CHAPTER 1
Introduction

1. Climatic Variations

The Earth has been experiencing climatic variations driven by various factors forced upon depending on historical to geological time scale. The most significant climatic change the Planet Earth undergone during the geological time scale is the Ice Ages. During Ice ages the global temperatures are relatively lower and high latitude regions of Earth is covered by massive ice sheets. During the history of the Earth at least five major Ice Ages have occurred with time intervals varying from million to tens of millions years. The earliest Ice age was occurred around 2 billion years ago and the most recent was occurred about 2 million years ago i.e. during Quaternary Period. Ice Age during Quaternary period is characterized by several short warm and cold periods that occurred with certain repeated cycles. The multiple shorter warmer periods when ice sheets retreat are termed as interglacials or interglacial cycles and multiple shorter cold periods when ice sheets advance are called glacial or glacial cycles.

Periodic advances and retreats of ice sheets and associated temperature shifts have left a signature in records of stable oxygen isotope ratios collected from both ice and ocean sediment cores. Oxygen isotopic records display a 100,000 year cyclical transition from glacial to interglacial conditions associated with changes in the physical relationship between the Earth and the Sun (Hays et al., 1976). The periodicities of Eccentricity (Earth to Sun distance), Obliquity (tilt of Earth's axis) and precession of the equinox are evident in multiple proxies which are used to study the climatic variations on glacial and interglacial time scales (Clemens et al., 2003).

Way back in eighteenth century, Serbian Mathematician Milutin Milankovitch developed mathematical formula upon which these orbital variations are based hence these cycles are named as Milankovitch Cycles.

Proxy measurements of polar temperature from the Greenland Ice Sheet Project (GISP) ice core, indicate that the glacial climate was punctuated with temperature swings of up to 10°C, which are superimposed on the glacial/interglacial sequence (Dansgaard et al., 1993). These sub-orbital or millennial-scale events, known as Dansgaard-Oeschger (D-O) events have a frequency of ~1500 years (Bond...
et al., 1993). This discovery of very rapid fluctuations in the earth’s climate, occurring on sub-orbital timescales, has drawn attention to the need for further exploration into the mechanism that is forcing climate change with a periodicity not explained by the Milankovitch orbital cycles. The rapidity of the D-O event transitions from warm “interstadial” periods to cool “stadials” brought about a paradigm shift in the field of climate change. Numerous evidences demonstrate that tropical monsoon system plays an important role on climate through air-sea interaction (Webster, 1987). Details of the monsoon are discussed in the following section.

2. Indian Monsoon

During the Northern Hemisphere summer high solar radiation causes intense sensible and latent heating over central Asia, especially over the Tibet Plateau, which causes an ascending air flow to the atmosphere leading to development of an intense low pressure cell in the Central Asia (Fig. 1.1). Simultaneously a high pressure cell develops in the southern Indian Ocean due to Southern Hemisphere winter and the greater specific heat of the water. Thus an atmospheric pressure gradient between the Asian continent and southern Indian Ocean induces a large scale meridional overturning forcing the Inter Tropical Convergent Zone (ITCZ) over Indian subcontinent causing 80% of annual rainfall in the region (Hastemath and Lamb, 1979). During winter the Asian sector is characterized by low solar radiation, cold temperatures sets in NE winds, which flow from the cold Asian Continent to the Arabian Sea (Fig.1.2).

![Fig.1.1. Low Pressure on the continent during Summer Monsoon.](image)

![Fig.1.2. High Pressure on the Continent during Winter Monsoon.](image)
These continental winter winds are characterized by relatively low velocity and therefore carry little moisture resulting in lesser rainfall during NE monsoon in the Indian subcontinent (Hastemath and Lamb, 1979).

The Indian monsoon (SW and NE monsoons) exerts a strong influence on the climatic conditions in the Asian subcontinent. The Indian monsoon affects millions of people across the Indian subcontinent, being responsible for both droughts and floods across this vast region. An understanding of the short-term climatic variability which causes these fluctuations in monsoon rainfall is poorly constrained, despite being of great importance. Monsoons not only influence the socio-economics of the south-east Asia but also play an important role on the biogeochemical cycles of the Indian Ocean. Therefore, understanding the forcing mechanisms of monsoon at various time scales would enhance the reliable forecasting of monsoon rainfall in the Indian subcontinent. Hence, reconstructing the history of the Indian Monsoon is a prerequisite for developing an understanding of past, present and future monsoons.

The influence of monsoon on the biogeochemical cycles also varied in different time scales, which depends on the several feed backs of the climate system. The intense monsoon winds, induced by the temperature gradient between the differential warming of the Asian continent and the Indian Ocean, leave an imprint in the Arabian Sea sediments which reflect changes in seasonal sedimentation and primary production. Expedition of Ocean Drilling Program Leg 117 along the western Arabian Sea enabled the paleoceanography community to study the monsoon variability and biogeochemistry on glacial and interglacial time scales (Clemens et al., 1991; Shimmield et al., 1991).

3. Oceanography of the Arabian Sea

Land locking on three sides of the Indian Ocean makes it a unique ocean as ocean arrests the north poleward oceanic heat transport, which results in accumulation of a massive pool of warm waters (28°C) in the surface layers. Hence the occurrence of seasonally reversing monsoons (Hastemath, 1994), transient pre summer and post summer monsoon cyclonic storms, is a natural regional geophysical necessity to drain this accumulated heat energy through intense air-sea interaction and energetic internal...
oceanic redistribution processes to accomplish thermal equilibrium over an annual cycle.

Arabian Sea is characterized by the seasonal reversal of winds during the SW and NE monsoons (Wyrtki, 1971). Variations of seasonal circulation patterns, seasonal sea surface temperature, sea surface salinity and productivity during SW and NE monsoons are described in the following section.

3.1. Southwest Monsoon

3.1.1. Surface Circulation

During the SW monsoon the surface low level south-easterly trade winds of the Southern Hemisphere extend southerly or south-westerly in the Northern Hemisphere. The frictional stresses of these south-westerly winds in turn drive the Somali Current (SC) flowing northward as western boundary current, the westward flowing South Equatorial Current (SEC) and eastward flowing Monsoon Current (MC) (Wyrtki, 1973). The South Equatorial Current (SEC) does not shift much to the north, but it becomes stronger, and most of its water turns north into the Somali Current (SC) off Sumatra, the Monsoon Current (MC) crosses the Equator and turns into the South Equatorial Current (SEC). These 3 currents, monsoon current (MC), South Equatorial Current (SEC) and Somali Current (SC), form a very strong wind driven gyre in the equatorial Indian Ocean (Fig. 1.3) (Wyrtki, 1973).

3.1.2. Sea Surface Temperature

In the Arabian Sea SST varies from 24°C to 30°C during SW monsoon (Fig. 1.4). Coldest temperature prevails in the western Arabian Sea and warmest temperatures in the Lakshadweep Sea, which is the core of the Indian Ocean warm Pool. During boreal summer strong SW monsoon winds blow over the Arabian Sea, these winds induce coastal upwelling along the Somalia and Oman Margin, which cools the western Arabian Sea and develops 4°C temperature gradient between west to east (Fig.1.4). On an annual cycle, mean SST in the western Arabian Sea is 2°C cooler than in the eastern Arabian Sea (Levitus and Boyer, 1994).
Fig. 1.3. A schematic representation of identified current branches during the Southwest Monsoon. Current branches indicated are the South Equatorial Current (SEC), South Equatorial Countercurrent (SECC), Northeast and Southeast Madagascar Current (NEMC and SEMC), East African Coast Current (EACC), Somali Current (SC), Southern Gyre (SG) and Great Whirl (GW) and associated upwelling wedges, Socotra Eddy (SE), Ras al Hadd Jet (RHJ) and upwelling wedges off Oman, West Indian Coast Current (WICC), Laccadive High and Low (LH and LL), East Indian Coast Current (EICC), Southwest and Northeast Monsoon Current (SMC and NMC), South Java Current (JC) and Leeuwin Current (LC). (Schott and McCreary, 2001).

Fig. 1.4. Sea Surface Temperature during South West Monsoon in the Arabian Sea.
3.1.3. **Sea Surface Salinity**

In the Arabian Sea salinity varies from 36.5%o to 35%o during SW monsoon (Figure 1.5). A distinct North to South salinity gradient exists with highest salinity 36.5%o in the Northern Arabian Sea and lowest salinity of 35%o along the western continental margin of India. The influx from Red Sea and Persian Gulf waters and more evaporation cause high salinity in the northern Arabian Sea. The low salinity is a result of high precipitation and river discharge along the western continental margin of India. The upwelling of water off Somalia causes the salinity to be lesser along the Somali Coast.

![Sea Surface Salinity Climatology in the South-West Monsoon (June, July, August and September) in the Arabian Sea and western equatorial Indian Ocean Region](image)

**Fig. 1.5. Sea Surface Salinity during South West Monsoon in the Arabian Sea**

3.1.4. **Productivity**

The Arabian Sea is known to be an area of high productivity region in the world oceans. During the SW monsoon strong SW winds blow across the Arabian Sea, which cause Ekman transport of surface waters away from the coast and develops intense seasonal upwelling cells along the Coast of Somalia, Oman and the southwest coast of India (Wyrtki, 1973). Upwelling processes increases the nutrient
concentrations (silicate, nitrate and phosphate) in surface waters which foster increased primary and secondary productivity (Fig. 1.6). Further, strong surface currents and eddies create a heterogeneous environment (Schott et al., 1990) that further enhances local biological productivity. Strong SW monsoon winds deepen the mixed layer in the central Arabian Sea which also increases the productivity in the region (Banse, 1987). In the north-western Arabian Sea near Oman, the flux of organic carbon reaches 80mgC/m²/day during the summer monsoon season (Honjo et al., 1999).

Fig. 1.6 Productivity during Southwest Monsoon
(Courtesy NASA/GSFC)

http://www-d.as.uwyo.edu/~geerts/cwx/notes/chap11/phyto.html
3.2. North East monsoon

3.2.1. Surface circulation

During the NE monsoon water movements to the north of the Equator are from east to west, forming the NE monsoon current. This flow starts developing in November, reaches its greatest strength in February and subsides in April. Current velocities are usually strongest to the south of Ceylon and in the southern Arabian Sea, where they exceed 1 knot. From November to January a strong branch of this current turns north and flows along the west coast of India, carrying low salinity water from Bay of Bengal into the eastern Arabian Sea (Fig. 1.7). The NE monsoon drift appears to be rather shallow and seems to exert little influence on the waters below the thermocline. Off the Somalia coast most of its water turns south, crossing the

![Fig. 1.7. A schematic representation of identified current branches during the Northeast Monsoon. Current branches indicated are the South Equatorial Current (SEC), South Equatorial Countercurrent (SECC), Northeast and Southeast Madagascar Current (NEMC and SEMC), East African Coast Current (EACC), Somali Current (SC), Southern Gyre (SG) and Great Whirl (GW) and associated upwelling wedges, Socotra Eddy (SE), Ras al Hadd Jet (RHJ) and upwelling wedges off Oman, West Indian Coast Current (WICC), Laccadive High and Low (LH and LL), East Indian Coast Current (EICC), Southwest and Northeast Monsoon Current (SMC and NMC), South Java Current (JC) and Leeuwin Current (LC). (Schott and McCreary, 2001).](image-url)
equator and forming the Equatorial Countercurrent. The southward flow off Somalia is strongest in December, but lasts until March (Wyrtki, 1973).

### 3.2.2. Sea Surface Temperature

During NE monsoon SST varies from 26.5°C to 28°C with a weak NW-SE gradient in the Arabian Sea (Fig. 1.8). Unlike off the coasts of Somalia, Oman and southwest India, upwelling does not occur during the SW monsoon season in the NE Arabian Sea hence SST are warmer during the SW monsoon (June through September) and coolest SST are noticed during January through March due to winter cooling in the NE Arabian Sea.

![Fig. 1.8. Sea Surface Temperature during North East Monsoon in the Arabian Sea.](image)

### 3.2.3. Sea Surface Salinity

In the Arabian Sea salinity varies from 36.5‰ to 35‰ with a North to South gradient during NE monsoon (Fig.1.9). High salinity in the northern Arabian Sea is caused due to high amount of evaporation during winter. The low salinity of 35‰ in the South is due to the North equatorial current which brings in low salinity water.
from Bay of Bengal in to the Arabian Sea. Low salinity water mass tongue is more prominent during the NE monsoon along the south-west coast of India.

![Sea Surface Salinity Climatology in the North-East Monsoon (October, November, December and January) in the Arabian Sea and western equatorial Indian Ocean Region (5°S- 30°N, 45°E- 80°E)]

Fig. 1.9. Sea Surface Salinity during North East Monsoon in the Arabian Sea.

3.2.4. Productivity

Reversal of wind directions from continent to the ocean during NE monsoon causes an onshore Ekman transport of surface waters, which suppresses upwelling and lowers the productivity in the western and southern Arabian Sea (Fig. 1.10). But the northern Arabian Sea continues to sustain fairly high biological production after the upwelling season and during much of the winter (north-east) monsoon (Banse and McClain, 1986, Banse and English, 1993, Yentsch and Phinney, 1992). The dry, cool, winter monsoon winds that originate in the Himalayas cools the surface water and increase the evaporation in the NE Arabian Sea causing convective mixing (Madhupratap et al., 1996; Reichart et al., 1998). As the mixing occurs the subsurface nutrient rich water comes up and increases the productivity.
4. Oxygen Minimum Zone

About 2/3 of the global continental margin area are exposed to severely \( O_2 \) depleted waters (<0.2ml-1) (Fig.1.11). Among these, western continental margin of India is one where pronounced oxygen depleted water conditions exist between 150 to 1200 m water depth (Wyrtki, 1973; Naqvi, 1991 and references therein). Three combined factors i.e. 1) the high seasonal productivity and associated high input of organic matter and part of its subsequent oxidation within and below the euphotic zone, 2) the sluggish intermediate water conditions, and 3) the strong tropical-subtropical thermocline, prevent downward mixing of oxygenated surface water causing the development of intense oxygen minimum zone (OMZ). The OMZ in the north-eastern Arabian Sea is thicker (Morrison et al., 1999) and more intense than
along the Somali coast and reaches values between 0-0.5 ml Oxygen/l between 200-1200 m water depth (Schulte et al., 1999).

Low dissolved oxygen levels and Hypoxia in the ocean influence biogeochemical cycling of elements, the distribution of marine species and economic well being of many coastal countries. Furthermore, global warming and associated stratification may lead to lowered oxygen content of the world oceans (Keeling and Garcia, 2002), and expansion of OMZs in selected areas. OMZ expansion or shrinkage may promote the evolution of species and genetic diversity maxima at mid-slope depths (Jacobs and Lindberg, 1998; Etter et al., 1999; Ulloa et al., 2001). The extent and severity of OMZs will change with alteration of ocean circulation, temperature and productivity (Reichart et al., 1998; Keeling and Garcia, 2002).
5. Previous research related to the objectives of the present study

5.1. Sea surface temperature reconstructions in the Arabian Sea

The oceanic surface heating is a result of the combination of several factors such as net solar insolation, air–sea turbulent heat fluxes, optical transparency of the near-surface water column, wind stress, mixed-layer depth, advection by eddies and ocean currents (Rao et al., 1993). The heat transfer between the atmosphere and ocean can be examined through the variations in the sea-surface temperature (SST), the most important variable that affects, for example, how much rain the monsoons will bring to India (Rao, 1986). Therefore, SST variations in the Arabian Sea play an important role on the prediction of SW monsoon onset and progress over India (Kershaw, 1988; Rao and Sivkumar, 1999).

Knowing the importance of SST in forecasting the Indian Monsoon Rainfall efforts have been made by the paleoceanographic community to reconstruct SST on glacial and interglacial time scales by using transfer function techniques which are based on the planktonic foraminiferal faunal census data (CLIMAP Project Members, 1976). Such estimates show that the average Ice Age zonal SST for the entire Indian Ocean was 1.4°C cooler during February and 1.5°C cooler during August compared with modern SST (Prell and Hutson, 1979) and Arabian Sea document greater seasonal SST changes during the last glacial period than in Holocene (Naidu and Malmgren, 2005). Arabian Sea SST estimates based on alkenone studies reveal that last glacial maximum temperatures were about 2°C cooler compared to modern values in the Arabian Sea (Rostek et al., 1993; Emeis et al., 1995; Chodankar et al., 2005). Attempts have been made to trace the upwelling strength based on the SST differences between glacial and interglacials mainly derived from oxygen isotopes and alkenones (Zahn and Pedersen, 1991; Emeis et al., 1995). These studies did not find any systematic relationship between SST changes and upwelling strength on glacial and interglacial time scales, because these authors dealt with annual SST rather than seasonal ones.
Over the last decade our understandings of SST variations in the Indian Ocean during Late Quaternary have improved considerably by the advent of Mg/Ca thermometry. SST reconstructions on glacial and interglacial time scales from the equatorial Indian Ocean exhibit similar patterns as the equatorial Pacific Ocean, which revealing a common forcing which controls the SST in both the regions (Saraswat et al., 2005). Dahl and Oppo (2006) reconstructed the SST variations in a set of sediment cores covering the entire Arabian Sea for four time intervals i.e. 0 ka, 8 ka, 15 ka and 20 ka. Their study reveals that at 20 ka and 15 ka Arabian Sea exhibits average negative temperature anomaly of 2.5°C - 3.5°C, which is attributed to the influences of glacial atmospheric CO$_2$ concentrations and large continental Ice sheets. However, it is apparent that Arabian Sea was ~3 to 4°C cooler during the Last Glacial Maximum than the present day (Saher et al., 2007; Anand et al., 2008; Govil and Naidu, 2010; Saraswat et al., 2013). SST reconstructions based on the geochemical proxies (alkenones and Mg/Ca) is expected to represent the annual SST changes, therefore to reconstruct the seasonal SST changes which are highly important in the Arabian Sea one need to depend on the faunal based SST reconstructions. Among the various faunal based SST reconstruction techniques Artificial Neural Network (ANN) provides precise SST estimates (Malmgren and Norlund, 1997). Thus, in the present study we have chosen ANN technique to reconstruct the seasonal SST changes in the Arabian Sea to unravel the seasonal SST changes in the western and eastern Arabian Sea.

5.2. Reconstruction of Carbonate ion

In today's ocean, marine organisms secrete calcitic hard parts at a rate several times faster than CO$_2$ being added to the ocean-atmosphere system (via planetary outgassing and weathering of continental rocks) (Broecker, 2003). While the state of saturation in the ocean is set by the product of the Ca$^{2+}$ and CO$_3^{2-}$ concentrations, calcium has such a long residence time ($10^6$ yr) that, at least on the timescale of a single glacial cycle ($\sim 10^5$ yr), its concentration can be assumed to have remained unchanged (Broecker, 2003). In contrast, the dissolved inorganic carbon in the ocean is replaced on a timescale roughly equal to that of the major glacial to interglacial cycle ($10^5$ yr). But, since in the deep sea CO$_3^{2-}$ ion makes up only ~5% of the total dissolved inorganic carbon, its adjustment time turns out to be only about one-
tenth that for dissolved inorganic carbon \(\sim 5,000\) yr (Broecker, 2003). Hence, the concentration of \(\text{CO}_3^{2-}\) has gradients within the sea and likely has undergone climate-induced changes. These changes involve both the carbonate ion concentration averaged over the entire deep ocean and its distribution with respect to water depth and geographical location. It is the global average carbonate ion concentration in the deep sea that adjusts in order to assure that the burial of \(\text{CaCO}_3\) in the sediments matches the input of \(\text{CO}_2\) to the ocean atmosphere system. As part of the Geochemical Section Studies (GEOSECS), Transient Tracers in the Ocean program (TTO), South Atlantic Ventilation Experiment (SAVE) and World Ocean Circulation Experiment (WOCE) ocean surveys, \(\Sigma\text{CO}_2\) and alkalinity measurements were made on water samples at various depths from the world oceans. Given the depth, temperature, salinity and phosphate it is possible to compute \textit{in situ} carbonate ion concentrations. Taro Takahashi from Lamont Doherty Earth Observatory has played a key role in these measurements and conversion to \textit{in situ} carbonate ion concentrations. A complete picture of the \(\text{CO}_3^{2-}\) ion concentrations in the deep sea is now available. Below 1,500m in the world ocean, the distribution of carbonate ion concentration is remarkably simple. For the most part, waters in the Pacific, Indian and Southern Oceans have concentrations confined to the range \(83 \pm 8\mu\text{mol kg}^{-1}\) (Takahashi, 2001). The exception is the northern Pacific, where the values drop to as low as \(60\mu\text{mol kg}^{-1}\). In contrast much of the deep water in the Atlantic has concentrations in the range of \(112 \pm 5\mu\text{mol kg}^{-1}\) (Takahashi, 2001). Attention is therefore now focused on distribution of \(\text{CO}_3^{2-}\) concentration in the deep sea for it alone sets the depth of the transition zone.

### 5.2.1. \textit{Paleocarbonate ion Proxies}

Calcite dissolution in marine sediments is driven by saturation state of the overlying waters and/or responds to sedimentary organic matter respiration and the acidification of pore waters that results from that (Emerson and Bender, 1981; Archer and Maier-Reimer, 1994). Specifically, differences in the carbonate ion concentrations of bottom water are believed to be responsible for the first-order variations in the depth of the lysocline between and within oceans (Peterson and Prell, 1985).
Attempts to reconstruct the carbonate ion history from sediments of the world oceans amongst others have yielded two important proxies; the size index (Broecker and Clark, 1999) and the shell weight method (Lohmann, 1995). Both these proxies have been applied largely to the Atlantic Ocean (Broecker and Clark, 1999; Broecker et al., 1999 and references therein), the Pacific Ocean (Broecker and Clark, 2003), the Indian Ocean (Broecker and Clark, 1999) and the Caribbean Sea (Broecker et al., 2003). Largely, in the above studies the shell weights of selected planktonic foraminifer species has been successfully utilized in understanding the carbonate ion variations during the Holocene and the Last Glacial Maxima (LGM).

The size index and planktonic foraminifera shell weight have also been employed to discuss calcite dissolution above the lysocline in the Atlantic, Pacific and Indian oceans (de Villiers, 2005; Schulte and Bard, 2003). Though these studies could quantify the CO$_3^{2-}$ concentrations to some extent, they could certainly identify the calcite dissolution and preservation events (Broecker et al., 2003). Recently, shell weights have been used to understand the CCD variations and hence carbonate ion over the course of the Cenozoic (Broecker, 2008). However, the paleocarbonate ion studies in the Indian Ocean are very sparse compared to more rigorously studied Atlantic and Pacific oceans.

5.3. Monsoon productivity and OMZ variability

The thermocline waters of the Arabian Sea are severely depleted in oxygen, which has direct bearing on the biogeochemical budget in the region. Quite a few studies in the Arabian Sea based on benthic foraminifera (Hermelin, 1991; Carbonate dissolution (Ten Kate, 1994), nitrogen isotopes and organic carbon (Altabet et al., 1995; Reichart et al., 1998; Ganeshram, et al., 2000), redox elemental fluctuations (Pattan et al., 2003), pteropod abundance and aragonite content (Naidu et al., 2014) have documented significant variation of OMZ intensity, which was primarily controlled by the productivity changes related to upwelling and convection processes. An excellent study by Reichart et al. (1998) from the north-western Arabian Sea reveals that the fluctuation of OMZ was related to the Earth’s Orbital cycles of Eccentricity, Obliquity and Precession. However, the millennial scale variability of the denitrification was tightly coupled with the Greenland temperature changes (Altabet et al., 2005). During stadials the ventilation of Antarctica Intermediate


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Waters also enhances the dissolved oxygen levels in the Arabian Sea (Boning and Bard, 2009). Though denitrification records in the Arabian Sea document a regional pattern (Altabet et al., 1995; Reichart et al., 1998), it is not clearly understood the interplay between monsoon-induced productivity, OMZ and aragonite and calcite preservation along the western continental margin of India on millennial time scale. Therefore, the present study is aimed to address these issues.

5.4. **NE Monsoon and Winter cooling**

As outlined above, winter cooling play an important role on the productivity in the NE Arabian Sea. The SST distribution and convection induced nutrient enrichments are governed by the anticlockwise circulation caused by NE monsoon winds. Thus the strength of NE monsoon winds and winter cooling are tightly coupled in NE Arabian Sea. Reconstruction of SW monsoon variability by using numerous proxies from several sediment cores covering various time intervals were carried out from the Arabian Sea and Bay of Bengal (Duplessy, 1982; Anderson and Prell, 1993; Overpeck et al., 1996; Naidu and Malmgren, 1996; Anand et al., 2008; Govil and Naidu, 2010; Govil and Naidu, 2011 and references therein). No information is available on the variability of NE monsoon on glacial and interglacial time scales and also on millennial time scales, but expect that NE monsoons were stronger during last glacial maximum than in Holocene (Duplessy, 1982). In this study an effort has been made to understand the winter convection/cooling influence by reconstructing winter SST over the last 40 kyr.

6. **Proxies used in the study**

6.1. **Planktonic foraminifera**

Foraminifera belong to Protista kingdom. Foraminifers are marine microorganism with a hard outer skeleton, mainly made up of calcium carbonate (CaCO$_3$). Planktonic foraminifera build calcite exoskeleton (tests or shells), of which the physical and chemical composition reflects the sea water conditions in which they are formed and floats on surface water often migrating from 0-200m. After the life cycle (3 to 4 weeks) many of these skeletons pile up on the sea floor sediment.
Distribution pattern of various modern planktonic foraminifera represent the surface water hydrography (Be, 1960; Zhang, 1985), water mass properties (Be and Tolderlund, 1971), and upwelling (Kroon and Ganssen, 1989) depending on the processes which are dominant in the particular region. Therefore, planktonic foraminifera serve as a main tool in various paleostudies. There are about 30-40 recent planktonic foraminiferal species and each species has its preference regarding depth, season and food source (Hemleben and Spindler, 1983).

Although fossil shells were recognized from beach sands and deep sea sediments as early as 1826 (d’Orbigny, 1826; Parker and Jonnes, 1865), it was until Owen (1867) and the scientific results of the challenger expedition (Brady, 1884) that the planktonic life habitats of these marine protozoans was clearly established. Applications of planktonic foraminifera in the field of paleoceanography and paleoclimate substantially increased the pioneering work of Schott (1935). Since then several researchers have been using planktonic foraminifera as a tool to study the paleoceanography in the world oceans. For example the relative abundance of *Globigerina bulloides* has been used to reconstruct monsoon upwelling history in the Arabian Sea (Prell, 1984; Naidu and Malmgren, 1996), Foraminifer assemblages, foraminifer sizes, coiling directions and other morphometric variations have been successfully used to address the various paleoceanographic problems in the India Ocean (eg, Prell, 1984; Kroon and Ganssen; 1989; Naidu and Malmgren, 1995; Peeters et al., 2004; Nigam, 2005).

For the last four decades planktonic foraminifera faunal assemblages have been used to estimate the sea surface temperature, for example transfer functions (Imbrie and Kipp, 1971), Modern Analogue Technique (Prell, 1985). More recently Artificial Neural Network (ANN) has been used to compute the seasonal SST in the Indian Ocean (Naidu and Malmgren, 2005). The primary principle behind such estimates lie in the empirical calibration between planktonic foraminifera faunal assemblages from core tops and modern measured SST from the same location. The most important property of a calibration data set is its coverage both in terms of geographic area and the range of SST that is calibrated. By applying the empirical calibration established between core top planktonic foraminifer faunal assemblages and measured SST to the planktonic foraminifera faunal assemblages in a sediment core, one can derive the SST of the past. Based on the ANN technique more precise...
SST could be computed better than transfer function technique and Modern Analogue Technique. Several statistical techniques, including the transfer function technique (Imbrie and Kipp, 1971), the modern Analogue Technique (Prell, 1985) and the Artificial Neural Technique (ANN) (Naidu and Malmgren, 2005) have been used to estimate annual, seasonal and monthly SST. The latter technique has at least the same potential for providing accurate SST as the Modern Analogue technique (Malmgren et al., 2001) and we used this technique to reconstruct SSTs for the months of May and August.

6.2. Oxygen isotopes

Oxygen has three stable isotopes $^{16}$O, $^{17}$O, and $^{18}$O. The most common isotope is $^{16}$O which constitutes approx 99.7% of all oxygen of the earth and another stable isotope that is often used in paleoclimate studies is $^{18}$O which constitutes about 0.2% and least abundant (0.04%) stable isotope of oxygen is $^{17}$O. These isotopes have slightly different physical properties, for instance $^{16}$O evaporates faster than $^{18}$O from the water and when they fall out as rain $^{18}$O is released faster than $^{16}$O. These phenomenon has several effects as rain clouds lose their $^{18}$O first and with continuing rain the precipitation becomes increasingly richer in $^{16}$O (Rozanski et al., 1992). This process leads to rainfall being increasingly depleted in $^{18}$O further from source and becomes more depleted further inland, therefore, precipitation in the Polar Regions are depleted with $^{18}$O and enriched in $^{16}$O. Thus oceans become depleted with $^{16}$O during glacial because most of the $^{16}$O gets locked up in the ice sheets when the ice caps are maximum in size. Measuring the $^{18}$O/$^{16}$O ratios in sea water would provide information about how much ice was stored on the continents and how the ocean water is affected by the evaporation and precipitation balance and riverine influx.

Another important application of stable oxygen and carbon isotopes in the field of paleoceanography comes through the fractionation of oxygen and carbon isotopes between the calcium carbonate crystallization (i.e. foraminifera) and ambient sea water, the fraction processes is temperature dependent. If temperature increases the incorporation of $^{16}$O increases and $^{18}$O decreases and vice versa in foraminifera. Urey was the first to demonstrate that fractionation of oxygen isotopes in the carbonate-water system is a measurable function of temperature and suggested that
these can be useful as a geological thermometer. He noted that, "calcium carbonate organisms are in equilibrium with the water depth in which it lives, and the shell sinks to the bottom of the sea. It is only necessary to determine the ratio of isotope of oxygen in the shell today in order to know the temperature at which the organism lived" (Urey, 1948). Subsequent research at the University of Chicago (McCrea, 1950; Epstein et al., 1953) have steered the study of oxygen isotopes ratios in the planktonic foraminifera into the forefront of fields studying climate changes and ocean history.

6.2.1. Oxygen Isotope as a Stratigraphic Tool in Paleoceanography

Oxygen isotopic studies on planktonic and benthic foraminiferal species have been carried out on important cores from different parts of the world ocean (Shackleton, 1977). The study of oxygen isotopic ratios relative to PDB standard calcite shows the synchronous pattern through geological past. Variation in oxygen isotopic composition in the ocean sediment taken a consideration of mixing of the ocean water (less than 1000yrs) and as isotopic composition is controlled by the amount of water stored in the continent ice (Shackleton and Opdyke, 1973) suggest that the variation in the isotopic composition record are synchronous in ocean sediment from any region. Ice sheet extent and melting of glaciers affect the sea level globally; it plays a major role in consideration to establishing the chronostratigraphy based on marine sediment. These synchronous variations enable correlations to be made between cores that may be thousands of kilometers apart (Bradley, 1999; Pisias et al., 1984; Prell et al., 1986). Based on the isotopic signals from the marine sediments from all over the world ocean, universally recognizable isotopic stages can be identified (Bradley, 1999; Pisias et al., 1984; Emiliani, 1955, 1966). Nevertheless, even after establishing the chronology based on the isotopic composition, it is important to add the dating techniques to make the absolute chronostratigraphy by using the $^{14}$C radiocarbon, U-series dating and paleomagnetism. Many investigators have shown the correlation between the climatic variability based on the orbital tuning and in isotopic signals in marine sediment as a change in sedimentation rate (Hays et al., 1976; Kominz et al., 1979; Martinson et al., 1987) and created a well-controlled chronostratigraphy. Warmer periods (Interglacial on Interstadials) are assigned odd numbers and colder (glacial) periods are assigned even numbers.
Marine isotopic stages (MIS) have been extensively used to reconstruct the time frame of marine sediment cores from the world oceans (Prell et al., 1986; Bassinot et al., 1994). By using the orbital time scale tuning to the $\delta^{18}$O record it has been established the chronology of MIS up to 2.5Ma (Shackleton, et al., 1990).

Oxygen isotope values of foraminifera depend on the local variation of salinity and temperature and globally with variations in continental ice volume. The relationship between $\delta^{18}$O of foraminifera, oxygen isotopic composition of the original water and temperature is clearly shown by the empirical equation of Craig (1965) given below.

$$T = 1609 - 4.2 (\delta c - \delta w) + 0.13 (\delta c - \delta w)^2$$

$T =$ calcification temperature of water in which the organism lives

$\delta c =$ is the per mil ($\%$) difference between the samples carbonate and PDB standard

$\delta w =$ is the per mil ($\%$) difference between the $\delta^{18}$O of water in which the sample precipitated and the SMOW standard.

6.3. Carbon Isotopes ($\delta^{13}$C)

There are a variety of naturally-occurring isotopes of carbon. These isotopes are characterized by differing atomic weights resulting from varying numbers of neutrons in the atomic nuclei. The relative abundances of these isotopes are given below

$^{12}$C = 98.89%, $^{13}$C = 1.11%, $^{14}$C = 1E-10%

$^{12}$C and $^{13}$C are both stable isotopes and $^{14}$C radioactive isotope which undergoes radioactive decay.

Mostly in all oceans, a linear correlation is observed between $\delta^{13}$C$\Sigma$CO$_2$ values and nutrient contents of deep and bottom water because the distributions of both are controlled by the interaction of biological uptake at the sea surface and decomposition in deeper water masses with the general circulation of the ocean (Kroopnick, 1980; Kroopnick, 1985; Mackensen and Bickert, 1999). The biological and chemical
fractionation of carbon isotopes in the ocean provide one of the most useful tracers for reconstructing past distributions of water masses and their properties (Curry et al., 1988). The distribution of δ¹³C in ocean is controlled principally by photosynthesis and remineralisation of organic carbon, and by mixing between water masses of different isotopic composition. The value of δ¹³C in seawater, after primary producers have removed all nutrients, is controlled by the mean δ¹³C and the mean nutrient concentration of the ocean (Broecker, 1982; Broecker and Peng, 1982).

6.4. Geochemical Elements

The concentration of minor and trace elements in marine sediment infer the range of chemical, oceanographic and sedimentary control on their supply and distribution in the sediment and their removal from the Ocean (Calvert and Pedersen, 1993). The chemical composition of the marine sediment is controlled by the relative contributions of particulate matter derived from various sources having variable compositions, by the uptake or fixation of elements during sediment and subsequent diagenetic changes. Thus, the concentrations of various geochemical elements were used in paleostudies, for example Barium has been used a tracer of productivity (Schmitz, 1987; Shimmield et al., 1991), Aluminium and Titanium as a tracer of terrigenous input (Pattan et al., 2003, Ivanchko et al., 2005 and references therein).

Earlier research on conditions required for the precipitation of various elements reveal that "some minor and trace elements are precipitated where free dissolved sulphide is present (Cu, Cd, Ni, Zn) without undergoing a valency change, whereas others undergo change in valency and are either more efficiently adsorbed onto solid surfaces under oxic (I) or anoxic (V) conditions or precipitated in anoxic conditions (Cr, Mn, Mo, Re U, V) (Calvert and Pedersen, 1993). Therefore, the enrichment of these minor and trace elements have been used to reconstruct the oxic and anoxic conditions during the geological past (eg. Reichart et al., 1998; Pattan, et al., 2013 and references therein).
6.5. **Nitrogen isotopes ($\delta^{15}N$)**

The isotopic fractionation of nitrogen associated with denitrification enriches $^{15}\text{NO}_3^-$ as $^{14}\text{NO}_3^-$ is preferentially converted to gaseous products. As denitrification proceeds, the residual nitrate in the oxygen-deficit sub surface waters becomes progressively enriched in $^{15}\text{NO}_3^-$. A consistent pattern of heavy $\delta^{15}N$ and high accumulation rates of biogenic components during the Holocene and lighter $\delta^{15}N$ and low accumulation rates of biogenic components during glacial periods are noticed in the Arabian Sea (Altabet et al., 1995; Ganeshram et al., 2000).

6.6. **Calcium carbonate**

Fifty percent of world’s ocean bottom is covered by carbonate sediments (>30% calcium carbonate) (Lisitzen, 1972). In near shore environment and continental shelves benthic organisms such as mollusks, bryozoan, algae and echinoderms are the major source of carbonate accumulation. By contrast, carbonate production over continental slopes and in the abyssal plains is almost exclusively planktic, dominated by coccolithophores, foraminifera and to a minor extent pteropods and calcareous dinoflagellates. Three factors control the calcium carbonate contents in the marine sediments: productivity, dissolution and dilution by non-carbonate material (Damuth, 1975). Thus, calcium carbonate content in the sediment can be used to trace the dilution (Naidu, 1991; Sirocko et al., 1993), dissolution (Le and Shackleton, 1992) and productivity (Naidu and Malmgren, 1999).

6.7. **Organic carbon**

Unicellular phytoplankton and zooplankton are the largest sources of organic carbon in the marine environment. Primary producers living in surface waters convert CO$_2$ into the organic matter. Part of this organic matter sinks to the deeper waters where it is oxidized, releasing initially fixed carbon back to water as CO$_2$ (Broecker and Peng, 1993). The differences between CaCO$_3$ and organic carbon in different areas are because of differential biological productivity in surface waters at the investigated area and/or differential preservations/dilution of CaCO$_3$ and organic carbon (Ortiz et al., 2004). Indeed, these two mechanisms are not decoupled and higher organic carbon
may cause enhanced CaCO₃ dissolution (Emerson and Bender, 1981; Guptha et al., 2005). Preservations of calcareous particles and organic matter follow different CaCO₃ saturation and oxygen concentrations of subsurface to bottom water masses which follow the glacial/interglacial cycles (Ivanova et al., 2003; Guptha et al., 2005).

7. Objectives

In view of the above discussions the present study is aimed to address the following objectives

1. To reconstruct seasonal SST variations over last 40 ka by using Artificial Neural Net Work Technique on the planktonic foraminiferal faunal census data in order to gain an understanding on the fluctuations of monsoon upwelling.

2. To establish the relationship between selected planktonic foraminifer species shell weights, relative abundance and carbonate ion in the upwelling and non-upwelling regions of the Arabian Sea.

3. To reconstruct glacial/interglacial monsoon productivity and OMZ variability along the Eastern Arabian Sea.

4. To reconstruct the productivity variations in NE Arabian Sea over last 40 ka and to evaluate the link between productivity and winter cooling and NE monsoon strength in the NE Arabian Sea.