and decreasing trends during the period 12–16 ka BP (Fig. 3.5).

Core SK145B/C-8

Sediments at the core site of SK 145B/C-8 also exhibit muddy texture (Fig. 3.3C). The sand content is <1% through out the core. Silt content is very high varying from 44 to 78% with an average of 60%. The silt content is very high (75%) at the core top, representing the period 0–7 ka BP During 7.3–8 ka BP it shows decreasing trend, whereas during 8.3-11.5 ka BP higher content of silt (75%) is recorded. Silt content shows an increasing trend, except at a few layers, during 11.9–18.7 ka BP (Fig.3.6).

Clay content varies from 22–60% with an average of 40%. The core top representing 0–4.1 ka BP constitutes comparatively low clay content (~30%), whereas during 4.3–5.7 ka BP clay content increases (~40%). During 6–7 ka BP the clay content is low (~ 30%), but from 7.3 - 9.9 ka BP higher amount of clay is recorded (60%). Less amount of clay is recorded (<32%) during 10.2–11.5 ka BP. During 11.9–14.8 ka BP the clay content is high (55%) whereas during 15.1–18.7 ka BP it is low (~30%) (Fig.3.6).

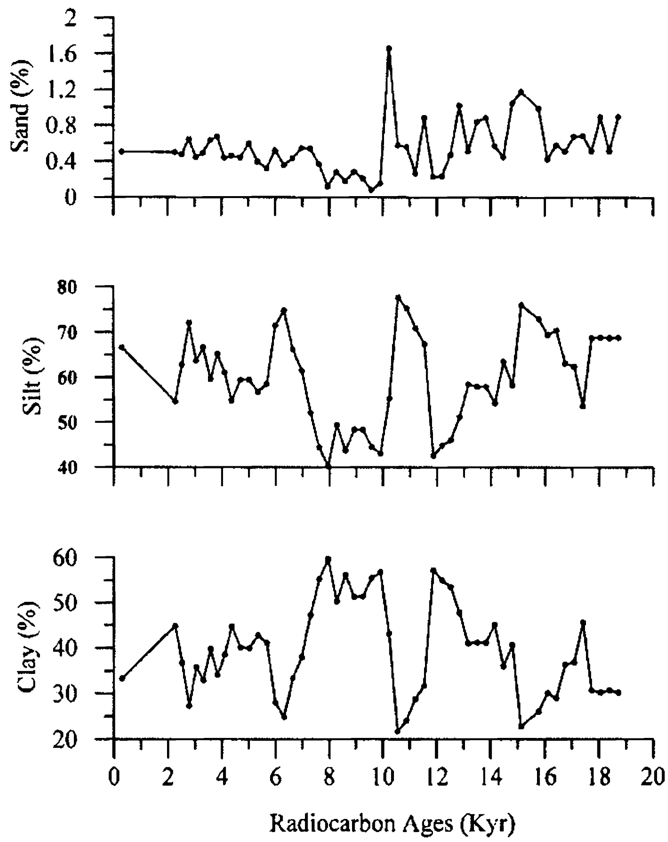

Figure.3.6. Temporal distribution of sand, silt and clay fractions in the core SK 145B/C-8

Textural Interrelationship of Cores

The core AAS 38-4, located in the upper slope, exhibits higher amount of sand compared to other cores located in the deeper water depths. Sand fraction is particularly higher (40-60%) during the middle to early Holocene (4-10 ka BP) on the upper slope region; correspondingly the silt content decreases followed by the clay content. The core site AAS 38-4 located relatively in shallow waters recorded gradual increase of silt with gradual decrease of clay during 2-4 ka BP, whereas during 4–6 ka BP higher silt and lower clay content are observed. During 6-9 ka BP slightly higher amount of silt and gradual increase and decrease of clay are recorded.

The cyclic variation of silt and clay, particularly in deep water core (AAS 38-5), is observed. The cyclic increase and decrease of silt and clay contents are distinctly observed at 2-4 ka BP, 4-6 ka BP, 6-8 ka BP and 10-14 ka BP.

The other deep water core (SK 145B/C-8) records high silt and low clay contents during 2-6 ka BP; it also records decrease in silt and increase in clay contents during this period. Silt shows further decreasing trend (75-40%) and clay an increasing trend (25 -55 %) during 6-8 ka BP. Lesser amount of silt (~40 %) and higher clay (50-60%) are observed during 8-10 ka BP. Silt decreases (75-45%) and clay increases (22-55%) during 10-12 ka BP. However, an increased silt (45-70%) and decreased clay (55-25%) contents were recorded during 12-15 ka BP. Again, during 15-17.5 ka BP decrease in silt and increase in clay are observed.

The western margin of India receives a large sediment input during SW monsoon (June-September) as it is associated with high rainfall. A southerly surface current, ~150km wide, in water depth of 50m on the shelf and an underwater current of about 40km wide in the depth interval between 100 and 250m prevail along the margin during the SW monsoon (Shetye et al., 1990). These currents play a significant role in transport of fine sediments from northern to southern region of the Indian coast.

The sea level variations may have also influenced the sediment texture and rates of deposition. For example, during the LGM the sea level was –120m, during 15-13 ka BP it was at –70m and at 12 ka BP the sea stood at –55m (Lambeck and Chappell, 2001) and the rivers must have been debouching approximately more than halfway between their present position and the core sites. This suggests that the rivers were debouching much far away than during the Holocene. Hence, the core site AAS 38-4 must have been receiving the coarse sediments and the higher sediment flux, and consequently higher sedimentation rates were recorded during this period.

The sediment textural patterns are controlled by hydrolysis, sediment load of rivers etc. The degree of hydrolysis on landmasses plays an important role in supply of sediments, particularly in a climatic regime dominated by seasonal precipitation, which favours both hydrolysis and hydromorphous conditions.

3.5. Organic carbon (C_{org})

Organic matter in marine sediments is a key in the global cycles of both carbon and oxygen (Berner, 1989), and provides a unique paleoenvironmental record. Understanding the factors controlling the distribution and preservation of organic matter in continental margin settings is of considerable importance, as continental margins represent important focal point for biogeochemical cycling. Further, the continental margins are known for high biological productivity and for the much of the ocean's carbon burial (McMannus et al., 2006).

The distribution of organic carbon content in all the three cores is represented in Figure 3.7. At the core site AAS 38-4, the organic carbon percentage varies from 0.75-2.95, it is 1.5% during 0-3.8 ka BP, and low (<1%) during 3-6.3 ka BP. Organic carbon content shows slight increase (~2%) during 6.5-19.4 ka BP, except in one or two sub-samples where slight decrease is observed (Fig. 3.7).

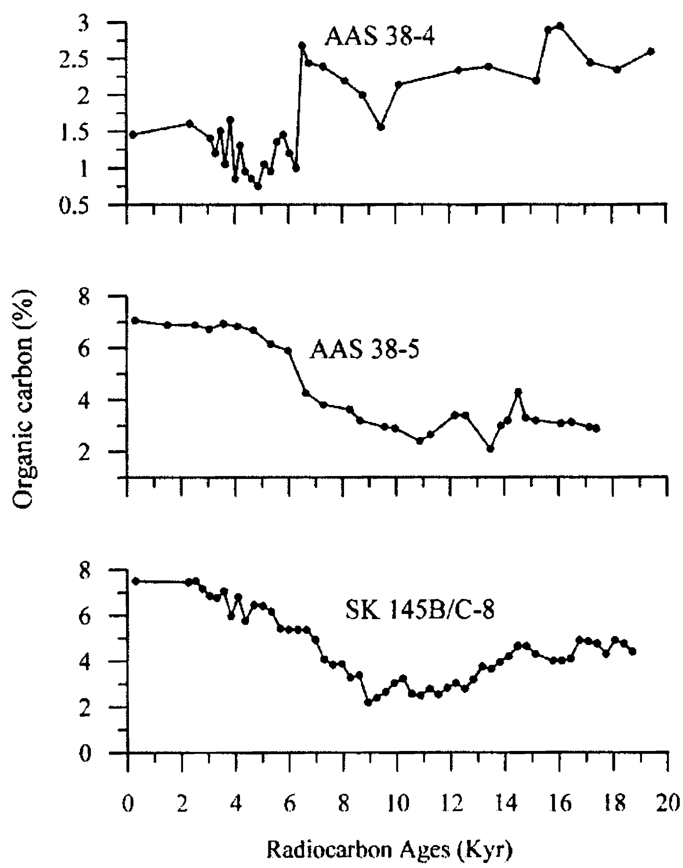

Figure.3.7. Temporal distribution of Corg in the core AAS 38-4, AAS 38-5 and SK 145B/C-8 respectively.

At the core site AAS 38-5 the organic carbon content varies significantly through out the core and ranges from 2 to 7%. Organic carbon is very high (~7%) in top portion of the core i.e., during 0–6 ka BP, whereas during 6-11 ka BP it is low (3-4%). It shows slight increase (~4%) during 11-12 ka BP and 14-18 ka BP (Fig. 3.7).

Organic carbon distribution in core SK 145B/C-8 is similar to that in core AAS 38-5. It is ~ 7% during 0-6 ka BP. It gradually decreases from 6-2% during 6-8 ka BP. Again it gradually increases from 2-4% during -14 ka BP, and remains around 4% during 14-18 ka BP.

Variation in organic carbon content may be attributed to changes in either productivity or preservation or both (Emerson, 1985; Emerson and Hedges, 1988; Hedges and Keil, 1995). Paropakari et al. (1993) argued that preservation of organic carbon is determined by the level of oxidation, respectively, the presence of an oxygen minimum zone in the Arabian Sea. Reichart et al. (1998) conclude that variations in organic carbon content in the northern Arabian Sea are not caused by fluctuations in the OMZ, and instead are primarily controlled by changes in surface water productivity. Various studies have mainly focused on the use of organic carbon in deep-sea sediments as a paleoproductivity index (Sarnthein et al., 1988; Lyle et al., 1988). Both production and oxygen availability can, however, not be independent of each other (Sarma, 2004). Another cause of changing organic carbon ratios could be dilution by inorganic components (Calvert et al., 1995).Increased sedimentation rate enhances organic carbon preservation in sediments (Heath et al., 1977; Sarnthein et al., 1988). Surface production and sedimentation rate are often positively correlated. However, increased sedimentation rates in the eastern Arabian Sea are not synchronous to increasing monsoon strength (Zahn and Pedersen, 1991). Phase-relationship between proxies of monsoonal forced bio-productivity and organic carbon accumulation rates along the western Arabian Sea imply that either bulk sedimentation rates impose a major influence on the preservation of organic carbon or that oceanic carbon production is not directly linked to monsoon-driven upwelling (Murray and Prell, 1992). A reliable organic carbon-based index of productivity must therefore be corrected for non-biogenic components. The higher sedimentary organic carbon on the continental slope off western India is widely believed to be due to the preferential preservation of deposited organic matter at water depths where the intense oxygen minimum intersects the sea floor. This region is considered to constitute one of the modern analogues for the environment of formation of organic-rich sedimentary facies that are common in the geological record. In addition to the high organic carbon concentrations in sediments currently accumulating in some oxygen minima, it is also widely accepted that the organic matter in these areas is better preserved compared with organic matter in sediments from more oxygenated environments (Demaison, 1991). Calvert (1987) has suggested that the locations of organic carbon maxima on continental slopes are not controlled by the bottom water oxygen levels but are probably produced by a combination of factors that control the texture of the sediments, the dilution of organic matter by other sedimentary components and the depth-related settling fluxes of organic carbon to the sea floor.

3.6. Calcium carbonate ($CaCO_3$)

The difference in CO_2 between glacial and interglacial periods documented in marine sediments and in ice cores and its links to changes in Earth's temperature (Neftel et al., 1982; Barnola et al., 1987) have got much attention among the paleoceanographers. Since the oceans cover about 70 % of the earth's surface, ocean–atmospheric exchange of CO_2 can take place over a vast area of the earth. Therefore, the oceans act as a major controlling device of atmospheric CO_2 through the chemistry of the oceans and preservation of calcium carbonate in deep –sea sediments (Berger, 1985). In order to understand past atmospheric CO_2 fluctuations it is necessary to know the dissolution and preservation patterns of calcium carbonate in the oceans. Fluctuations in $CaCO_3$ dissolution are largely controlled by changes in the global mass balance of $CaCO_3$ and ocean circulation. These processes, which actively promote the cycling of carbon among several reservoirs, serve as important controls of both oceanic carbon chemistry and atmospheric CO_2 content over orbital time scales (Broecker and Peng, 1987).

Arrhenius (1952) was the first to document the glacial/interglacial cyclicity of $CaCO_3$ in the equatorial Pacific. Since then it has been debated extensively whether the productivity of $CaCO_3$ secreting organisms or dissolution of $CaCO_3$ exert an influence on the $CaCO_3$ cycles. In addition, dilution by terrigenous material may also influence the $CaCO_3$ content. In general, the amplitude of carbonate cycles increases and the average values of $CaCO_3$ decrease in areas influenced by terrigenous dilution and/or dissolution in the Pacific (Arrhenius, 1952; Farrel and Prell, 1989; Berger 1992), Atlantic (Ruddiman, 1971; Damuth, 1975; Gardner, 1975) and Indian Ocean (Olausson, 1965; Naidu 1991; Naidu et al., 1993). Thus, dissolution and dilution are important factors in governing the fluctuations in $CaCO_3$ records.

In the Indian Ocean some sites exhibit the Pacific pattern, where $CaCO_3$ content is higher during glacials and lower during interglacials (Olausson, 1965; Oba, 1969; Naidu, 1991), whereas others show Atlantic patterns, where $CaCO_3$ content is generally higher during interglacials and lower during glacials (Peterson and Prell, 1985; Naidu et al., 1993), some sites show both Atlantic and Pacific patterns (Naidu and Malmgren, 1999). The changes in carbonate preservation was quite different between the Pacific and Atlantic Oceans, which caused the dichotomous fluctuations between these two oceans during glacial and interglacials (Volat et al., 1980). Recently Gupta et al. (2005) observed that in the eastern Arabian Sea $CaCO_3$ changes do not show a consistent pattern with reference to glacial and interglacial climatic fluctuations. Therefore, these authors suggest that $CaCO_3$ changes are controlled by regionally varying conditions of productivity and sedimentation than by global climate change.

The $CaCO_3$ data for two cores (AAS 38-4 & AAS 38-5) are presented in Figure. 3.8. As mentioned in the previous chapter, we could not analyse core SK 145B/C-8 for $CaCO_3$ determination. In core AAS 38-4 the calcium carbonate content vary from 11-57%. During 0-3 ka BP $CaCO_3$ is less (~30%), whereas during 3-6 ka BP it is high (40%). $CaCO_3$ is less (~32%) during 6-12 ka BP. Again during 12–13 ka BP, $CaCO_3$ increases to 21%. During 14–16 ka BP it further decreases and during 16-19.4 ka BP $CaCO_3$ gradually increases (Fig.3.8).

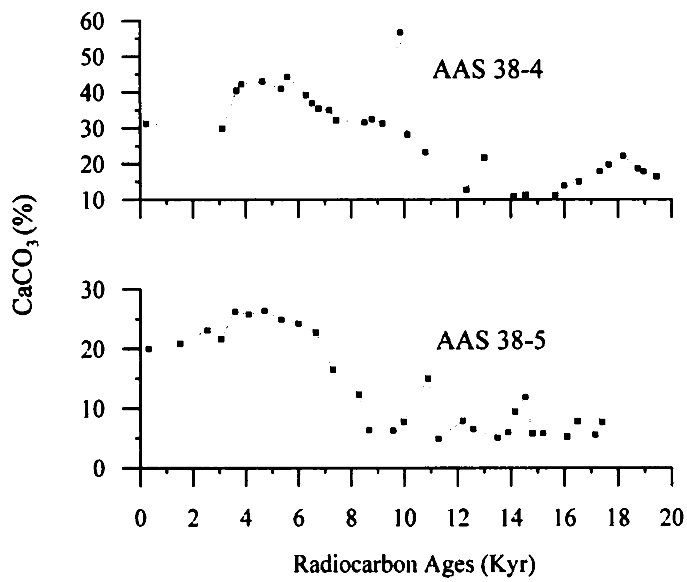

Figure.3.8. Temporal distribution of CaCO3 in the core AAS 38-4 and AAS 38-5

In the core AAS 38-5, calcium carbonate percentage varies from 6 - 26 %. $CaCO_3$ is high (25%) during 0-6 ka BP, whereas during 7.3-9.5 ka BP it decreases (16–6%). It again increases to 15% during 10-11 ka BP. A decrease (~6%) in $CaCO_3$ is recorded during 11-14 ka BP whereas during 14–14.5 ka BP a slight increase (~ 10%) is observed. Again during 14.7–17.4 ka BP a slight decrease (6 -7%) in $CaCO_3$ that remained around 6% was recorded (Fig.3.8).

3.7. Implications of paleoenvironment and paleomonsoon intensity

Rivers are the main source of sediments for the continental shelf and slope regions of the west coast of India. Hence, sediment accumulation reflects the intensity of the prevailing monsoonal and resulting weathering conditions along the adjacent hinterland. Temporal variations in sediment accumulation records can be used to reconstruct the sedimentary and paleomonsoonal conditions. Prins et al. (2000) observed higher sedimentation rates with higher continental humidity (summer-monsoon precipitation) in sediment cores on the Makran continental slope, Arabian Sea.

The sediments on the western continental shelf and slope of India are mainly terrigenous (Kolla et al., 1976; Nair et al., 1982; Rao and Rao, 1995), and sediment accumulation on the continental margin reflects the intensity of the prevailing monsoonal and resulting weathering conditions in the adjacent hinterland. Sediment accumulation rates calculated for different age intervals show significant temporal variations which could be attributed to the variations in sediment supply to the region as the result of fluctuations in paleomonsoonal conditions.

Sediment accumulation records along the western continental margin of India reveal high sedimentation rates during the Pleistocene-Holocene transition period as compared with immediately later and earlier time intervals. This shows a higher sediment input to the entire western continental margin of India during the Late Pleistocene-Early Holocene, implying higher rates of weathering and erosion at the hinterland due to high rainfall during this transition period. Based on other paleoclimatic proxy records in sediment cores, a more intense monsoon in the region during the Late Pleistocene-Early Holocene has been suggested by several workers (Van Campo et al., 1982; Van Campo, 1986; Prell et al., 1990; Naidu and Malmgren, 1996; Rajagopalan et al., 1997; Naidu, 1998). Mahiques et al. (2002) state that the input of terrigenous sediments to the continental shelf and slope off southeastern Brazil is controlled by the LGM and Holocene climatic changes.

In most of the earlier studied cores, sedimentation rates for the topmost dated layers are observed lower than those for deeper sections suggesting that the sediment input to the western continental margin of India has been comparatively low in recent years. The southwest coast of India characterized by such landforms as lagoons, barrier islands, beach ridges, paleo-strandlines, coastal alluvial plains, marshy plains and flood plains, acts as a major sink for river derived sediments (Narayana, 2006).

The variation in supply of coarser material to the marine environment is directly related to the intensity of the monsoon. High and low contents of sand fraction reflect the strong and weak monsoon conditions respectively on the adjacent hinterland. The down-core variation in silt and clay contents in the inner continental slope also indicates similar monsoonal signal. Low supply of sand content from 19.5 ka BP to 12.6 ka BP and sudden increase since 10.3 ka BP in the upper continental slope region (239 m depth), and steep increase in sand content during 15.9 - 10.7 ka BP at the site of the core AAS-38/5 (875 m water depth) indicate varied hydrolysis. The enhanced hydrolysis on land tends to be enriched in coarse fraction. These observations corroborate with the high sedimentation rates and strong monsoon signals.

Sediment trap studies have shown high biological productivity and more terrigenous supply in to the Arabian Sea during southwest monsoon (Haake et al. 1993). Earlier studies have outlined the monsoonal influence on the biological productivity and terrigenous supply in the Arabian Sea during the Late Quaternary (Sirocko and Sarnthein 1989; Shimmield et al. 1990; Clemens et al. 1991; Murray and Prell, 1992; Naidu, 1991; Naidu et al., 1993; Naidu and Shankar, 1999; Bhushan et al., 2001; Agnihotri et al., 2003; Gupta et al., 2005). It is generally understood that the summer monsoon was stronger during interglacials than in glacials (Prell et al., 1992). Productivity during the last glacial maximum (LGM) was observed to have decreased in the western Arabian Sea (Emeis et al., 1995; Spaulding and Oba, 1992) and in southeastern Arabian Sea (Pattan et al., 2003; Ivanova et al., 2003), where as the northern and eastern regions of the basin exhibited increased LGM -productivity (Cayre and Bard, 1999; Rostek et al., 1997; Schulte et al., 1999; Thamban et al., 2001;

Banakar et al., 2005). The decreased LGM-productivity is linked to the weakened summer monsoon (Anderson and Prell, 1993) leading to reduced upwelling (Naidu and Malmgren, 1996), whereas the increased LGM-productivity is linked to the intensification of winter monsoons (Duplessy, 1982; Rostek et al., 1997) leading to an increased deep-water nutrient injection in to the photic zone (Cayre and Bard, 1999).

Organic carbon has been used as a proxy for paleoproductivity in the Arabian Sea (Fontugne and Duplessy, 1986; Thamban et al., 2001; Banakar et al., 2005). Accordingly, it was suggested that the productivity of eastern Arabian Sea was higher during LGM than in Holocene (Rostek et al., 1997; Schulte et al., 1999; Beaufort et al., 1999; Prabhu and Shankar, 2005; Banakar et al., 2005). These authors have attributed high productivity during glacials due to the strong northeast monsoon winds, which was responsible for bringing nutrient rich bottom water to the euphotic zone by breaking the water column stratification, thus inducing high primary productivity. By contrast, more recently Gupta et al. (2005) have demonstrated that organic carbon changes during glacials and interglacials were not consistent in the eastern Arabian Sea, therefore one would be cautious in interpreting the organic carbon record in terms of productivity. The present study also confirms that the organic carbon variations at the core sites are highly inconsistent. AAS38-4 shows higher organic carbon values during LGM and lower values during the Holocene. On the contrary, AAS38-5 document lower and higher values during LGM and Holocene, respectively. Therefore, it is very unlikely that the organic carbon concentrations in these two cores provide a reliable proxy for productivity along the eastern Arabian Sea.

Calcium carbonate and organic carbon content in the cores presently studied are largely different; however, show similar trend in $CaCO_3$ variations, whereas organic carbon changes show opposite trend during Holocene and LGM. Strikingly, AAS 38-5 core raised from deeper depth shows lower $CaCO_3$ values than in the core AAS 38-4 that was raised from shallower water depth and more proximal to the present day shoreline. Further, both cores exhibit comparatively low $CaCO_3$ values during LGM and high values during Holocene. The difference in $CaCO_3$ concentrations between the two cores appears to have been caused by the dilution with terrigenous material. The dilution is greater in AAS38-5 as compared to the AAS 38-4. This finding signifies that the terrigenous material supply is higher at the location of AAS 38-5. High Al and Ti values in the core AAS 38-5 (see Chapter. 5) also lend support to the above argument suggesting greater terrigenous dilution at this core site. Therefore, variation in $CaCO_3$ content is primarily due to dilution by terrigenous material supply to the core sites. Water depth at both the core locations are much above the foraminiferal lysocline depth in the Arabian Sea (Cullen and Prell, 1984), therefore, dissolution factor of $CaCO_3$ may be negligible in this region.

The input of terrigenous material to the Arabian Sea varies on a regional scale and relates to monsoon intensity (Sirocko, 1989). Higher linear sedimentation rates during Holocene at the core sites 38-4 and 38-5 correlate with high concentrations of $CaCO_3$, possibly caused by two factors; i) less terrigeneous dilution and, ii) enhanced productivity. Sea level during LGM was -120m below the present day level and the rivers were debouching into the sea far away from the present coast. Further, the lower sea level during LGM may have favoured the erosion of then exposed continental shelf and shortened the pathway for the transport of suspended load of rivers to the sea at these core locations. Hence, the concentrations of lithogenous materials were higher resulting low $CaCO_3$ content in the sediment.

It has been suggested earlier that the $CaCO_3$ records from the Arabian Sea are mostly influenced by terrigenous dilution, caused by variations in the terrigenous lithogenic flux derived from the Arabian and Somalian Peninsulas during the summer monsoon season (Murray and Prell, 1992), and by riverine material from the Indus River in the eastern Arabian Sea (Naidu, 1991). The similar dilution of $CaCO_3$ in the southeastern Arabian Sea is caused by the terrigenous lithogenic flux derived from the southwestern part of India and exposed continental shelf during the low sea levels i.e., the LGM.

Flux of lithogenic matter, bulk carbonate, organic carbon, and foraminiferal tests is tightly coupled to monsoon intensity in the western, central, and eastern Arabian Sea, as shown by sediment trap studies (Haake et al., 1993). A distinct bi-modal distribution, representing high fluxes during summer and winter monsoons and minimum flux during inter-monsoon (Nair et al., 1989; Curry et al., 1992; Rixen et al., 1996), implies that the biological productivity in the modern Arabian Sea is higher during monsoons and lowest during inter-monsoons.

3.8. Summary

Sediment core exhibit varied textural facies. On the upper slope region the sediments are of muddy sands to sandy muds and the sand content is higher during the Holocene. In the deeper slope region, sediments are of clayey silts. Silt content is dominant in the Late to Middle Holocene and during last glacial period. The sediment supply to the core sites must have been influenced by the monsoon variability, transport pathways and sea level changes. The sedimentation rates reveal that sedimentation was low since 2 ka BP, where as during the mid-Holocene the sedimentation rate was significant. The low sedimentation rate was recorded in the deeper slope core sites also since last 2 ka BP. Sedimentation rates decrease with increasing water depths. The sedimentation rates in the southeastern Arabian Sea are controlled by hydrolysis and sediment load of rivers. Higher sedimentation rates during late Pleistocene – Early Holocene are associated with high rainfall and consequent weathering and erosion.

Increased sedimentation rates in the eastern Arabian Sea are not synchronous to increasing monsoon strength (Zahn and Pedersen, 1991). The observed sedimentation pattern may suggest the hinterland of southwest India as the source for sediments to the region. The non- consistent trend of sedimentation rates may be the result of prevailing adverse depositional conditions. It is reported that the bottom topography, hydrography, water currents and gravitational redeposition control the sediment dispersion on the Indian western continental slope (vonstackelberg, 1972 and Sirocko and Lange, 1991). It is also possible that the observed low sedimentation rates at shallow water depth may be an isolated case since we do not have the necessary detailed hydrographic and seismic survey data of the study area to provide definite explanation on the prevailing depositional conditions. However the present data clearly suggests that the generally presumed decreasing sedimentation rates with increasing water depths is not always true on the southwestern continental slope of India.

High organic carbon since last glacial period to Mid-Holocene and low organic content since 6 ka BP are recorded. High organic carbon in the present study area is controlled by productivity and sediment texture. The upper slope core site recorded higher organic carbon during LGM and lower during the Holocene. At the deeper water core site, low organic carbon content during last glacial period and high during the Holocene. Therefore, there is a limitation in using organic carbon as a productivity indicator.

Calcium carbonate values are lower in deeper water core than in shallow water core. The difference in $CaCO_3$ values appear to have been caused by the dilution with terrigenous material. The dilution is greater in core AAS 38-5 than in AAS 38-4. The dilution of calcium carbonate is caused by the terrigenous, lithogenous flux derived from the hinterland and exposed continental shelf during the lower sea levels.

CHAPTER-IV

Clay Mineral Records of Sediment Cores: Provenance and Paleomonsoon

4.1. Introduction

Clay minerals are the weathering products of rocks and soils and their composition largely depends on climate, geology and topography of the area. Once formed in weathering profiles, clay minerals are eroded, carried away by transporting agents such as river, glacier and wind, and accumulate in sedimentary basins on land or in marine environment.

Clay minerals represent the most ubiquitous components of sediments, fluvial to marine environments. The production, supply and composition of clay minerals in the marine environment largely depend on geology and drainage of the hinterland area, and climate variations (Weaver, 1989; Chamley, 1989). The nature and composition of clay minerals will provide valuable information regarding the type and intensity of weathering on land. It has been demonstrated that clay mineral assemblages in marine sediments are particularly informative as to sources of the sediments, especially in relation to the integral effect of provenance, lithology and climate (Naidu and Mowatt, 1983; Chamley, 1989). Several studies have suggested that the temporal variation of clay minerals in marine sediments can be utilized as paleoclimatic proxies (Sirocko and Lange, 1991; Gingele, 1996; Thamban et al., 2002). Clay mineral distribution patterns in oceans can offer potentially useful proxy data on long term dispersal of water masses and associated pollutants.

Distribution of clay minerals in oceans is complicated by various processes viz., circulation patterns, supply of sediment from multiple sources, size sorting, flocculation and organo-mineral interactions (Grim, 1968; Gibbs, 1977; Kolla et al., 1981). In the marine environment, where the terrestrial input is dominant, the down core variation in sediment characteristics may reveal changes in the intensity of weathering or variations in depositional conditions depending on climatic conditions and/or sea level changes (Kolla,1981). Therefore, the study of clay minerals is considered to be a reliable tool for the reconstruction of paleoenvironmental conditions in the region.

Numerous studies have indicated the following: (a) Clay minerals in the marine environment are largely detrital (Biscaye, 1965); (b) There are no diagenetic effects in recent marine clays and therefore the provenance of different clay minerals can be identified (Grim, 1968); (c) Crystallinity of particular clay minerals can also be used as an evidence of climatic change (Jacobs and Hays, 1972); (d) As rainfall is the most controlling factor determining the leaching rates in weathering profiles, the composition of clay minerals in the sediments are useful indicators of paleoclimatic conditions (Singer, 1984).

Most oceanic basins reflect the existence of various controls on the distribution of terrigenous clays. The climate, land petrography, and near shore hydro dynamic constraints intervenes dominantly in turn, depending on the location of the different parts of a given basin. In addition the long distance transportation processes by marine currents and by the wind may strongly modify the distribution of detrital assemblages with respect to the terrestrial production zone. Some *in situ* formations of clay and associated species occur in certain basins, and complicate the final distribution of mineral suites on the sea floor.

The clay mineral distribution in the western Indian Ocean is largely influenced by the climate and the land geology, but is also controlled by physiographic patterns, sub-marine volcanism and marine currents (Kolla et al., 1976). Quartz- smectite rich clays occur along the Indian margin, south of the Indus River; they result from the climate, weathering of Deccan basalts and are associated with quartz derived from Pre-Cambrian metamorphic rocks, and appear to be primarily dispersed southerly, and to some extent northerly, by surface currents.

Detrital fine grained sediments are abundant both on the shelf and continental slope of the west coast of India. They occur largely as clay minerals, weathered from the rocks of the hinterland and are primarily transported by rivers. The parent rocks in the drainage basins of the rivers have, however been extensively lateritised. Rivers are the principal agents of transport of detrital sediments in to the eastern Arabian Sea. The Indus River is the largest one, bringing enormous amounts of sediment (Haq and Milliman, 1984). Clay particles supplied from land to sea often experience further transportation through long shore and density currents or through reworking processes. The variable influence of long shore currents that strongly depend on meteorologic and seasonal conditions also interferes with the slow and progressive settling phenomena.

Studies on clay mineral composition in the western continental margin of India were mainly focused on the surficial sediments to understand the provenance and transport path ways of fine grained land derived sediments (Kolla et al., 1981; Nair et al., 1982; Rao et al., 1983; Narayana and Pandarinath, 1991; Rao, 1991; Rao and Rao, 1995; Chauhan and Gujar, 1996) and studies on sediment cores were few (Sirocko and Lange, 1991; Chauhan et al., 2000; Thamban et al., 2002, Kessarkar et al., 2003). The clay size fraction of the Arabian Sea lithogenic sediments is derived from a mixture of river run-off from India and eolian dust contributions from Arabia, Pakistan and northern India (Kolla et al., 1981; Naidu et al., 1985). The maximum of clay sedimentation off southwest coast of India is corroborated by both concentration values and accumulation rates and can be attributed to high abundances of fine grained river borne sediments along the southern Indian coast (Naidu et al., 1985). Most of the clay minerals are ubiquitous in the Arabian Sea and could not be related to specific source areas and transport pathways, but the large amounts of smectite off southern India are clearly derived from Indian rivers (Sirocko and Lange, 1991). Clay minerals consist of hydrous layer silicates that constitute a large part of the family of phyllosilicates.

The fine grained clastic fractions, mainly of clay minerals, in the marine environment are the weathering products of rocks and soils on land. Several studies have suggested that the temporal variations of clay minerals in marine sediments can be utilized as paleoclimatic proxies, provided that they are detrital and have not been subjected to alteration by diagenesis (Sirocko and Lange, 1991; Gingele, 1996; Vanderaveroet et al., 1999). Diagenetic modifications of the detrital clay minerals in marine environment during the recent past are considered to be negligible (Grim, 1968; Chamley, 1997).

Previous studies on clay mineral composition of sediments from the eastern arbian Sea were mainly based on surficial sediments to understand the provenance and transport pathways of fine grained terrigenous sediments (Kolla et al., 1981; Nair et al., 1982; Konta, 1985; Rao, 1991; Chauhan, 1994; Rao and Rao, 1995). The studies on clay mineralogy of sediment cores in the Arabian Sea and their utility as paleoclimatic proxies are very few (Kolla et al., 1981; Sirocko et al., 1993; Thamban et al., 2002; Kessarkar et al., 2003).

In the present study, clay mineral distribution patterns since Last Glacial Maximum (LGM) in three gravity cores are discussed. The ratios of clay mineral assemblages and their relative abundance were utilized to interpret the plaeoenvironment and paleomonsoon record. The main focus of this chapter is to understand the provenance and to infer paleoclimate and paleomonsoon during late Quaternary.

4.2. Clay mineral abundance

The clay mineral types and the proportions of the individual clay minerals in marine sediments, therefore, depend on the climatic conditions on land and on the nature of the source rocks. The distribution of different clay minerals in the present-day oceans reveals a zonation that strongly reflects the pedogenic zonation and climatic conditions on the adjacent continental land masses (Biscaye, 1965; Griffin et al., 1968; Lisitzin, 1972; Windom, 1976). Clay mineral assemblages in marine sedimentary sequences are, therefore, useful tools for reconstructing the paleoclimate through time. Clay minerals are also being useful for deciphering and reconstructing the sedimentary processes.

Smectite is derived from chemical weathering of parent alumino silicates and ferromagnesian silicates under warm and humid conditions, and also from chemical weathering of basaltic rocks. Kaoilinite is readily found in soils of inter tropical land masses charecterised by a warm, humid climate, and therefore displays a strong climatic dependence controlled by the intensity of continental hydrolysis (Chamley, 1989). Kaolinite is common on steep slopes within the drainage basin, where there are good drainage conditions. The detrital chlorite mainly results from the chemical weathering of plutonic and metamorphic rocks. The presence of illite and chlorite which reflect the decrease of hydrolytic processes in continental weathering and an increase of direct rock erosion under cold and arid climatic conditions. In addition, illite could also be formed by the weathering of non-layer silicate, such as feldspar from granites under moderate hydrolysis conditions, and by the degradation of micas.

Smectite is considered as settling preferentially in areas of decreased current and associated grain sorting. The variable influence of long shore currents that strongly depend on meterologic and seasonal conditions also interferes with the slow and progressive settling phenomena. The ability of smectite minerals to settle preferentially allows identification of current activity (Chamley, 1989, p.120). The relative abundance of smectite displays a distribution that does not parallel the zonal distribution of main weathering processes. This indicates the accessory control of climate, and the dominance of other allochthonous and /or autochthonous processes. The increased amounts of marine smectite recorded off the temperate to sub-arid regions suggest that the mineral partly reflects conditions intermediate between those of cold-dry and warm-humid climate.

The distribution of kaolinite in marine sediment reflects warm, humid climate control and, hence kaolinite is called "low latitude mineral" (Griffin et al., 1968). Kaolinite abundance increases towards the equatorial regions in all ocean basins, and therefore expresses a strong climatic dependence controlled by the intensity of the continental hydrolysis. Marine kaolinite derived from inter tropical soils is generally associated with abundant iron oxides (mainly goethite and sub-amorphous components), and often with gibbsite (Biscaye, 1965). A decrease in kaolinite abundance with increased distance from coast in Indian Ocean sediments is envisaged as conditions of marine transportation (Gorbunova, 1962).

The distribution of chlorite in marine sediments typically reflects high latitude climatic conditions. The relative abundance of illite tends to increase towards high latitudes parallel to chlorite, which reflect the decrease of hydrolytic processes and the increase of direct rock erosion under cold climatic conditions. In addition, abundant detrital illite characterizes the oceanic areas that are linked to either to high altitude cold climate regions like the Himalayas or to desert climate regions.

A detailed study in the Arabian Sea by Kolla et al. (1981) had lead to identify the role of climate, petrographic sources, and currents on the distribution of clay minerals and quartz. Smectite rich clays occur along the Indian margin south of the Indus river; they result from the chemical weathering of Deccan basalt traps, are associated with quartz issued from local Pre-Cambrian metamorphic rocks, and appear to be primarily transported southerly by surface currents. Illite rich clays dominate in most of the rest of the Arabian Sea.

The transportation agents appear to consist of surface currents and turbidity currents off the Indus river area, and mainly of winds in other areas. The effect of arid to desert climate in the northern and western areas favours the dominant control of land petrography on detrital associations as well as the essential action of wind. In the southern and southeastern region, kaolinite rich sediments probably derive from inter tropical soils of Africa, Madagascar and southern India.

Based on the studies of cores from the eastern Arabian Sea, Aoki and Sudo (1973) reported an increase in relative abundance of smectite and kaolinite during interglacial stages, while glacial stage sediments are relatively enriched in illite and chlorite; these clay mineral variations are attributed to changes in the degree of continental hydrolysis, interglacial stages favouring weathering processes on Asian land masses. Under tropical humid latitudes where kaolinite forms abundantly in well-drained soils during both glacial and interglacial periods, the variations recorded in the composition of terrigenous clay input may reflect temporary changes in the intensity and geographic location of rainfall.

The clay minerals present in the sediment cores of the study area in the decreasing order of abundance are - smectite, kaolinite, illite and chlorite. X-ray Diffractograms of some selected samples are shown in Figures 4.1–4.3. The relative abundance of clay minerals varied significantly with in the cores and among the cores and is described in detail in the upcoming section.

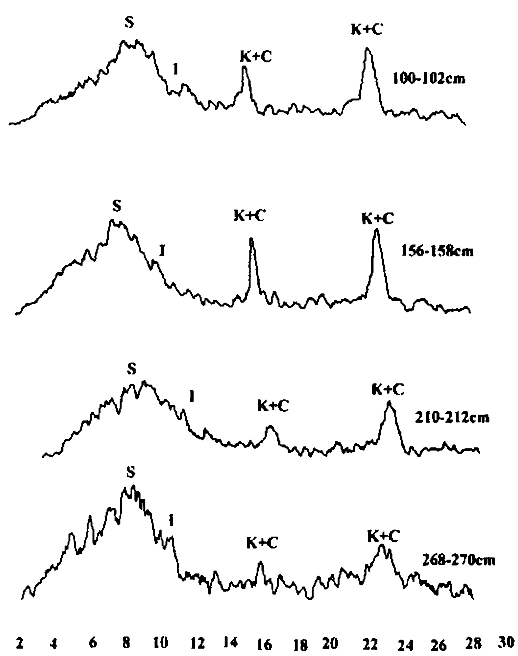

Figure 4.1. X- ray diffractograms of some samples at different core depths of the core AAS38-4.

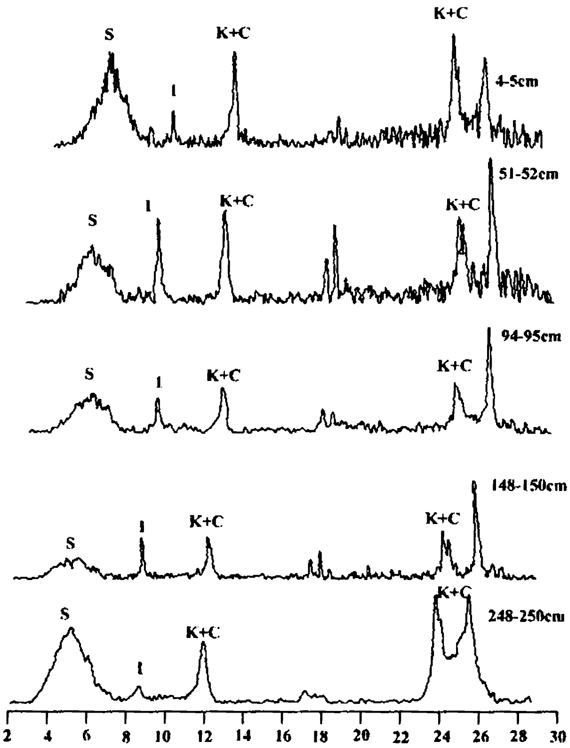

Figure. 4.2. X- ray diffractograms of some samples at different core depths of the core AAS38-5.

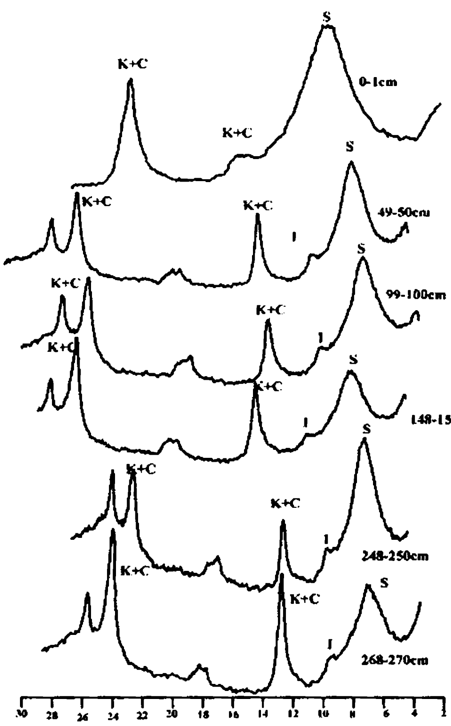

Figure. 4.3. X- ray diffractograms of some samples at different core depths of the core SK145B/C-8.

Core AAS 38-4

Smectite, illite, kaolinite and chlorite are the dominant clay minerals in the decreasing order of abundance on the upper slope region, where the core AAS 38-4 is located. Smectite varies from 23–68 % with an average of 52%. Illite content varies from 10–32 % with an average of 21%. Kaolinite percentage varies from 10–26 % with an average of 18 %. Chlorite percentage varies from 4–15% with an average of 8 % (Fig. 4.4).

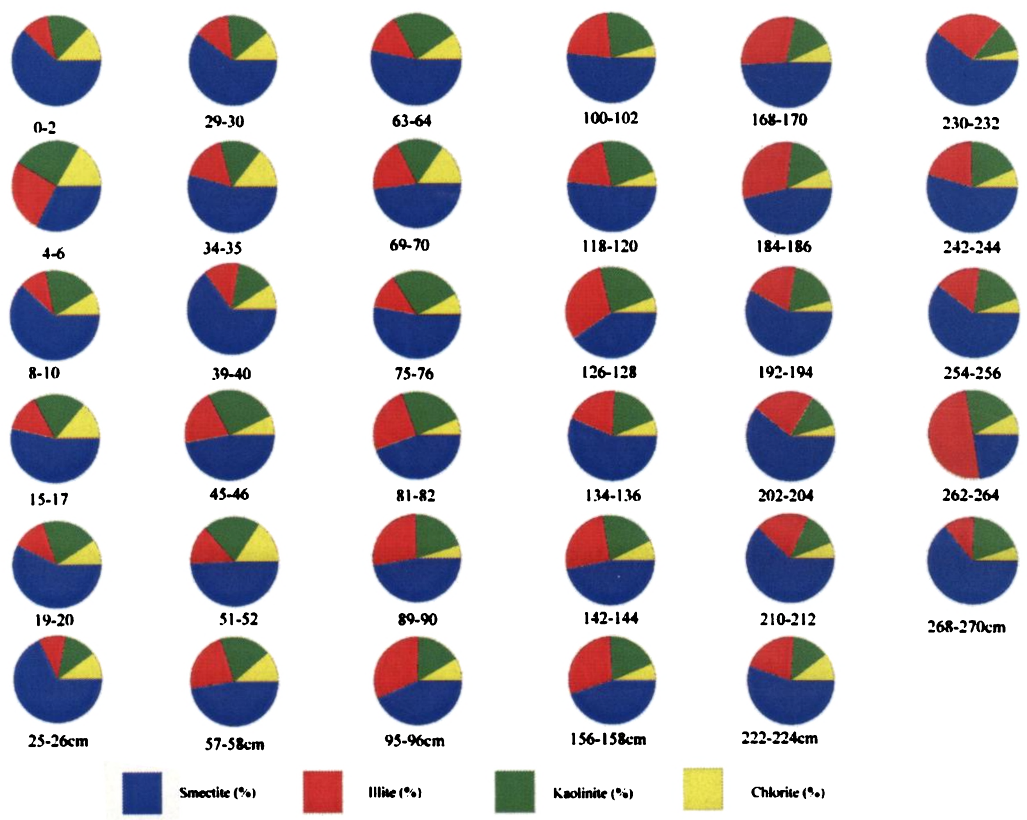

Figure. 4.4. Pie diagram showing the relative abundance of clay minerals - smectite, illite, kaolinite and chlorite -at different core lengths in the core AAS 38-4.

Core AAS 38-5

The abundance and distribution of clay minerals in core AAS 38-5 (located in deep waters) is similar to that of core AAS 38-4. Smectite, illite, kaolinite and chlorite are the dominant clay minerals in the decreasing order of abundance in the deeper slope region i.e., at the core site AAS 38-5. Smectite percentage varies from 33 to 70% with an average of 55%. Illite content ranges between 8 and 28% with an average of 18%. Kaolinite content varies from 10–26 % with an average of 17%. Chlorite content ranges from 5 to 20% with an average of 11% (Fig. 4.5).

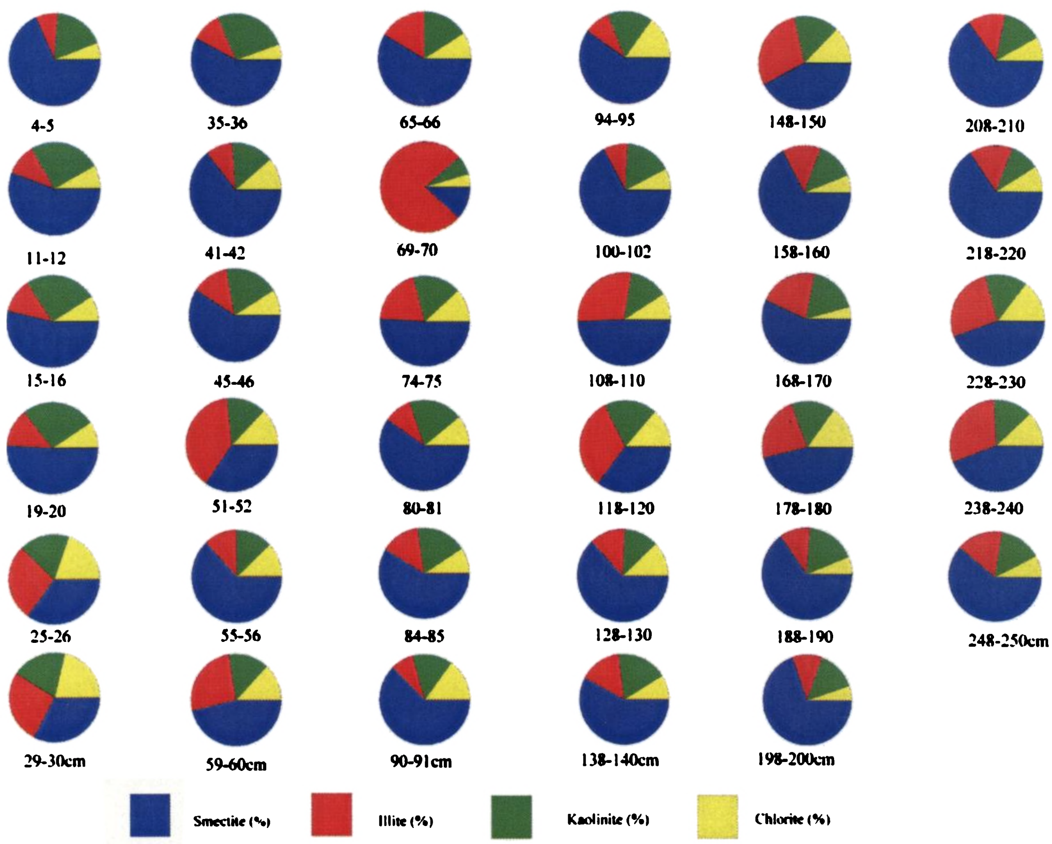

Figure.4.5. Pie diagram showing the relative abundance of clay minerals - smectite, illite, kaolinite and chlorite -at different core lengths in the core AAS 38-5.

Core SK 145B/C-8

In core SK 145B/C-8, although smectite is the dominant mineral, the order of abundance of other clay minerals varies. Smectite is followed by kaolinite and illite in abundance, which is contrary to clay mineral variations of core AAS 38-4 and AAS 38-5. In this core the second abundant clay mineral is kaolinite whereas in other two cores illite was second abundant. As observed in other two cores, in this core also very high smectite content is recorded and it ranges from 29 -73 % with an average of 59 %. Kaolinite varies from 7 -47 % with an average of 20 %; the relative abundance of illite varies from 7-29 % with an average of 18% (Fig. 4.6).

It is interesting to note that chlorite is almost absent through out the core length, except at a few randomly distributed horizons. In top, middle and bottom portion of the core chlorite is almost absent.

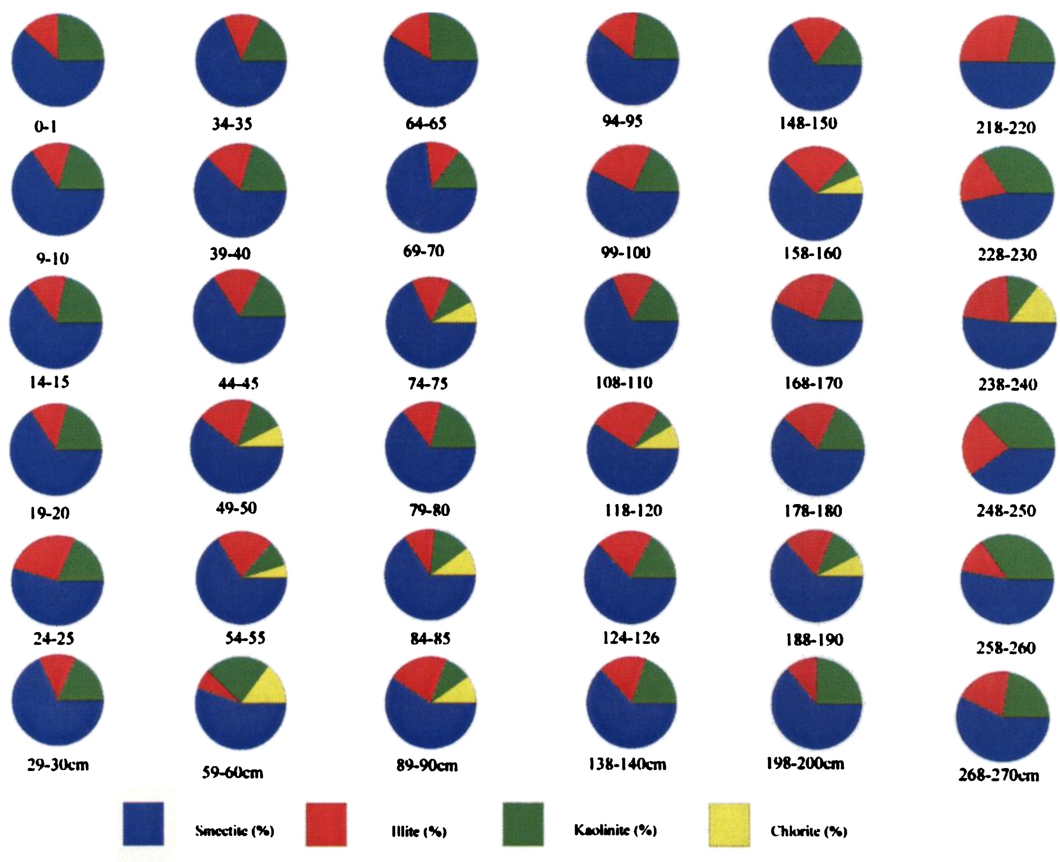

Figure.4.6. Pie diagram showing the relative abundance of clay minerals - smectite, illite, kaolinite and chlorite - at different core lengths in the core K145B/C-8.

4.3. Temporal distribution of clay minerals

Core AAS 38-4

At the core site AAS 38-4, smectite content is more (~ 60 %) during 0-4 ka BP, whereas during 4.2-6.6 ka BP it decreased to 47 % (Fig. 4.7). During 6.8-9.2 ka BP an increase of (~52 %) in smectite content is recorded, and again decreased to ~ 45% during 9.7 –14 ka BP. An increase (to 60 %) in smectite content is recorded during 15-19.4 ka BP (Fig.4.7). Illite content is ~10 % and ~15 % during 0-3.5 ka BP and 3.5-5.6 ka BP respectively. Illite increases in relative abundance up to 31% during 5.9-14 ka BP, whereas gradual decrease (~ 11 %) is observed during 15-19.4 ka BP (Fig.4.7).

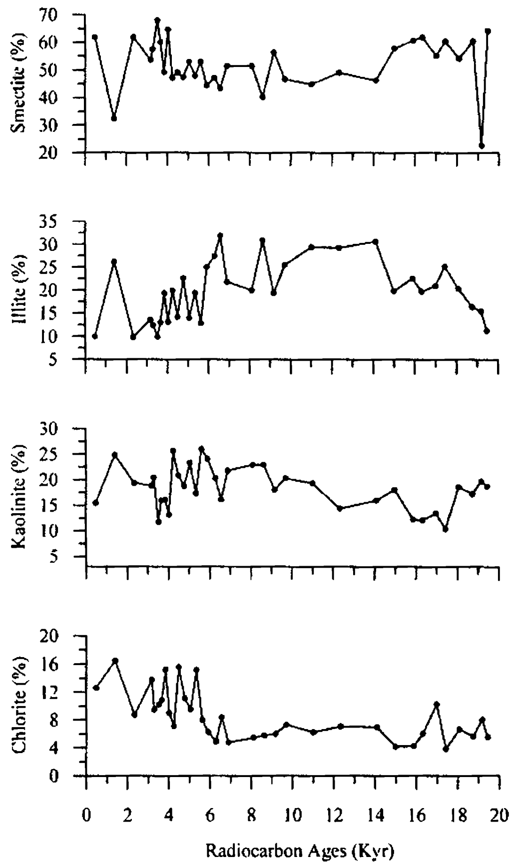

Figure.4.7. Temporal distribution of smectite, illite, kaolinite and chlorite in the core AAS 38-4.

The relative abundance of kaolinite is higher (~20 %) during 0-3.3 ka BP and it is about 15 % during 3.5 - 4 ka BP. During 4.2 - 15 ka BP kaolinite content is about 23 %, whereas during 16 – 17.4 ka BP it decreases to ~12 %. Again kaolinite abundance increases to 20 % during 18 - 19.4 ka BP (Fig.4.7). Chlorite content is more (~12 %) whereas during 0-5.3 ka BP and is 8 % during 5.6-16.3 ka BP. During 17-19.4 ka BP, chlorite content further decreases to 6 % (Fig. 4.7).

Core AAS 38-5

In core AAS 38-5 also smectite is the dominant one followed by illite, kaolinite and chlorite (Fig. 4.8). During 0-2.8 ka BP the smectite content is more (~ 55 %), whereas during 3.1-3.3 ka BP it is less (~ 30 %). The abundance of smectite is very high (~65%) during 3.6-8 ka BP and during 8-13 ka BP it decreased to about 55%. During 13. 5–15.5 ka BP smectite increases upto 70% and during 16–16.7 ka

BP it decreased to 44 %. An increase of smectite to 60 % during the period 16.7-17.4 ka BP is recorded (Fig. 4.8).

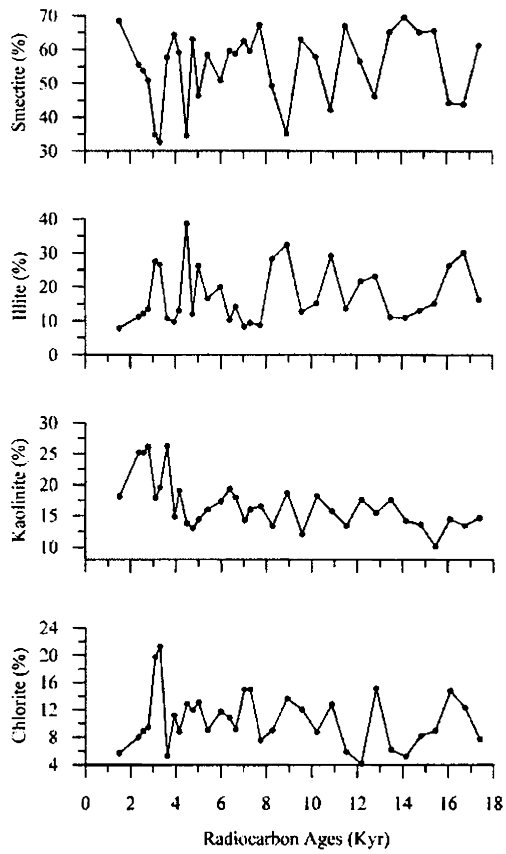

Figure.4.8. Temporal distribution of smectite, illite, kaolinite and chlorite in the core AAS38-5.

During 0-3.3 ka BP, illite content progressively increases from 8 to 27%, while it decreases to 10 % during 3.6– 4 ka BP. At 4.5 ka BP the illite content shows a rapid increase upto 39 %. This might be an outlier. A gradual decrease (from 20-8 %) in illite content is observed during 4.8-7.8 ka BP, while 32 % of illite is recorded during 8-9 ka BP. During 13.5-16.5 ka BP, illite content is about 10%, and it gradually inreases to 30 % at 17 ka BP (Fig. 4.8).

Kaolinite content is slightly higher (22 %) and low (15 %) during 2–5 ka BP and 4–8 ka BP respectively. Further it slightly increases to 18% during 8–15 ka BP. About 10 % of kaolinite was recorded during 14–17.5 ka BP (Fig. 4.8).

Chlorite content is very low (8 %) during 0-3 ka BP, but it increased to 20 % around 3 ka BP. The chlorite abundance is of 12 % during 4-5 ka BP and about 10 %

during 7-8 ka BP. From 8 to 17 ka BP the chlorite content varies in cyclicity with increasing and decreasing order and ranges between 8-16 % (Fig. 4.8).

Core SK 145B/C-8

In this core also smectite content is higher like in other two cores. Smectite content is uniformly higher (~65 %) from 0–14 ka BP, whereas during 15–18.7 ka BP it gradually decreases (to 50 %) (Fig. 4.9).

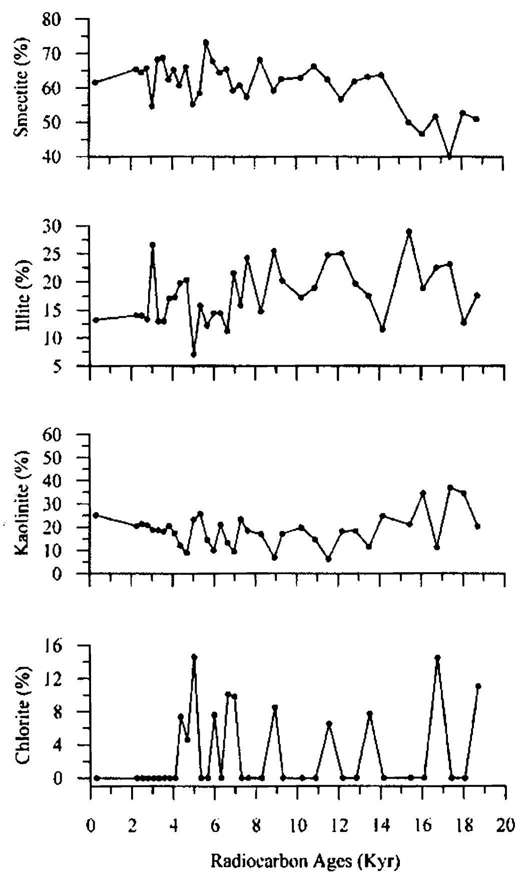

Figure.4.9. Temporal distribution of smectite, illite, kaolinite and chlorite in the core SK 145B/C-8.

Illite also varies significantly through out this core. It is comparatively less (~17 %) on the core top representing 0-5 ka BP, whereas during 5-13 ka BP an increase (~22 %) in illite content is recorded. A decrease (20-11 %) in illite content is recorded during 13-14 ka BP, and it increases (to 23 %) during 15–17.4 ka BP. Further, a decrease (to13 %) of illite is recorded during 18–18.7 ka BP (Fig. 4.9).

At this core site kaolinite is the second dominant mineral and it varies through out the core. Kaolinite content is more at the core top (~25 %) representing 0-4 ka BP.

During 4-13.5 ka BP a decrease (to 18 %) in kaolinite content is recorded, whereas during 14–18.7 ka BP an increase (to 35 %) in kaolinite content is recorded (Fig. 4.9).

It is interesting to note that chlorite is almost absent through out the core length, except at few core depths representing 4-6 ka BP, 8-9 ka BP, 11-12 ka BP, 13-14 ka BP and 16-1 7 ka BP where a small amount of chlorite is observed (Fig. 4.9).

4.4. Clay mineral ratios

Clay mineral ratios are extensively employed to decipher the climatic conditions of the region. The ratios of clay minerals rather than individual abundances will help to eliminate the effect of mutual dilution on clay mineral assemblages. Since kaolinite is formed under warm humid conditions and chlorite and illite under arid to cold climate, an increase in the ratio of kaolinite/chlorite and kaolinite/illite may indicate enhanced humidity.

Clay mineral ratios - kaolinite/chlorite (K/C), kaolinite/illite (K/I) and chlorite/illite (C/I) are shown in Figures 4.10, 4.11 and 4.12.

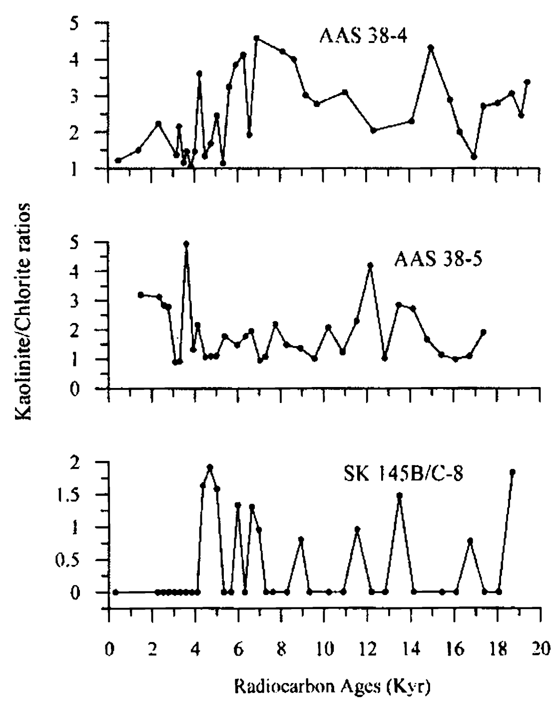

Figure.4.10. Temporal distribution of kaolinite and chlorite ratios in the cores.

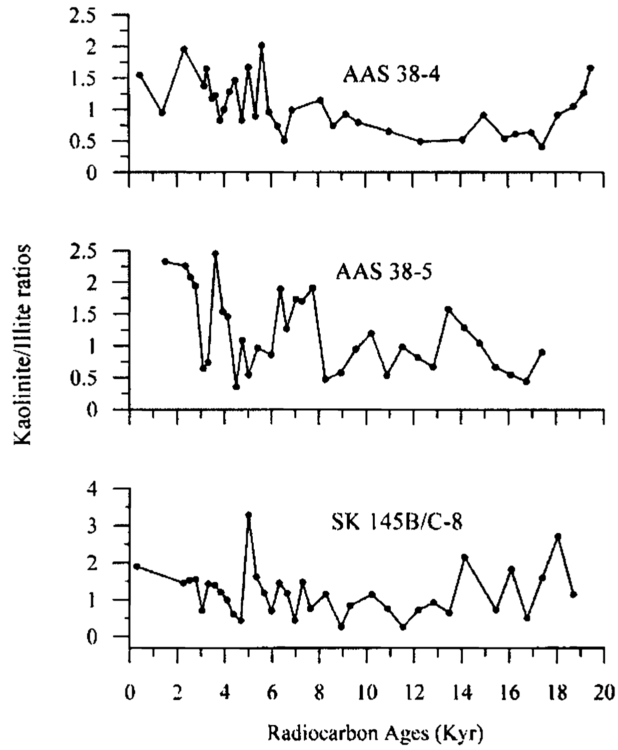

Figure.4.11. Temporal distribution of kaolinite and illite ratios in the cores.

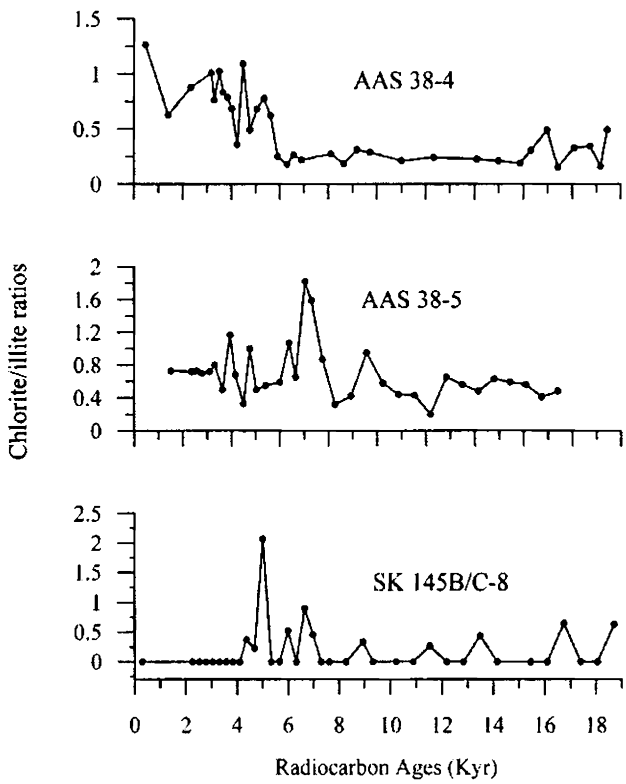

Figure.4.12. Temporal distribution of chlorite and illite ratios in the cores.

Kaolinite /Chlorite (K/C) ratio:

In core AAS 38-4, K/C ratios, in general, range from 1 to 4.5 (Fig. 4.10). K/C ratios range between 1 and 2 during 0–5 ka BP. During 5.5–11 ka BP K/C ratios range from 2.5 to 4.5 and during 11–14 ka BP the ratio is just around 2. The ratios decrease from 4.5 to 2 during14–17 ka BP and during 17–19.4 ka BP the ratio is around 3 in the core AAS 38-4 (Fig. 4.10).

In core AAS 38-5, K/C ratio is 3 during 0–3 ka BP and the ratio varies from 1 to 2 during 3–12 ka BP. Both increasing and decreasing trends ranging from 1 to 3 were recorded during 12–18 ka BP in the core AAS 38-5 (Fig. 4.10).

In core SK 145 B/C-8, as chlorite content is zero, most of the samples during different time intervals exhibit K/C ratios as zero. Because of this, the trend of K/C ratio cannot be explained with a definite conclusiveness in the core SK 145 B/C-8 (Fig. 4.10).

Kaolinite/Illite (K/I) ratio:

In general, the core AAS 38-4 exhibits high (1-1.5) K/I ratio during late Holocene (0–5.5 ka BP); the ratio is ~1 during 6–8 ka BP and during 8–17 ka BP, K/I ratio is <1 (0.5-1). K/I ratio increased from 0.5 to 1.5 during 17–19 ka BP in core AAS 38-4 (Fig. 4.11).

In core AAS 38-5, K/I ratios range from 0.5-2.5. The ratios are ~2 and 0.5 during 0–2 ka BP and at 3 ka BP respectively. It gradually decreases from 2 to 0.5 during 3–4 ka BP, and increases from 0.5 to 1.5 during 4-7 ka BP. The ratio ranges around 0.5 during 8-13 ka BP and increases to 1.5 at 13.5 ka BP. It gradually decreases from 1.5-0.5 during 13.5-17.4 ka BP (Fig. 4.11).

In core SK 145B/C-8 also, K/I ratio generally range from 0.5-2.5. K/I ratio is around 1 during 2-13.5 ka BP. It shows increasing trend (up to 3) during 13.5-18 ka BP (Fig. 4.11).

Chlorite/Illite (C/I) ratio:

Chlorite/illite ratios in core AAS 38-4 range from 0.2 to 1.2. It is about 0.8 during 1 -5 ka BP and it gradually reduces from 0.8 to 0.2 during 5 -6 ka BP. The ratio is around 0.2 and 0.4 during 6–16 ka BP and 16–19.4 ka BP respectively in the core AAS 38-4 (Fig. 4.12).

In core AAS 38-5, C/I ratio is about 0.8 during 0 – 6 ka BP. During 6.5–8 ka BP C/I ratio gradually reduces from 1.6 to 0.4. The ratio varies from 0.4 to 0.8 during 8–17.4 ka BP in the core AAS 38-5 (Fig. 4.12).

In core SK 145 B/C-8, C/I ratio also behaves similar to that of K/C ratio, as chlorite content is zero. The C/I ratio is zero throughout the geological time span of the sediment column (Fig. 4.12).

4.5. Provenance

Clay mineral characteristics of the aquatic sediments reflect the prevailing climatic conditions, hydrography, geology and topography of the continental source area (Chamley, 1989). The variations in clay mineral abundances, therefore, are a tool for deciphering the sediment sources and transport pathways to an area. Clay minerals in the marine environment are found to be largely detrital and widely dispersed (Biscaye, 1965). Since the clay minerals are ubiquitous in the marine sediments and their measurement is relatively easier and economic than the other sophisticated techniques, they provide a potential tool for paleoclimatic reconstruction. Presuming that the geology and geomorphology of the source region remained fairly stable for the time period in consideration in a tropical region, rainfall seems to be the main factor determining the composition of clay minerals in the marine sediments (Singer, 1984; Chamley, 1989; Thamban et al., 2002). Clay mineral composition basically indicates the intensity of weathering, especially the degree of hydrolysis, at source rock regions that can be used as paleoclimatic indicators (Chamley, 1989). Diagenetic effects on different clay minerals are considered to be insignificant and therefore the provenance of different clay minerals can be identified (Grim. 1968). Nevertheless a cautious approach is warranted as many other processes are acting upon clay minerals simultaneously or immediately after reaching the marine environment, suppressing the climate induced differentiation of clay minerals. These processes include size sorting of clay minerals during transportation (Gibbs, 1977), flocculation at lower salinities (Grim, 1968), selective deposition of certain clay minerals in relation to organo-mineral interactions (Degens and Ittekot, 1984), texture related clay mineral variations (Maldonado and Stanley, 1981), dispersal of clay minerals due to the prevalence of regional and local currents (Kolla et al., 1981; Naidu et al., 1985), redistribution of settled clays during reworking (Biscaye, 1965). As a consequence of these factors, selective transportation, deposition or mixing of clay minerals from different sources takes place.

Clay mineral distribution in the surficial sediments along the western margin of India indicates three different sources: (i) Himalayan source in the north charecterised by high illite and chlorite that are transported mainly by Indus river, (ii) a Deccan Trap source in the central region charecterised by abundant smectites which are the erosion products of basic volcanic rocks of central India, and (iii) a gneissic province charecterised by high kaolinite and gibbsite that are the weathering products of Precambrian gneissic rocks and laterites of southern India (Rao and Rao, 1995).

Smectite occurs along the Indian margin south of the Indus river, which resulted from the weathering of Deccan trap basalts, and appear to be primarily dispersed towards south by surface currents. In the southern most part of the Indian margin, kaolinite rich sediments probably derive from tropical soils of southern India, and are mainly distributed by north equatorial currents. The Himalayan source of clay minerals and their possible transport by the Indus river to the present study area can be ruled out, as the influence of Indus river on the shelf sedimentation is largely limited to the north of Gulf of Kachchh because of the presence of macro tidal currents, which act as barrier for longshore sediment transport to the south (Nair et al., 1982). High smectite and relatively low illite and chlorite values are characteristic of the clay minerals in the study area. Low concentrations of illite were also recorded in sediment cores from the continental slope of southern India by Sirocko and Lange (1991). Abundance of smectite in present study area suggests its derivation from the erosion products of basic volcanic rocks from central India under a semi-arid climate, and transported southward in the shelf and slope regions during the southwest monsoon because of strong southward currents. Further, the smectite might have also been derived from the submerged basic volcanic rocks in the continental shelf and slope regions and in the Lakshadweep area of southwest coast of India.

Similar to the results of present study, Rao et al. (1983) and Rao and Rao (1995) have also recorded smectite, kaolinite and illite in the decreasing order of abundance in the surficial sediments of the continental shelf and slope off southwestern India. But Thamban et al. (2002) have observed kaolinite, illite, smectite, and chlorite in order of decreasing abundance and opine that higher amounts of kaolinite and illite clearly reflected their derivation from the hinterland, which consists mainly of Precambrian crystalline rocks. Intense chemical weathering and leaching of crystalline rocks under the tropical humid climates leads to the formation of mainly kaolinite and gibbsite (Chamley, 1989). It is suggested that high kaolinite in sediments of the upper continental slope of southwest India is due to the supply from hinterland source rocks and cross-shelf transport processes. Kessarkar et al. (2003) argue that low illite and high kaolinite off Cochin may be explained on the position of Western Ghats and the sedimentary formations lying between the coast and the Western Ghats. As the steep cliffs of the Western Ghats are situated on the Quilon–Cape Comorin, it is likely that residual illite and chlorite were released from the gneisses and schists under intense rainfall conditions and subsequently transported and deposited on to the continental slope. Since the Ghats are away from the coast between the Bhatkal and Quilon, the soils between the Ghats and the shore i.e., alluvium and the Warkalli beds, may have been subjected to active hydrolysis and drainage have preferentially released large kaolinite (Kessarkar et al., 2003).

We suggest that variation in distance by a few kilometers between the Western Ghats and the coast may not make much difference in the hydrological processes as the terrain has more or less similar geomorphic feature, except the wide coastal plains off Cochin, and as it experiences similar humid climate and high rainfall all along and undergo similar hydrolysis. Further, the rainfall is higher in the central than the southern segment of Western Ghats. Therefore, in our view, the residual and sedimentary kaolin deposits that occur as 10-20 m thick beds at several places along the southwest coast of India act as major source for kaolinite in the offshore region, apart from laterites and crystalline rocks of hinterland. Further, rhyolites and dacites of acid volcanic rocks that occur in and around St. Mary Islands and in the shelf region of Mangalore area, further north of the study area, might also be contributing for the abundance of kaolinite in the offshore region through the long shore transport of weathered products. It can be suggested that the sediment source of southern India is constantly delivering kaolinite and illite, but the sediment dispersion on the Indian continental slope is controlled by bottom topography, water currents and gravitational deposition. These mechanisms also control the distribution pattern of other clay components in this area. Smectite and kaolinite accumulation are found to be moderate to high and show only small differences between Glacial and Holocene distribution patterns.

4.6. Paleomonsoon and paleoenvironmental scenario

The climate fluctuations may be the prime controlling factor for the distribution of clay minerals. Several processes such as – size sorting during transportation (Gibbs, 1977), flocculation at low salinities (Grim, 1968), selective deposition of some clay minerals in response to organo-metal interactions (Degens and Ittekkot, 1984), dispersal due to the prevalence of regional and local currents (Kolla et al., 1981) etc. - act on the distribution of clay minerals in the marine environment. As a consequence of these processes, selective deposition or mixing of clay minerals from different sources takes place. Therefore, a proper approach is essential for meaningful climatic interpretations of clay minerals.

Flocculation of clays under changing pH conditions is limited to the coastal ocean and may not be significant in open ocean regions (Thamban et al., 2002). Close association between organic matter and fine-sized clay minerals has been suggested to be of importance in clay mineral concentrations (Degens and Ittekkot, 1984). Although the cores studied in the present work contain high organic matter, it might not have influenced the concentration of smectite as this mineral is mostly derived from the Deccan trap source rocks from the hinterland. Grain size variations also seem to have no relation to the clay mineral content as observed in the down-core variations at all the three core sites. Oceanic currents are good conduits for the transport of fine-grained terrigenous particles within the ocean (Kolla et al., 1981). Based on this, Rao and Rao (1995) suggested that the high illite content along the western continental slope off India might have northern source, and carried by the southerly monsoon currents. But the predominant Al-rich illite in the cores of present study suggests a hinterland origin for the clays along the southwestern margin of India.

The erosion and transportation of soils is climate dependent, which are the main source for the supply of terrigenous clays along the continental margin of India. As the western hinterland of India is characterized by steep slopes with heavy monsoonal rainfall, it is presumed that clays were not stored on land for periods more than the geological time span discussed in this study. Such a relatively faster terrigenous transport to the ocean is consistent with the higher average sedimentation rates along the western margin of India as reported earlier (Pandarinath et al., 2004 and references therein) and also in the present study.

As the sediments in the study area are mainly detrital, derived from the adjacent hinterland through the rivers, the temporal variations in the sedimentation rates and characteristics reflect the changes in weathering conditions as a result of fluctuations in the intensity of monsoon in the region. Under tropical humid conditions where kaolinte forms abundantly in well-drained soils during both glacial and interglacial periods, the variations recorded in the composition of terrigenous clay input may reflect changes in the intensity of monsoon (Chamley, 1989, p.436). In the Atlantic, off Niger River, abundant supply of kaolinite from about 13,000 to 4500 yr BP suggests very active precipitation over the continental river basin, especially in middle and upstream areas where kaolinite preferentially formed in Cenozoic soils (Pastouret et al., 1978).

The variations of sea level during alternating glacial-interglacial periods may have induced noticeable changes in clay mineral composition derived from river drainage basins or continental shelves. Indirect control of late Quaternary climate on clay assemblages through sea level variations are reported in the literature (e.g. Japan Sea, Oinuma and Aoki, 1977; Behring Sea, Naidu et al., 1982).

The specific clay mineral ratios are employed to infer climatic conditions. Since the kaolinte is formed under warm humid conditions and illite and chlorite under cold to arid conditions, changes in ratios of K/C and K/I serve as indicators of humidity (Chamely, 1989; Vanderveroet et al., 1999). Kaolinite content, and K/C and K/I ratios, which serve as proxies for continental humidity, indicate the distinct events of monsoon intensification. Crystallinity and chemistry of illite provide proxies for the intensity of hydrolysis on land (Chamley, 1989; Gingele, 1996). The characteristic high rainfall and temperature in the coastal Peninsular India would lead to strong hydrolisation of illite.

The last glacial period as recorded in the cores is characterized by significant contribution of physical weathering products, indicating relatively arid conditions prevailed at source regions. Proxies for continued humidity (kaolinite content, K/C and K/I ratios) seem to indicate that distinct events of monsoon intensification punctuated the generally weak summer monsoon. Based on the records of enhanced values of kaolinite content and K/C ratios during ~22000-19000 yr BP, Thamban et al. (2002) have inferred an intensification of monsoon, and consequently an increased transport of chemically weathered products, during this period, which may indicate an enhanced precipitation related to monsoons and/or sea level fall. However, the summer monsoon activity during Last Glacial Maximum (LGM) was reported to be weak (Duplessy, 1982), and this period corresponds to rapid sea level fall during the late stages of Pleistocene glaciation (Fairbanks, 1992). During the lowered sea level, rivers may have debouched directly on the slope region and consequently they would have contributed hinterland derived weathered products such as kaolinite in enhanced quantities.

A significant increase in hydrolysis and erosion on land that occurred during the early deglaciation suggest a major climatic amelioration (Thamban et al., 2002). Several proxy records from Indian monsoon regime also support such an early deglacial initiation of summer monsoon conditions (Sirocko et al., 1993; Overpeck et al., 1996; Naqvi and Fairbanks, 1996). A steep increase in kaolinite concentration, K/C and K/I ratios during the late deglaciation between 8800 and 6400 yr BP, a period of reduced humid conditions, was recorded by Thamban et al. (2002) and they attributed such increase to enhanced chemical weathering and fluvial inputs. They further observed increased K/C and K/I ratios at about 6400 yr BP and argued that precipitation around this time must have been sufficient enough to evolve perennial river systems, and eroding and transporting the terrigenous materials.

Various studies have reported that the intensity of SW monsoon was high between ~13000 yr BP with a maximum at ~9000 yr BP (Prell, 1984; Van Campo, 1986; Naidu and Malmgren, 1996; Rajagopalan et al., 1997). But Thamban et al. (2002) suggest that the monsoon intensity was low during 13000-9000 yr BP and enhanced precipitation started after 9000 yr BP. They further observed that the relative variations in clay mineral contents and their ratios in the late Holocene suggest a reduced summer monsoon activity since ~5600 yr BP, and this is similar to the observations made by others (Van Campo, 1986; Rajagopalan et al., 1997; Enzel et al., 1999) with regard to paleomonsoon records. However, no such markers could be established in the present study.

The temporal variations documented in clay minerals in the present study show that kaolinite, chlorite and illite levels show an oscillating trend. Smectite levels are constant through out, kaolinite and chlorite levels are high since 6 ka BP, while illite is higher than both kaolinite and chlorite during 19.5 – 6.8 ka BP. High ratios of kaolinite/chlorite from 19.5–6.3 ka BP suggest the prevalence of humid conditions. Low K/C ratios since Mid-Holocene to the present indicate reduced monsoonal activity and consequently low weathering rates. Chlorite forms under arid conditions and high chlorite/illite ratio from 6.3 ka BP to the Present provide evidence for the extent of aridity. Generally, illite and chlorite form under dry/arid conditions and kaolinite under humid conditions. The gradual decrease in illite and increase in kaolinite from 17.4 ka BP suggest that the climate has gradually turned to warm humid conditions since then in this region. Based on clay mineral proxies, Gingele et al. (2004) have inferred more humid conditions between 11 and 6 ka BP and the onset of more arid conditions around 5.5 ka BP, which reached a maximum at 3.6 ka BP. It appears that these arid conditions are responsible for the low sediment supply in the last 3 ka in the southwestern margin of India.

Periods of relatively warm water in the sea correlate on land to relatively strong hydrolysis (i.e., high rainfall and temperature), responsible for the degradation of illite, the production of pedogenic kaolinite and smectite. Opposite conditions marked by low rainfall and temperature appear to have prevailed during periods of cold sea water, leading to the preservation on land of pre-existing, rather well-crystallized illite, chlorite and smectite (Chamley, 1989). In addition to clay mineral data, the levels suggesting enhanced hydrolysis on land tend to be enriched in coarse sediment fraction, calcium carbonate and in feldspars. Therefore, the paleoclimate information deduced from clay mineral assemblages in marine sediments basically concerns the degree of hydrolysis at the surface of exposed landmasses. Under tropical humid latitudes where kaolinite forms abundantly in well-drained soils during both glacial and interglacial periods, the variations recorded in the composition of terrigenous clay input may reflect temporary changes in the intensity of rainfall. Abundant supply of kaolinite correlates with very high sedimentation rates, which suggests very active precipitation over the continental river basin (Pastouret et al. 1978). On the other hand, low sedimentation rates correspond to enhanced supply of detrital smectite and chlorite, suggesting either a decrease in the precipitation (i.e., dryer climate), or preferential rainfall as observed by Pastouret et al. (1978) on the coastal zones of North-Equatorial Western Africa.

A careful evaluation of the modern distribution and possible factors affecting the clay mineral abundance along the eastern Arabian Sea indicate that the past variations of clay mineral proxies may be used to reconstruct the paleoclimatic conditions on the Indian subcontinent. Since lithogenic sedimentation along the southeastern Arabian Sea is mainly run-off related, the clay minerals would carry signatures of the intensity of hydrolysis on land and the fluvial strength.

Since kaolinite is formed under warm humid conditions and chlorite and illite under arid and cold climate respectively, an increase in the ratio of kaolinite/chlorite and kaolinite/illite may indicate the prevalence of humid climate (Gingele, 1996). During most of the last glacial period, the eustatic sea level remained lower than the present level with a lowest value (~120m) during the LGM (Fairbanks, 1989). Widespread arid conditions (Prell, 1984) in association with the lowered sea level during the LGM would have lead to the enlargement of desert area of the northwestern India and the dust laden coastal plains. Clay mineral data suggest that the illite concentration increased during glacial period. The increased illite and chlorite input and decreased kaolinite concentration could be attributed to an enhanced dust input related to an expanded desert region and reduced fluvial supply during LGM.

4.7. Summary

Clay minerals basically express the intensity of weathering, and especially of hydrolysis, on the land masses adjacent to sedimentary basins. Since the lithogenic sedimentation along the western continental margin of India is mainly attributed to river run-off, the clay fraction should contain signatures of past monsoonal climate on land.

The dominant clay minerals recorded in the sediment cores are smectite, kaolinite, illite and chlorite. Their relative abundance varies significantly within and among the cores. Smectite is the dominant mineral in all the cores followed by illite in upper as well as deeper slope regions, except in one core where kaolinite is the second abundant. There is a distinct variation in abundance during the Holocene and LGM.

The temporal distribution of clay minerals is controlled by the monsoon intensity, hydrolysis, sea level changes and climate of the region. High ratio values of clay minerals, particularly, K/C and K/I suggest strong humid conditions since last glacial period.

Since kaolinite is formed under warm humid conditions and chlorite under semi-arid to arid and cold climate, an increase in the ratio of kaolinite/chlorite may indicate enhanced humidity. Illite forms under cold humid conditions and an increase in chlorite/illite ratio would suggest an arid climate on land. As smectite is formed by intense chemical weathering of hinterland basaltic rocks under semi-arid conditions, increase in smectite/illite ratio should also indicate increased precipitation on hinterland.

Monsoon intensity, southerly coastal currents and humid climate played a major role in the clay mineral distribution and abundance. The study strengthens the idea of employing the clay minerals as proxies to decipher paleomonsoon, paleoclimate and paleoenvironmental conditions during the Late Quaternary.

CHAPTER-V

Major and Trace Elements of Sediment Cores: Terrigenous and Climate Controls

5.1. Introduction

Sediments possess a record of past conditions of the ocean and they are, therefore, important proxies for the quantification of the various fluxes contributing to the global geochemical cycles. Ocean sediments exert a strong influence on oceanic chemical budgets as well as biogeochemical cycling of redox sensitive elements, if they are organic rich and sub-oxic in nature. The occurrence and propagation of abrupt climate change between the high and low latitudes and during glacial and interglacial periods has become an important focus of paleoclimatic and paleoceanographic research. The causes of abrupt change have significant implications for understanding the future manifestations of similar forcings under late Holocene boundary conditions.

The chemical and mineralogical composition of marine sediments provides important information about their origin and can be used to reconstruct the chemical and physical conditions of the paleoenvironment, and in some cases to recognize climatic events. In the last two decades, Late Pleistocene-Holocene marine sequences have been the focus of several studies aiming to reconstruct the paleoclimatic and paleoceanographic aspects.

The variations in the composition of bulk sediments with regard to time are related to climate and these variations are controlled by the changes in the chemical composition of source material, bioturbation etc. (Finney et al., 1988). The concentrations of major, trace and minor elements in marine sediments reflect the range of chemical, oceanographic and sedimentary controls on their supply to, and their distribution in the ocean. Such controls include the composition of sedimentary detritus delivered to the ocean (Calvert and Pedersen, 1993). Examination of the behaviour of a group of elements will provide valuable information on the chemical state of the environment of deposition of ancient sediments provided the behaviour of such elements in the modern ocean is understood.

The chemical composition of marine sediments is controlled by the relative contributions of particulate materials derived from different sources having variable compositions. Trace elements are initially supplied to the site of deposition from terrestrial sources via rivers or atmosphere, and from biological productivity in the ocean. The distribution and vertical flux of most of the trace elements are controlled by the production, sinking and deposition of the biogenic particles (Bruland et al., 1994). The monsoon winds transporting dust are found to be responsible for the spatial and seasonal variations in fluxes of particles in the Arabian Sea (Nair et al., 1989; Rixen et al., 1996). Seasonally varying particle flux patterns make the southeastern Arabian Sea an ideal site to study the trace metal fluxes and their implications to the paleoclimate.

Bulk sediment geochemical data was utilized in deciphering the sedimentation history of Arabian Sea. Major element composition provides insight into the relative proportions of major components in marine sediment: terrigenous detritus, biogenous material, and diagenetic products. Elements and element ratios related to terrigenous material also help to identify provenance characteristics and thus changes in climate and/or sediment supply. Changes in paleoproductivity are mirrored by elements related to biogenous processes, and the oxygenation state of the water column may be deduced from the abundance of redox-sensitive elements. Beside these paleoceanographic studies, study of the whole sediment column can, therefore, help to locate the depth of past and still-active biogeochemical processes.

Several recent studies of marine sediments from the Arabian Sea have demonstrated their utility for reconstructing abrupt climate change during the last glacial period. In particular, indicators of marine paleoproductivity, sea surface temperature (SST) and dust flux record intensity fluctuations in the dominant monsoon climate (e.g. Prins et al., 2000; Pattan et al., 2003; Agnihotri et al., 2003, Sirocko et al., 1993, 2000). Sirocko et al. (2000) have studied 37 cores from the Arabian Sea for major and trace element geochemistry for four time periods of Holocene and the last glacial to detect tracers of past variations in the intensity of monsoon circulation and its hydrographic circulation. They suggest that the geochemical multi-tracer approach provides information on the history of deep sea ventilation, which was lower during the last glacial maximum than during the Holocene. Nath et al. (1997) have analysed trace and rare earth elemental variations in sediments of Oxygen Minimum Zone (OMZ) of the Arabian Sea and observed that sediment deposited in OMZ exhibit enrichment in U and organic carbon. They argue that the enrichment of Uranium and organic carbon could be due to high supply of organic matter to the sea floor.

There a few studies on the paleoproductivity and paleomonsoon based on Arabian Sea sediments. Based on geochemical evidences, Higginson et al. (2004) have discussed abrupt changes in relative strength of Arabian Sea monsoon during stadial/interstadial climate transition. Banakar et al. (2005) have studied the monsoon related changes in sea surface productivity and water column denitrification in the eastern Arabian Sea during the last glacial cycle. They suggest that eastern Arabian Sea was more saline during LGM than in the Holocene due to fresh water flux to the region and intensified winter monsoon. Chodankar et al. (2005) have reconstructed the past salinity gradient in the eastern Arabian Sea that serves as a proxy for the variation in fresh water flux and summer monsoon intensity. Late Quaternary paleoproductivity variations in the eastern Arabian Sea were studied by Agnihotri et al. (2003) using the accumulation rates, organic carbon, Sr and Ba. They observed the decreased surface productivity during the last glacial - interglacial transition.

The application of the geochemical approach indicates that the intensity of the southwest monsoon was low during the LGM, intensified slightly at the end of this interval. The initial studies of the inorganic geochemistry of the Arabian Sea were restricted to the continental margins; Fe and Ti were concentrated in the lithoghenic, quartz rich and coarse grained sediments of the shelf and upper slope, while Mn, Ni, Cu and Zn occurred most frequently in the clay mineral rich, fine grained sediments of the lower slope and deep ocean. The Quaternary paleoclimate records have shown that the southwest monsoon intensity has fluctuated in the past many times. Strong surface winds associated with the SW monsoon cause coastal upwelling, allowing the deeper and nutrient rich water to rise in to the photic zone, hence by increasing the surface productivity, which results in the enrichment of organic matter to the bottom sediments, since upwelling is wind-driven, surface water productivity is sensitive to changes in climate conditions (Summerhayes et al., 1992). Combination of high productivity and moderate ventilation of the thermocline leads to an intense oxygen minimum zone at water depths between 150 and 1250m in the Arabian Sea (Wyrtki, 1971; Naqvi, 1987). Productivity in the Arabian Sea is high, and the major contribution comes from during southwest monsoon period (Nair et al., 1989). Hence, past variations in the intensity of southwest monsoon could have been well recorded in the marine sediments of Arabian Sea.

While using the trace element concentrations, in paleoenvironmental reconstruction, it is essential to assess whether they are relatively enriched or depleted. Commonly, the degree of enrichment or depletion of a trace element in a sample is evaluated relative to its concentration (Wedepohl, 1971, 1991; Mclennan, 2001). Most of the times, trace elements in fine grained siliciclastic sediments are used for paleoenvironmental interpretations.

The recent development of a suite of geochemical (Calvert and Pedersen, 1993; Piper and Isaacs, 1995; Rosenthal et al., 1995) and isotopic (Calvert et al., 1992; Altabet and Francois, 1994; Ganeshram et al., 1995; Farrel et al., 1995) indicators of paleooxygen and or paleoproduction changes throw light on historical rates of ventilation and export production in coastal basins. Thus the geochemical behaviour of Cu, Cr, V, Ni, Ba and major elements - Fe, Ti, Al and Mn - can be used to track directly the history of deep water renewal, upwelling and productivity on the continental margin. When combined with information on a variety of other data such as sedimentology, micropaleontology and the chronostratigraphy of the cores, it is possible to reconstruct more accurately the paleoceanographic history of the region.

In this study, an attempt is made to assess the productivity changes and terrigenous dilution effects in response to last glacial and interglacial oscillation through biogenic and inorganic proxy (major and trace elements) in three cores recovered from the southeastern Arabian Sea. In the present study, major (Si, Al, Ti, Fe & P) and trace elements (V, Cr, Co, Ni, Cu, Sr and Ba) are employed to infer paleoclimate and paleoproductivity in the southeastern Arabian Sea.

5.2. Major elements

Al, Fe, Ti and Si, major components of aluminosilicates, tag terrestrial inputs to the sediments. Abundances of these elements (except Fe) are not affected by redox changes resulting from oxygen utilization during organic carbon oxidation in the water column and in the sediments. Down core variation of accumulation rates of these elements are mostly controlled by supply of detrital material during the past. Aluminium has been used as an indicator of clay detritus mainly derived from continental weathering by several researchers (Somayajulu et al., 1994; Reichart et al., 1998; Agnihotri et al., 2003; Pattan et al., 2003). Hence down core variations of major elements normalized with Al can be successfully used to infer past changes in provenance associated with those in wind patterns/intensities (Sirocko et al., 1993, 2000; Reichart et al., 1997). Phosphorous (P) is also considered as an important indicator of productivity since P is entering the sediment via biogenic material, i.e., through fish remains.

The major element concentrations in sediment cores analysed were Si, Ti, Al, Fe, Mn, Mg, Ca, Na, K, P and percentages were computed for all these elements. However, in the present work Si, Ti, Al, Fe and P only are discussed. The data on major elements is available only for two cores i.e., AAS 38-4 and AAS 38-5, and the core samples of SK 145B/C-8 could not be analysed because of limited analytical facility.

Silicon (Si)

Among the major elements, Si is the dominant one in both the cores and it ranges between 36 and 45%. At the core site AAS 38-4, the silica content is about 35-37% during 0-10 ka BP and it is about 43% during 10-19 ka BP. The supply of Si to the upper slope region is slightly decreased in the entire Holocene period, where as the Si supply was slightly higher during the last glacial period (Fig.5.1.). At the deeper slope region, i.e. at the core site AAS 38-5, the Si content is about 40% during 0-7 ka BP. It was increased by 5% i.e., upto 45% during 9-19 ka BP (Fig.5.2.).

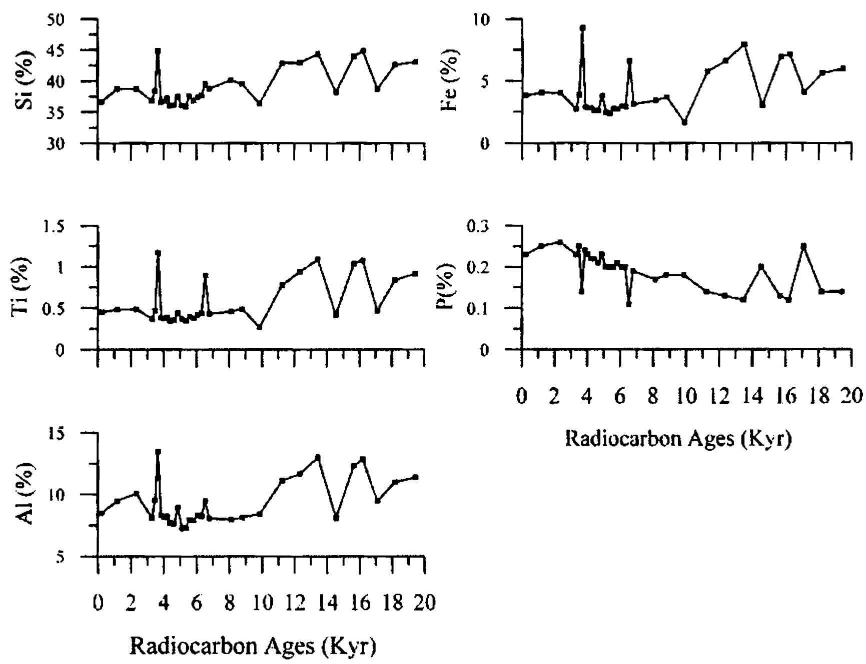

Figure. 5.1. Temporal variation of major elements in the core AAS 38-4.

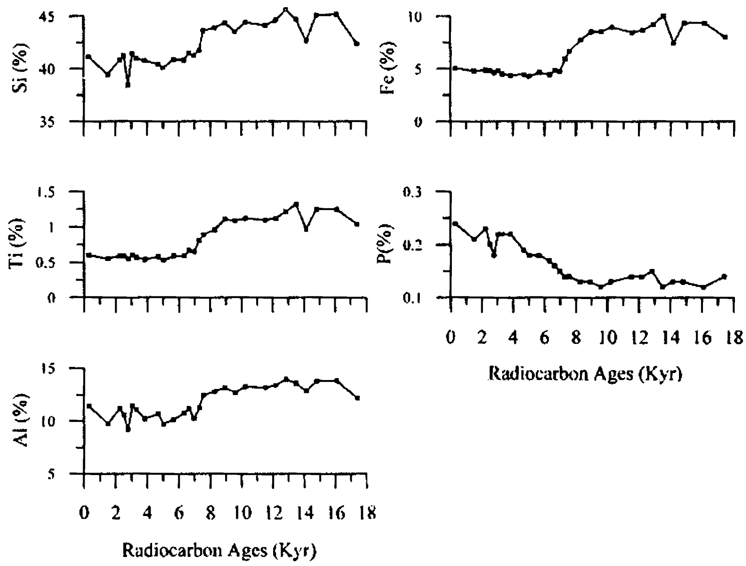

Figure. 5.2. Temporal variation of major elements in the core AAS 38-5.

Titanium (Ti)

Titanium distribution shows two distinct variations, - (i) Ti content is <0.5% during 0-10 ka BP i.e., Holocene and (ii) >0.5% during the last glacial period in the upper slope region (core AAS 38-4) (Fig.5.1.). In deep water region (core AAS 38-5), Ti content is ~0.6% during 0-7ka BP; it gradually increases to 0.9% during 7-8.5 ka BP and is >1% during 8.5-17.5 ka BP i.e., early Holocene to last glacial period (Fig.5.2.).

Aluminium (Al)

Aluminium behaviour is almost similar to that of Si and Ti in core AAS 38-4. It is slightly lower ~8% during 0-10 ka BP i.e., Holocene, and 11% during the last glacial period in the core AAS 38-4 (Fig.5.1.). In core AAS 38-5, Al content is ~10% during 0-7 ka BP and 12 to 14% during 7-18 ka BP (Fig.5.2.).

Iron (Fe)

Iron content is ~4% during 0 -8 ka BP and gradually increases from 2 to 7% during 8 -19 ka BP in core AAS 38-4 (Fig.5.1.); whereas in core AAS 38-5, Fe content is ~5% during 0-7 ka BP and it gradually increases from 5 to 9% during 7 - 9 ka BP. Its content is 8 to 9% during 9-18 ka BP (Fig.5.2.).

Phosphorous (P)

Distribution of phosphorus distinctly differs with that of other major elements discussed in this work. In core AAS 38-4, it is ~0.2% during 0-6.5 ka BP and is low (0.1 to 0.18) during 6.5 -19 ka BP (Fig.5.1.). In core AAS 38-5, similar pattern is observed, i.e., it is slightly higher during 0-7 ka BP and lower (0.14%) during 7-18 ka BP (Fig.5.2.).

5.2.1. Major elemental ratios

The reconstruction of changes in monsoon intensity during late Quaternary relies on the combination of productivity and lithogenous input. The percentage of Al in sediment record generally reflects the concentration of terrigenous aluminosilicate detritus (Calvert et al., 1993; Reichart et al., 1997; Shimmield et al., 1991a). Normalising elemental data to Al is helpful for assessing compositional variation in the lithogenic record as well as identifying changes in the biogenous and authigenous sources of certain elements. Therefore, both major and trace elements of two cores are presented as ratios to Al.

Si/Al ratio

Si/Al is higher in the core top (~5) representing 0–9 ka BP. During 9–13 ka BP a gradual decrease (5-3.5) in Si/Al ratio is observed. During 14–15 ka BP, the ratio increases considerably, and during 15–19.5 ka BP, it decreases to 3.5 in the core AAS 38-4 (Fig.5.3.).

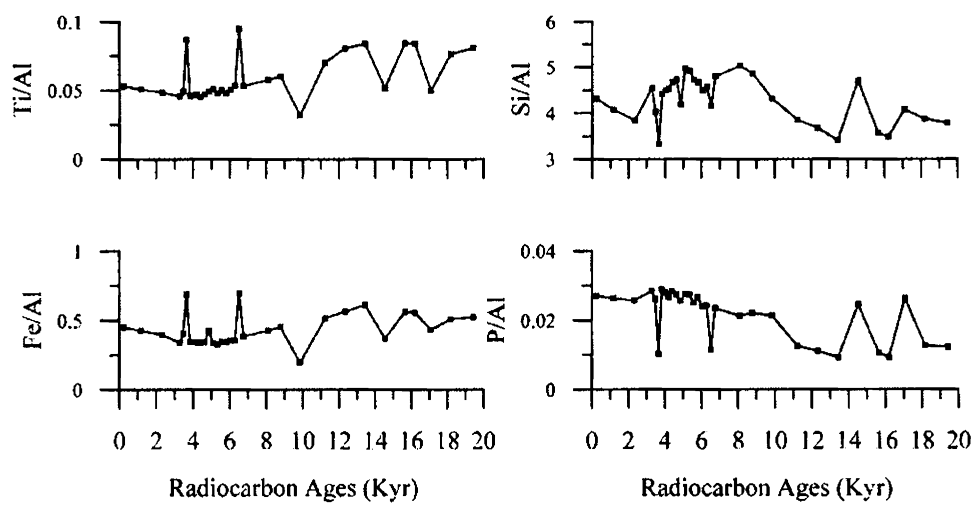

Figure. 5.3. Temporal variations of major elemental ratios in the core AAS 38-4.

In core AAS 38-5, Si/Al ratio exhibits the opposite trend to that of both Ti/Al and Fe/Al. Si/Al ratio is around 4 - 4.5 during 0 -7 ka BP, whereas during 7 -16 ka BP, the ratio is reduced to around 3 (Fig. 5.4).

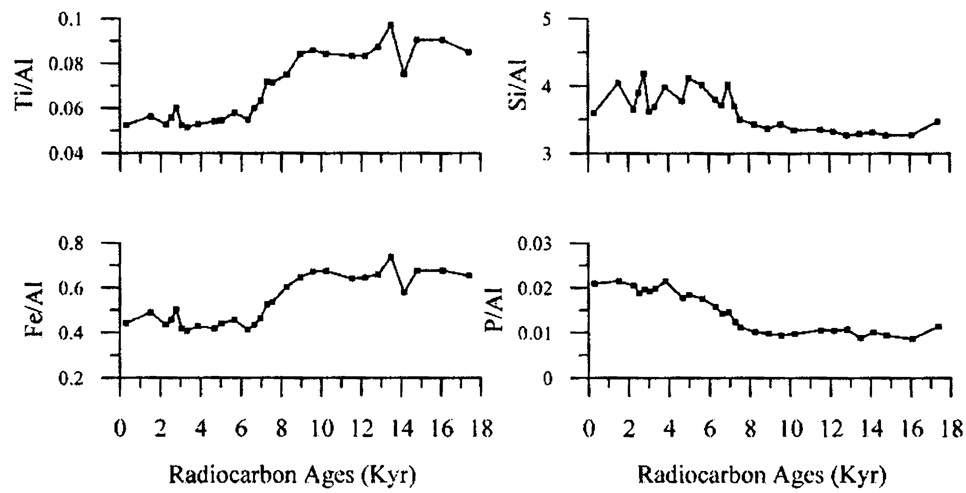

Figure. 5.4. Temporal variations of major elemental ratios in the core AAS 38-5

Ti/Al ratio

In core AAS 38-4, Ti/Al ratio is around 0.05 during 0–9 ka BP. During 9–10 ka BP it decreases to 0.025, and the ratio increased to 0.075 during 10–14ka BP. During 14–19.54 ka BP, Ti/Al ratio shows cyclicity of increase and decrease; but, in general, Ti/Al ratio increases during this period in the core AAS 38-4 (Fig.5.3.). In core AAS 38-5, Ti/Al ratio is around 0.05 (similar to that of AAS 38-4) during 0–7 ka BP. During 7–14 ka BP, Ti/Al ratio steadily increased to 0.1 and during 14–17.4 ka BP, Ti/Al exhibits a decreasing trend. Core AAS 38-5 exhibits the same pattern as that of AAS 38-4 in Ti/Al ratio (Fig. 5.4).

Fe/Al ratio

In core AAS 38-4, the behaviour of Fe/Al ratio is also similar to that of Ti/Al but the content of iron is more. Fe/Al ratio is ~0.5 during 0–9 ka BP and it decreases to 0.25 during 9–10 ka BP. During 10–14 ka BP, Fe/Al ratio increases to 0.6 and during 14 –19.4 ka BP; it shows a cyclic pattern, but in general, an increasing trend of Fe/Al is observed (Fig.5.3.). During 0–7 ka BP Fe/Al is ~0.5, whereas during 7–14 ka BP, the ratio increased to 0.8. Fe/Al ratio exhibits a decreasing trend during 14 – 17.4 ka BP similar to that of Ti/Al in the core AAS 38-5 (Fig. 5.4).

P/Al ratio

In core AAS 38-4, P/Al exhibits the similar trend as that of Si/Al (Fig.5.3.). During 0 –10 ka BP, P/Al ratio is ~0.02 and during 10–14 ka BP, P/Al ratio decreases gradually to 0.01, whereas during 14–19.5 ka BP, the ratio shows a rhythmic pattern with ups and downs. In core AAS 38-5, P/Al ratio also follows the same trend as that of Si/Al. P/Al is ~0.02 during 0 –7 ka BP, whereas during 7 –17.4 ka BP, P/Al ratio decreased to 0.01 (Fig. 5.4).

5.3. Trace Elements

While using the trace element concentrations for paleoenvironmental reconstruction, generally, the degree of enrichment or depletion of a trace element in a sample is evaluated (Wedepohl, 1971, 1991; Mclennan, 2001). Most of the times, trace elements associated with fine-grained siliciclastic sediments are used for paleoenvironmental reconstruction.

The surface productivity is expected to induce changes in oxygen level of bottom waters/sediment-water interface. Redox conditions of bottom waters/sedimentary column strongly influence the distribution of some of the trace elements such as V and Cr (Agnihotri et al., 2003). In pore waters, their oxidation states are determined by the ambient redox condition and hence these elements have different mobilities in oxic and anoxic environments. V and Cr are found to be depleted in the oxic zone and get enriched in the reducing strata (Dean et al., 1997). Downcore variations of these elements normalized to Al are often used to infer paleo-redox conditions at the time of deposition. The concentration of Ba in marine sediments has been widely used as a paleo-tracer of export production. Barite ($BaSO_4$) is thought to precipitate in sub-surface waters in microenvironments created by decaying organic matter (Ganeshram and Pedersen, 1998). Since barite formation is directly linked to biogenic material, Ba/Al can be used as an upwelling index (Dymond et al., 1992; Sirocko et al., 1996). Of the trace elements of biogenic origin, Nickel (Ni) is very similar to that of Corg. Hence the Ni/Al ratio can be related to the nutrient cycling and reflects the accumulation of organic matter, i.e., productivity. Since Strontium is always found associated with $CaCO_3$, the Sr concentration corrected for detrital contribution (Sr/Al) can also be used as an indicator of surface productivity (Reichart et al., 1997). Copper (Cu) concentration corrected with Al can also be used to infer the paleo-redox conditions during the deposition (Francois, 1988; Calvert and Fontugne, 2001).

In the present study, trace elements – V, Cr, Co, Ni, Cu, Sr and Ba - are discussed in relation to their chronology, paleoprodutivity and paleooceanography.

Vanadium (V)

V content is ~60 ppm during 0-9 ka BP, which increased to 137 ppm during 9.8-17 ka BP. A slight decrease is observed in the core AAS 38-4 during 18-19.5 ka BP (Fig.5.5).

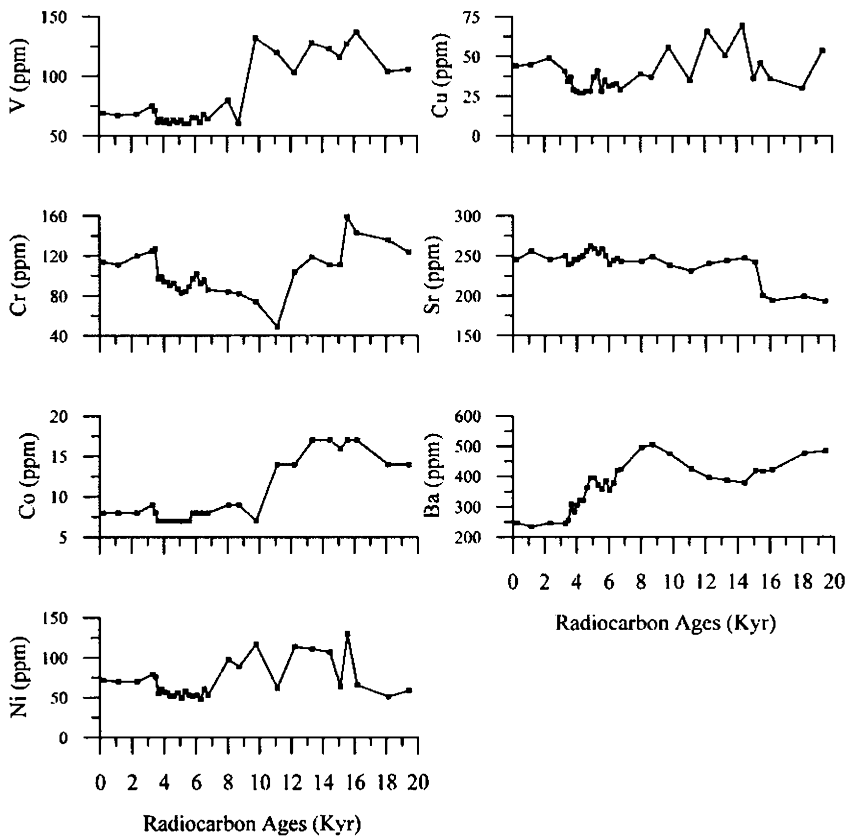

Figure. 5.5. Temporal variation of trace elements in the core AAS 38-4.

In core AAS 38-5, during 0-3 ka BP, V content shows a decreasing pattern (109-77 ppm), whereas during 3.3-4 ka BP, it increases to 107 ppm. During 4.5 -6 ka BP, V decreases gradually to 84 ppm. It increases to 180 ppm during 6-15 ka BP, and during 16–17.4 ka BP, a slight decrease (130 ppm) in V content is observed (Fig. 5.6). In core SK 145B/C-8, during 0-3 ka BP, V content shows a decreasing pattern (109-77 ppm), whereas during 3 - 4 ka BP it increased to 107 ppm. V decreases gradually to 84 ppm during 4.7-6 ka BP, and during 6-15 ka BP, it increases to 180 ppm. Again slight decrease (130 ppm) in V content is observed during 16–17.5 ka BP in the core SK 145B/C-8 (Fig.5.7).

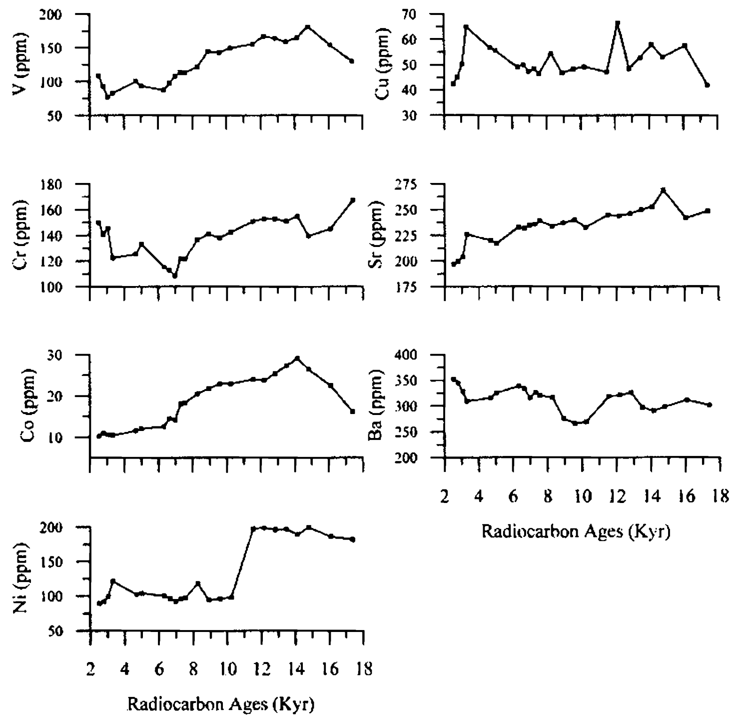

Figure. 5.6. Temporal variation of trace elements in the core AAS 38-5.

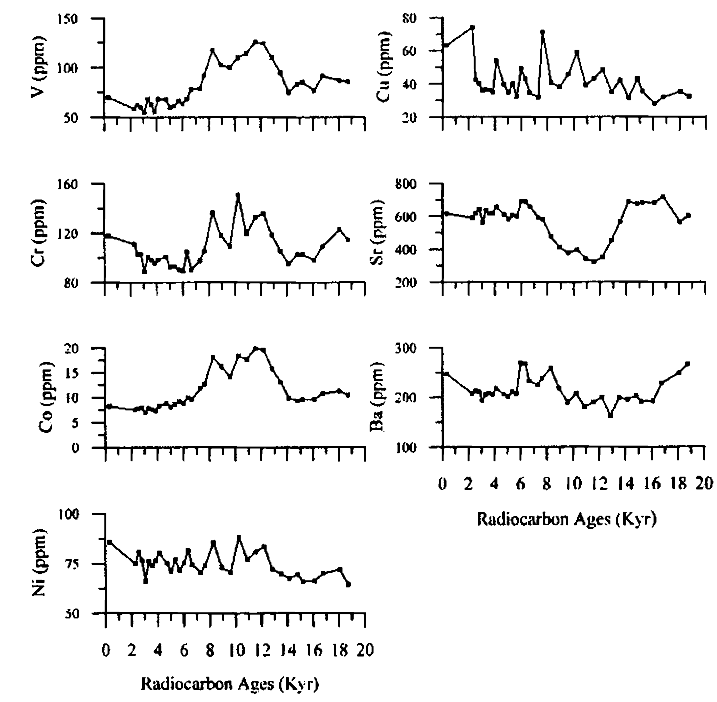

Figure. 5.7. Temporal variation of trace elements in the core SK 145B/C-8.

Chromium (Cr)

In core AAS 38-4, Cr content is more (120 ppm) during 0 - 3.5 ka BP, whereas during 3.5 -11 ka BP it decreases gradually to 49 ppm. During 12-15 ka BP, Cr content increases (to 110 ppm) gradually, whereas during 15.5–16 ka BP, it increased further to 160 ppm, and again decreased gradually (to 120 ppm) during 16 - 19.5 ka BP in the core AAS 38-4 (Fig.5.5). In core AAS 38-5, Cr content shows a decreasing (150-110 ppm) pattern during 0 -7 ka BP. During 7-17.5 ka BP, in general, Cr increases gradually to 153 ppm, whereas during 14 -17.4 ka BP, a small decrease (to 120 ppm) in Cr content is observed (Fig.5.6). In core SK 145B/C-8, Cr content is ~100 ppm during 0 - 4.7 ka BP whereas during 5 -6 ka BP it decreased to 90 ppm. Cr content increased to 150 ppm during 7.3–12 ka BP. During 13 - 14 ka BP, Cr content decreased to 95 ppm, and again it increased to 100 ppm during 14 -16 ka BP and during 16 -18.7 ka BP Cr content further increased to 120 ppm (Fig.5.7).

Cobalt (Co)

The distribution of Co content shows distinct variation from the Holocene to the last glacial period, with low content during the Holocene and higher content during the last glacial period (Fig.5.5). In core AAS 38-4, Co content is 7 – 10 ppm during 0-10 ka BP and 10-14 ppm during 10-12 ka BP. It is about 17 ppm during 12-19 ka BP (Fig.5.5). In core AAS 38-5, Co content is about 10 ppm during 0-7 ka BP, and it gradually increases from 15 to 28 ppm during 7 -14 ka BP. Again it decreases from 28 to15 ppm during 14 -18 ka BP (Fig.5.6). Core SK 145B/C-8 exhibits Co content of ~7 ppm during 0 -8 ka BP, and it is 15-17 ppm during 8-12 ka BP. Co sharply reduces from 17 to7 ppm during 12 -14 ka BP, and it is 7 ppm during 14 -18 ka BP (Fig.5.7). All the cores, in general, exhibit low Co content during the Holocene and higher content during last glacial period.

Nickel (Ni)

Core AAS 38-4 exhibits subtle variations of Ni content with in the Holocene and also during the last glacial period (Fig.5.5).Ni content is 75 ppm during 0 -3 ka BP, and about 50 ppm during 3 -7 ka BP. Ni content sharply increases to 100 ppm and almost remains the same amount during 8 -14 ka BP. It again reduces to 50 ppm during 15 -19 ka BP (Fig.5.5). Interestingly, core AAS 38-5 exhibits distinct variation in Ni content during the Holocene and the last glacial period (Fig.5.6). Ni content is ~100 ppm during 0-10 ka BP and it is ~200 ppm during last glacial period (Fig.5.6). The behaviour of Ni content in core SK 145B/C-8 is quite contrasting and is about 80 ppm during 0 -12 ka BP, whereas it is about 60 ppm during 13 – 19 ka BP (Fig.5.7).

Copper (Cu)

Copper distribution varies from one core site to the other. In core AAS 38-4, Cu content is 45 ppm at the core top that represents the time span of 0 -3 ka BP, whereas it decreased to about 25 ppm during 3.5 -7 ka BP (Fig.5.5). It increases gradually to 70 ppm during 8 -14.5 ka BP. However, Cu content decreases from 70 to 30 ppm in this core during 15- 19.5 ka BP (Fig.5.5). Core AAS 38-5 also exhibits increased Cu content at the core top i.e., during 0- 3 ka BP (Fig.5.6). It gradually decreases from 65 to 40 ppm during 3- 7.5 ka BP. It remains constant at 45 ppm during 7.5- 11.5 ka BP, but increases gradually from 45 to 60 ppm during 11.5 -16 ka BP (Fig.5.6). The Cu concentration in core SK 145B/C-8 is slightly lower than in other two cores. It generally ranges between 30- 50 ppm during 2- 19 ka BP (Fig.5.7). Cu content doesn't show any distinct variation between the Holocene and the last glacial period.

Strontium (Sr)

Sr accumulation is very high in all the three cores particularly in the core SK 145B/C-8 (Figs. 5.5- 5.7). In core AAS 38-4, Sr content is about 250 ppm during 0 - 15 ka BP, and it is 200 ppm during 15 -19 ka BP (Fig.5.5). In core AAS 38-5, Sr content increases gradually from about 200-240 ppm during 0- 14 ka BP, and it is about 220 ppm during 14 -18 ka BP (Fig.5.6). Sr concentration is much higher in core SK 145B/C-8 when compared to other two cores, and it ranges between 300 -600 ppm. Sr content is about 600 ppm during 0 -6 ka BP, it decreases gradually from 600 - 300 ppm during 6 -11 ka BP, and again it gradually increases from 300 to 640 ppm during 11 -14 ka BP. Sr content is about 650 ppm during 14 -19 ka BP (Fig.5.7).

Barium (Ba)

Barium content is generally high in all the three cores. In core AAS 38-4, Ba content is about 250 ppm during 0 -3.5 ka BP, and it gradually increases from 250 to 500 ppm during 3.5- 8.5 ka BP (Fig 5.5). Ba content gradually decreases from 500 to

370 ppm during 8.5-15 ka BP, and it gradually increases to 475 ppm during 15 -19 ka BP (Fig. 5.5). Ba content in core AAS 38-5, ranges between 300 -350 ppm during 0 -8 ka BP, and decreases to 270 ppm during 8 -10.5 ka BP. During 10.5 -13 ka BP, it increases from 270 to 370 ppm, whereas its abundance is about 300 ppm during 13 - 18 ka BP (Fig 5.6). In core SK 145 B/C-8, Ba content is about 200 ppm during 0 -6 ka BP; it is about 250 ppm during 6 -8 ka BP. Ba content is about 200 ppm during 9 -16 ka BP and it increases upto 300 ppm during 16- 19 ka BP (Fig. 5.7).

5.3.1. Trace elements/Al ratios

It is a general practice to plot Al vs trace elements in order to normalize the elemental ratios with respect to terrigenous supply and to evaluate the paleoproductivity of the region. Al is plotted against vanadium (V), chromium (Cr), cobalt (Co), nickel (Ni), copper (Cu), strontium (Sr) and barium (Ba) and the temporal variations of these ratios are shown in Figures.5.8 and 5.9. As Al content is not determined for the sediment samples of the core SK 145B/C-8, the ratios for this core are not available. V/Al, Cr/Al, Co/Al, Ni/Al ratios show similar uniform trend during 0 -7 ka BP (Fig.5.8).

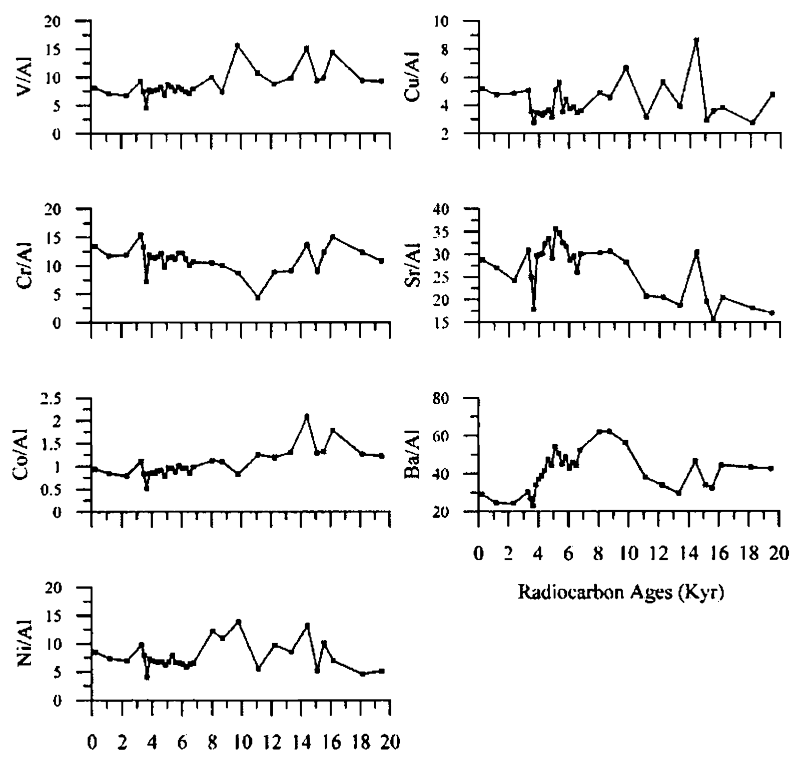

Figure. 5.8. Temporal variations of trace elements/Al ratios in the core AAS 38-4.

The V/Al ratios are generally higher during early Holocene and last glacial period in core AAS 38-4. It is observed from the Figure.5.9 that V/Al ratio is 10 during 0- 2.5 ka BP, and it gradually increases from 6 to 12 during 3 -14 ka BP in core AAS 38-5 (Fig.5.9). Cr/Al ratio is ~12 during 0- 3ka BP and it gradually decreases to 10 during 3 -7 ka BP in core AAS 38-4 (Fig.5.8). The Cr/Al ratio shows irregular pattern with both increase and decrease during 7- 19ka BP in core AAS 38-4. It is also observed that Cr/Al ratio varies from 10-15 during 0- 5 ka BP and the ratio is 10 during 6 -18 ka BP (Fig.5.9). Co/Al ratio is ~1 during 0-4 ka BP and it gradually increases to 2 during 4-19 ka BP in core AAS 38-4 (Fig.5.8). Co/Al ratio behaviour is similar to that of V/Al ratio in core AAS 38-5 and it gradually increases from 1-2 during 2 -14 ka BP. The ratio decreases from 2 to 1 during 14 -18 ka BP (Fig.5.9).

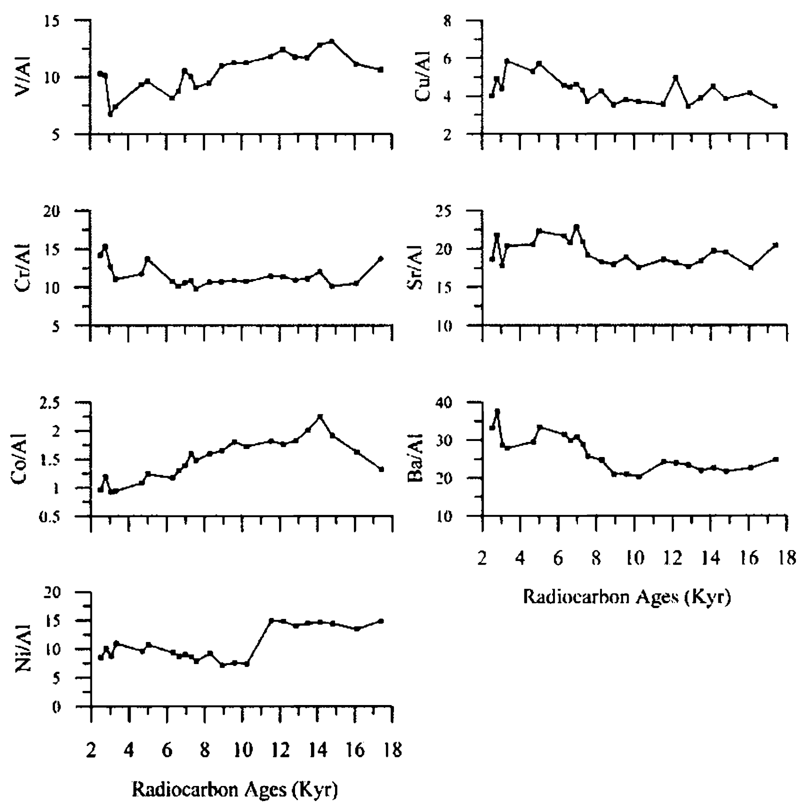

Figure. 5.9. Temporal variations of trace element/Al ratios in the core AAS 38-5.

Ni/Al ratio is ~8 during 0- 7 ka BP, whereas during 7 -19 ka BP, in general it increases to 12 but shows an irregular pattern with both increase and decrease in core AAS 38-4 (Fig.5.8). In core AAS 38-5, Ni/Al ratios clearly exhibit the demarcation between the Holocene and the last glacial period. The Ni/Al ratio is about 10 during 0

- 10 ka BP, and it is 15 during 11-18 ka BP (Fig.5.9). Cu/Al ratio is 5 during 0 -3 ka BP whereas it is 3 during 3 -7 ka BP. The ratio shows irregular pattern with increasing and decreasing trends during 7 -19 ka BP in core AAS 38-4 (Fig.5.8). Cu/Al plot also shows slightly higher values (4-6) during 0-7 ka BP, and 3 to 4 during 7.5-18 ka BP in core AAS 38-5 (Fig.5.9). It is interesting to note that Sr/Al ratio is higher (~27) during Holocene, whereas it is ~20 during the last glacial period in core AAS 38-4 (Fig.5.8). In core AAS 38-5, Sr/Al ratio is about 20 during 0-7 ka BP, whereas the ratio is about 7 during 8 -18 ka BP (Fig.5.9).

Ba/Al ratio is low (~25) during 0 -4 ka BP, and the ratio gradually increases to 50 during 4 - 5.5 ka BP in core AAS 38-4 (Fig.5.8). The ratio is about 45 during 5.5 - 7 ka BP, which later increases to about 60 during 7 -10 ka BP. Ba/Al ratio, then sharply decreases to around 30 during 11 - 16 ka BP. The ratio again increases to 40 during 16 -19.5 ka BP (Fig.5.8). In core AAS 38-5, Ba/Al ratio is higher (~30) during 0 - 8 ka BP, and it is about 20 during 8- 18 ka BP (Fig.5.9).

5.4. Terrigenous Supply and Dilution Effects

Abundances of major elements Al, Fe and Ti in cores AAS 38-4 and 38-5 are employed to investigate the changes in supply and source of detrital material to the eastern Arabian sea during the last glacial and the Holocene. The terrigenous records of cores reveal that the most significant, well-resolved environmental variations are evidenced during the early Holocene, associated with abrupt oscillations in the terrigenous signals. Thamban et al. (2007) have observed the monsoon reaching its peak about 9500 and 8000 years BP, as supported by various terrestrial proxies like elemental concentrations and magnetic susceptibility. This is coeval with the maxima in Indus discharge as well as the establishment of thermocline anoxia in the Arabian Sea (Staubwasser et al., 2002). Thamban et al. (2007) have also observed significantly increased Fe concentrations starting at 9100 years BP, suggesting a significant increase in precipitation-derived terrigenous supply to the eastern Arabian Sea. They have further recorded significant weak monsoon periods between 6000 and 5500 years BP and this event has well reflected in the records of all terrigenous proxies like Fe, Al and Ti. Such an event is also recorded in the planktonic foraminiferal oxygen isotope records of sediment cores from the south west 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). Another weak summer monsoon event is observed in the terrigenous proxy records around 3500 years BP. Sarkar et al. (2000) also suggests such change in Indian summer monsoon precipitation.

A rapid decrease in terrigenous supply at 1300 years BP (Thamban et al., 2007), and severe weakening of the summer monsoon around 1500 years BP were evidenced in the Arabian Sea (Gupta et al., 2003). But in this study, the terrigenous proxies such as Fe, Al and Ti do not show any decrease in concentration, which are the indicators of weak monsoon, rather they exhibit the increasing trends of concentrations. Downcore variations of Al content and its accumulation rate in both the cores show an enhance supply of detrital material during the LGM. Such enhanced Al content during the LGM was also observed by Agnihotri et al. (2003).

Al and Ti concentrations in marine sediments are the best indicators of lithogenous material flux to the marine environment (Pattan et al., 2003). Both these elements may have been derived from the extensive laterite deposits present in the hinterland of southwestern India and are supplied to the southeastern Arabian Sea. High Al and Ti values in core 38/5 lends support to the argument suggesting greater terrigenous dilution at this core site. The input of terrigenous material to the Arabian Sea varies on a regional scale and relates to monsoonal activities (Sirocko, 1989). Low Al and Ti concentrations during the Holocene and higher Al and Ti concentration during the LGM are recorded.

The Al concentrations in sediments reflect the provenance and dilution effects. The low values of Al (~10%) in sediments of the cores under present investigation is an intriguing factor because the hinterland consists of laterites and ultramafic rocks which are enriched in Al content. The low Al values in the eastern Arabian Sea sediments could be a dilution with lithic components that are low in Al content.

The Ti content in the core samples indicates a slightly higher content during the last glacial period than in the Holocene. This observation is opposite to that observed earlier by Sirocko et al. (2000). They have reported higher Ti content in the sediments of the Arabian Sea during the Holocene. Generally, higher Ti content occurs in shallow waters near the coast. As the cores under present investigation are located in the slope region, the enrichment of Ti is not effected.

The values of P content are usually below 0.1%. But, values of > 0.15% occur in deeper waters core, where the fine grained sediments are accumulated. The P content in the Arabian Sea sediments has been reported as the values that range between 0.07 and 0.15% (Sirocko et al., 2000). The recorded values in the present study also fall within this range.

The enriched Si content of the eastern Arabian Sea sediments reveals a maximum input of silicates to the southeastern region. The earlier workers have reported a maximum of Si content off the coast of Pakistan and gradient towards the south. This pattern resembles the distribution pattern of quartz reported by Sirocko and Lange (1991) and Sirocko et al. (2000). In addition to its direct relationship to the quartz content, the patterns of Si concentration resemble the distribution pattern of the sediment particle sizes. The pattern and high values of Si/Al ratio suggest that Si content is considerably higher throughout. Ti/Al ratio is considerably lower. However, the slightly higher ratio of Ti/Al during the last glacial period and early Holocene probably reflect the increase in contributions from the basalt provinces during that period. The relative abundance of Ti bearing heavy minerals such as ilmenite, titanomagnetite derived from the coastal sediments control large part of Ti content.

Higher concentrations of Al and Ti coincide with the low $CaCO_3$ values (see section 3.6) during the last glacial period suggests greater terrigenous dilution during LGM. It has been suggested earlier by many that the $CaCO_3$ records from the Arabian Sea are mostly influenced by terrigenous dilution, caused by variations in the terrigenous lithogenic flux derived from the Arabian and Somalian Peninsulas during the summer monsoon season (Murray and Prell, 1992), and by riverine material from the Indus river in the eastern Arabian Sea (Naidu, 1991). The similar dilution of $CaCO_3$ in the southeastern Arabian Sea is caused by the terrigenous lithogenic flux derived from the southwestern part of India, when the sea level was low and during the LGM.

5.5. Paleoproductivity

Marine sediments provide important information on paleoproductivity, past oceanographic and climatic changes. Because these changes cannot be determined directly, geochemical proxies of climatic variations and paleoenvironmental conditions have been developed during the last decades (Boyle, 1983; Calvert and Pedersen, 1993; Shimmield, 1992, Schneider et al., 1997; Wehausen and Brumsack, 1999; Schnetger et al., 2000). Most of the paleoproxies have restrictions that confine their application to well-defined environments. A large number of geochemical proxies, including major and trace elements, have been used to reconstruct the history of the productivity record.

The paleoceanographic, paleoclimatic history is inferred from the proxy records of these cores in terms of variations in productivity, $CaCO_3$ and organic carbon. Sirocko et al. (2000) reported the late Quaternary climate changes retrieved from various geochemical proxies in several sediment cores from the Arabian Sea. They identified three major climatic extremes during the last 25, 000 years-(i) early Holocene period of 8-10 ka BP is recognized as a humid interval when monsoon intensity was maximum at northwestern Arabian Sea; (ii) the period 16-18 ka BP is the period when surface productivity and monsoon intensity were found to be the lowest (Sirocko et al., 1993, 2000). The time interval 16-18 ka BP reported for minimum upwelling and surface productivity (Sirocko et al., 2000); (iii) the period 21-23 ka BP is the LGM where no major changes are observed. Sr/Al and Ba/Al ratios have been extensively used as indicators of surface productivity in the Arabian Sea as well as in other upwelling regions (Rixen et al., 2000; Schnetger et al., 2000; Ganeshram and Pedersen, 1998; Reichart et al., 1997; Agnihotri et al., 2003). Agnihotri et al. (2003) have reported low surface productivity during the LGM based on Sr/Al and Ba/Al ratios. The inferred low surface productivity during the LGM appears to be spatially and temporally consistent and is in good agreement with other paleoceanographic records in the eastern Arabian Sea (Sarkar et al., 1990, 1993; Agnihotri et al., 2003). Sr/Al and Ba/Al ratios show an abrupt decrease during the early Holocene, thereby indicating a sharp decrease in productivity during this period. Such change has also been observed by Agnihotri et al. (2003) in the eastern Arabian Sea. This time interval (8-10 ka BP) coincides with the early Holocene humid interval, a period of intensified southwest monsoon in the Arabian Sea (Van Campo, 1986; Overpeck et al., 1996; Sirocko et al., 2000).

A possible explanation for the observed sharp change during the early Holocene could be increased monsoonal precipitation on coastal regions, resulting in large amounts of fresh water flux to the core sites and as a consequence upwelling-induced surface productivity would have weakend/ceased (Thamban et al., 2001 and references therein).

It is well known that Ba is concentrated by marine organisms forming barite (Bertram and Cowen, 1997) which then resists remineralisation in water column. Therefore, it may be used as a productivity proxy under certain circumstances. Association of Cu with organic carbon is known from sediment trap studies (e.g., Jickells et al., 1984). Cu appears to be less sensitive to redox conditions than other metals (e.g., V, Ni, Zn) also enriched in biogenic material, and therefore, also may be used as a productivity proxy (Shaw et al., 1990).

The use of Ba or Ba/Al ratios as a paleoproxy has been documented in a large number of publications (see von Breymann et al., 1992; Shimmield, 1992; Wehausen and Brumsack, 1999). Although the exact mechanism is not fully understood, many researchers feel that Ba is incorporated into bioaggregates in the water column or is trapped by the active precipitation of barite by the living phytoplankton cells (Brumsack, 1989; Bertram and Cowen, 1997). The barite content of deep-sea sediments has been suggested to be a function of surface water productivity, water depth and mass accumulation rate (Dymond et al., 1992). Schnetger et al. (2000) have reported enrichment of productivity-related elements (Ba, Ca, Sr, P) during interglacial times. They further observed that during interglacial periods with higher productivity, the deepening of the OMZ in the Arabian Sea is a likely phenomenon.

5.6. Summary

The major and trace elements are employed to infer paleoclimate and paleoproductivity in the southeastern Arabian Sea. The abundance and ratios of major elements - Si, Ti, Al, Fe and P - are presented. All the major element abundances show distinct variation between the Holocene and the last glacial period. Si, Ti, Al and Fe are distinctly lower during the Holocene than in the last glacial period. However, P abundance is higher in the Holocene than in the last glacial period.

Trace elements - V, Cr, Co, Ni, Cu, Sr and Ba - are employed in the present study to understand the paleoceanography and paleoproductivity. V, Co, Ni and Ba are distinctly low during mid – and late Holocene. Although Ba content varies from one core site to the other core site, in general, Ba content is higher in all the three cores. Trace elements/Al ratios suggest that there are distinct temporal variations in all the three cores. The variation in the input of terrigenous material to the Arabian Sea relates to monsoonal activities. Al concentrations in sediments reflected the provenance and dilution effects. High Al and Ti values coincide with the low $CaCO_3$ values during the last glacial period, which suggests greater terrigenous dilution.

Geochemical records provide important markers of paleoproductivity and past oceanographic and climatic changes. The elemental/Al ratios show low surface productivity during the LGM. An abrupt decrease of Sr/Al and Ba/Al ratios during the early Holocene indicate sharp decrease in productivity. Ba/Al ratio as a paleoproxy has been signified in this study too. Productivity related elements have been enriched during interglacial times. Most of the productivity indicated elements are lowered during the Holocene than in the last glacial period.

CHAPTER-VI

Paleoenvironment and Paleoclimate during Late Quaternary - A Summary

Paleoenvironment and Paleoclimate during Late Quaternary
A Summary

Evaluating past oceanographic and climate changes through the analyses of deep- ocean sediments necessarily involves the use of proxy indicators. Proxy indicators by virtue of their physical, biological and or chemical origins are indirectly related to paleoceanographic and climatic variables such as sea surface temperature, wind speed, nutrient content, precipitation, global ice volume. No proxy currently in use can claim a unique direct association with a single climate variable; all have the potential to be influenced by processes other than climatic changes or oceanographic variable of direct interest. Some of the more common processes that complicate the interpretation of proxy variables include regeneration of biological material in the intermediate and deep waters, post depositional diagenesis, erosion and redeposition. In addition, some proxies are influenced by a number of climatic and or oceanographic variables making it difficult to isolate the effect of any single variable. Each of them is physically, chemically, biologically or isotopically linked to the variable of interest but impacted differently by unrelated processes such as diagenesis, dissolution etc.

The paleoceanographic and paleoclimate data obtained from the Arabian Sea particularly in the eastern Arabian Sea is rather sparse. The proxy records from the Arabian Sea sediments have provided useful information regarding monsoon induced upwelling, which has close relation with the climate prevailed in Asia during that period. Moreover, the understanding of productivity has a close linkage to pCO_2 levels of the atmosphere during glacial interglacial time scales (Broecker, 1982; Broecker and Denton, 1989; Sigman and Boyle, 2000). The traditional proxy for paleoproductivity is organic carbon which has been successfully used in many of the oceanic regions by various researchers (e.g. Rixen et al., 2000; Kawahata et al., 1998; Ganeshram et al., 1999). But still there are certain uncertainties and controversies regarding Corg accumulation and its linkage to monsoon and surface productivity particularly in the eastern Arabian Sea.

In order to understand the paleoenvironmental and paleoclimatic changes in the southeastern Arabian Sea during the late Quaternary period, the present study was taken up. The data generated in the present work would serve as an archive and add to the existing data on the southwest coast of India. To achieve the objectives, three sediment cores from the continental slope region off Cochin in the southeastern Arabian Sea were studied in detail. The core samples studied have provided information on sediment texture, sedimentation rates, organic carbon, calcium carbonate, clay minerals, and major and trace element compositions, which in turn are employed to infer the paleoenvironment and paleoclimatic changes.

The terrigenous material supply to the Arabian Sea is mostly influenced by the monsoon intensity and hinterland geology. The radiocarbon ages recorded for the sediment layers were upto 19500 years BP. Based on the radiocarbon ages, sedimentation rates were computed to understand temporal variations, depositional conditions and the paleomonsoon intensity in the region. The sedimentation rates varied from 4-27 cm/ka for the dated entire sedimentary column of the core. The higher sedimentation rate was recorded during the mid Holocene and the low sedimentation rate for the last 2 ka. Sedimentation rates generally decrease with increasing water depths as a result of reduced influence of the terrestrial input. The higher sedimentation rates were reported in the northeastern Arabian Sea, and reduced sedimentation rates in the southern Arabian Sea. The large-size rivers, when compared to small-size rivers in the south, that drain in the northern region through volcanic rocks may be the prime cause for high sedimentation, apart from the detritus from Indus river. Further, volcanic rocks that carpet the northern hinterland are more susceptible for physical and chemical weathering than the granitic gneisses and charnockites that occur in the catchment regions of southern rivers. The varied sedimentation rates reported for the eastern Arabian Sea by various workers can be attributed to the different dating techniques employed. For example, AMS C-14 dating technique provides more accurate results for the late Quaternary sediments than any other technique.

Texturally, all the three sediment cores exhibit varied grain-size particles and sandy mud to muddy texture. The core located in the upper slope region has recorded upto 60% of sand during the early Holocene. In general, sand content is higher in the shallow water core than in the cores of the deeper water, which exhibit silty clays and clayey silts. At the core site in the upper slope, gradual increase of silt with gradual decrease of clay was recorded. In the deeper slope region, the increased amount of clay compared to silt content was recorded. The currents although play a significant role in transport of fine sediments along and across the shore, which have influenced the deposition of silts and clays, the role of sea level in the sediment distribution on the sea floor is not ruled out. Sea level variations may have influenced both the sediment texture and rates of deposition. As the sea level was -120 m during the LGM, the exposed continental shelf was wider and subsequently the river discharges are much far away from the present coastline. This might be the main reason for the coarse sand fraction at the core site in the upper slope region. Further, the enhanced hydrolysis on land due to intense monsoonal precipitation tends to enrich the coarse fraction of the sediments on the sea floor.

Organic matter reflects paleoproductivity and sediment dispersal and deposition. The organic carbon was higher in the cores of the deeper region than in the sediments of the upper slope. The oxidation and preservation of organic carbon has been widely debated and has been extensively employed as a paleoproductivity index. In general, increased sedimentation rate enhances organic carbon preservation in sediments. It is suggested that locations of organic carbon maxima on continental slopes are not controlled by the bottom water oxygen levels, but by a combination of factors viz., the sediment texture, dilution of organic matter by other sediment components and the settling fluxes of organic carbon. $CaCO_3$ content varied from 11-57 % in the upper slope region whereas it varied from 6-26 % in the deep water region. $CaCO_3$ has also been extensively employed as a proxy for productivity. It has been debated whether dissolution or dilution by terrigenous material influences the $CaCO_3$ content in the oceans. The $CaCO_3$ records in the Arabian Sea are mostly influenced by terrigenous dilution, caused by variations in the terrigenous lithogenic flux during the summer monsoon seasons. The similar dilution of $CaCO_3$ in the southeastern Arabian Sea is caused by the terrigenous lithogenic flux derived from the southwestern part of India and exposed continental shelf during the LGM.

Clay mineral studies indicate that smectite, illite, kaolinite and chlorite are the dominant clay minerals in the southeastern Arabian Sea. The temporal distributions reveal that smectite is dominant through out the dated ages of the sedimentary layers. The clay mineral ratios are varied during early-middle-late Holocene and last glacial period. The study shows that K/C ratio is lower from middle Holocene to present and higher during the early Holocene in the upper slope region. K/I ratios are slightly higher from the middle to late Holocene period in both upper slope and deeper slope regions. Most of the previous studies, except Thamban et al. (2002), have recorded higher smectite content in the southeastern Arabian Sea. Abundance of smectite in the present study area suggests its derivation from the erosion of basic volcanic rocks of central India under semi-arid climate, and transported southward in the shelf and slope regions by strong monsoonal currents. Kaolinite and illite reflect their derivation from the Precambrian crystalline rocks associated with the hinterland. It is suggested that sediment derived from the southwestern India is constantly supplying kaolinite and illite to the continental slope. Further, deposition by gravitational currents is also an important mechanism in the dispersal and deposition of clay minerals in the continental margin region.

The climate variations during the late Quaternary might have also influenced the clay mineral distribution in the southeastern Arabian Sea. The variations of sea level during glacial/interglacial periods might have induced changes in clay mineral composition. The clay mineral ratios, K/C and K/I, serve as indicators for continental humidity, which indicate the distinct events of monsoon intensification. The characteristic high rainfall and high temperature in southwestern region of India would have led to strong hydrolisation of illite. The last glacial period as recorded in the cores is characterized by significant contribution of weathered products, indicating relatively arid conditions prevailed on the hinterland. Kaolinite content and ratios of K/C and K/I, proxies of continued humidity, indicate distinct events of monsoon intensification.

The temporal variations documented in clay minerals in the present study show that kaolinite, chlorite and illite levels show an oscillating trend. Smectite levels are constant through out, kaolinite and chlorite levels are high since 6 ka BP, while illite is higher than both kaolinite and chlorite during 19.5 – 6.8 ka BP. High ratios of kaolinite/chlorite from 19.5–6.3 ka BP suggest the prevalence of humid conditions. Low K/C ratios since Mid-Holocene to the present indicate reduced monsoonal activity and consequently low weathering rates. Chlorite forms under arid conditions and high chlorite/illite ratio from 6.3 ka BP to the Present provide evidence for the extent of aridity. Generally, illite and chlorite form under dry/arid conditions and kaolinite under humid conditions. The gradual decrease in illite and increase in kaolinite from 17.4 ka BP suggest that the climate has gradually turned to warm humid conditions since then in this region. Based on clay mineral proxies, Gingele et al. (2004) have inferred more humid conditions between 11 and 6 ka BP and the onset of more arid conditions around 5.5 ka BP, which reached a maximum at 3.6 ka BP. It appears that these arid conditions are responsible for the low sediment supply in the last 3 ka in the southwestern margin of India.

Clay mineral abundances and ratios express the intensity of weathering and hydrolysis on the adjacent landmass. The temporal distribution of clay minerals is controlled by the monsoon intensity, hydrolysis, sea level changes and climate of the region. High ratio values of clay minerals, particularly K/C and K/I suggest strong humid conditions since last glacial period. Monsoon intensity, southerly coastal currents and humid climate played a major role in the clay mineral distribution and abundance. The study strengthens the idea of employing the clay minerals as proxies to decipher paleomonsoon, paleoclimate and paleoenvironmental conditions during the Late Quaternary.

Abundances of major elements Al, Fe and Ti in cores AAS 38-4 and 38-5 are employed to investigate the changes in supply and source of detrital material to the eastern Arabian sea during the last glacial and the Holocene. The terrigenous records of cores reveal that the most significant, well-resolved environmental variations are evidenced during the early Holocene, associated with abrupt oscillations in the terrigenous signals.

The Al concentrations in sediments reflect the provenance and dilution effects. The low values of Al (~10 %) in sediments of the cores under present investigation is an intriguing factor because the hinterland consists of laterites and ultramafic rocks which are enriched in Al content. The low Al values in the eastern Arabian Sea sediments could be a dilution with lithic components that are low in Al content.

The Ti content in the core samples indicates a slightly higher content during the last glacial period than in the Holocene. Higher concentrations of Al and Ti coincide with the low $CaCO_3$ values during the last glacial period suggest greater terrigenous dilution during LGM. It has been suggested earlier by many that the $CaCO_3$ records from the Arabian Sea are mostly influenced by terrigenous dilution, caused by variations in the terrigenous lithogenic flux derived from the Arabian and Somalian Peninsulas during the summer monsoon season (Murray and Prell, 1992). The similar dilution of $CaCO_3$ in the southeastern Arabian Sea is caused by the terrigenous lithogenic flux derived from the southwestern part of India, when the sea level was low and during the LGM.

Trace elements of marine sediments provide important information on paleoproductivity, past oceanographic and climatic changes. Because these changes cannot be determined directly, geochemical proxies of climatic variations and paleoenvironmental conditions have been developed intensively during the last two decades. A large number of geochemical proxies, including major and trace elements, have been used to reconstruct the history of the productivity record. Sr/Al and Ba/Al ratios show an abrupt decrease during the early Holocene (8-10 ka BP), thereby indicating a sharp decrease in productivity during this period. This time interval coincides with the early Holocene humid interval, a period of intensified southwest monsoon in the Arabian Sea. A possible explanation for the observed sharp change during the early Holocene could be increased monsoonal precipitation on coastal regions, resulting in large amounts of fresh water flux to the core sites, and as a consequence upwelling-induced surface productivity would have weakened/ceased.

The use of Ba or Ba/Al ratios as a paleoproxy has been extensively documented. Enrichment of productivity-related elements - Ba, V, Sr, P - during interglacial times has been recorded in the present study.

The major and trace elements are employed to infer paleoclimate and paleoproductivity in the southeastern Arabian Sea. All the major element abundances show distinct variation between the Holocene and the last glacial period. Si, Ti, Al and Fe are distinctly lower during the Holocene than in the last glacial period. Trace elements - V, Co, Ni and Ba - are distinctly low during mid– and late Holocene. Trace elements/Al ratios suggest that there are distinct temporal variations in all the three cores. The variation in the input of terrigenous material to the Arabian Sea relates to monsoonal activities. Al concentrations in sediments reflected the provenance and dilution effects. High Al and Ti values coincide with the low $CaCO_3$ values during the last glacial period, which suggests greater terrigenous dilution. The elemental/Al ratios show low surface productivity during the LGM. An abrupt decrease of Sr/Al and Ba/Al ratios during the early Holocene indicate sharp decrease in productivity. Ba/Al ratio as a paleoproxy has been signified in this study too. Productivity related elements have been enriched during interglacial times. Most of the productivity indicated elements are lowered during the Holocene than in the last glacial period.

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