CHAPTER I
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Most elements of lower atomic masses occur in nature as mixtures of stable isotopes, for example, hydrogen ($^1\text{H}$, $^2\text{H}$), carbon ($^{12}\text{C}$, $^{13}\text{C}$) and oxygen ($^{16}\text{O}$, $^{17}\text{O}$, $^{18}\text{O}$). Harold Urey (1947) showed that significant differences in physico-chemical properties exist among these isotopes due to their finite mass differences resulting in their fractionation during chemical reactions. In nature, the fractionation or partitioning of these isotopes is caused mainly by equilibrium isotope exchange reactions or kinetic reaction processes (Hoefs 1980). Of these two processes, isotope exchange reactions can be considered as special cases of general chemical reactions. The present work deals with a particular exchange reaction system, $\text{CaCO}_3$-$\text{H}_2\text{O}$ system. In such a system, where $\text{CaCO}_3$ (calcium carbonate) is precipitating from $\text{H}_2\text{O}$ (water), the relevant exchange
reaction can be written as:

\[ \frac{1}{3} \text{Ca}^{16}O_3 + H_2^{18}O \xleftrightarrow[\text{K}]{\text{1/3 Ca}^{18}O_3 + H_2^{16}O} \]  

(1)

where \( \text{K} \) is the equilibrium constant. The partitioning of two oxygen isotopes in the two phases namely \( \text{CaCO}_3 \) and \( \text{H}_2\text{O} \) can be expressed by introducing the concept of fractionation factor which is defined as:

\[ \alpha = \frac{R_{\text{CaCO}_3}}{R_{\text{H}_2\text{O}}} \]

where \( R_{\text{CaCO}_3} \) and \( R_{\text{H}_2\text{O}} \) are \( \text{18O}/\text{16O} \) ratios in \( \text{CaCO}_3 \) and \( \text{H}_2\text{O} \) respectively. If the isotopes are randomly distributed over all possible positions in each of these two compounds, then

\[ \alpha = K^{1/n} \]

where \( n \) = number of atoms exchanged. For the exchange reaction given in (1), where monoatomic exchange occurs,

\[ \alpha = K = \frac{\left[\text{Ca}^{18}O_3\right]^{1/3} \left[H_2^{16}O\right]}{\left[\text{Ca}^{16}O_3\right]^{1/3} \left[H_2^{18}O\right]} \]

\[ = \frac{\left[\text{18O}/\text{16O}\right] \text{CaCO}_3}{\left[\text{18O}/\text{16O}\right] \text{H}_2\text{O}} \]

Since equilibrium constant (K) or \( \alpha \) depends on temperature (T), carbonates precipitated from water of a constant isotopic composition but at different temperatures, will have different \( \text{18O}/\text{16O} \) ratios. This idea is the basic
concept behind the attempt for a quantitative determination of palaeotemperatures of ocean water.

Isotopic ratios are usually reported in terms of $\delta$-values, defined as the relative deviation of the sample isotope ratio from that of an international standard and is expressed in per mil ($^{\circ}/oo$):

$$\delta = \left( \frac{R_s}{R_{std}} - 1 \right) \times 10^3^{\circ}/oo$$

where $R_s = ^{18}_0/^{16}_0$ in sample

$R_{std} = ^{18}_0/^{16}_0$ in standard

Empirical relation between $\delta^{18}O$ of precipitating CaCO$_3$ from a given water solution and solution temperature was first determined by McCrea (1950) and has the following form:

$$T = a + b \left( \delta_c - \delta_w \right) + c \left( \delta_c - \delta_w \right)^2$$

where $T =$ temperature in $^\circ$C, $\delta_c = \delta^{18}O$ of CO$_2$ obtained from carbonate by reaction with 100% orthophosphoric acid at $25^\circ$C, $\delta_w = \delta^{18}O$ of CO$_2$ (tank gas) equilibrated isotopically at $25^\circ$C with water from which the carbonate was precipitated. Both $\delta_c$ and $\delta_w$ are measured against the same mass spectrometer standard gas.

McCrea's calibration was made on inorganic CaCO$_3$. But it was not clear whether the same equation can be applied to fossil CaCO$_3$ for determining temperatures of ancient ocean water (Emiliani 1981). Subsequent investigators measured $\delta^{18}O$ in CaCO$_3$ of both biogenic (e.g. mollusks, foraminifera,
etc.) and abiogenic nature as a function of temperature. Table I.1 gives different values for the coefficients a, b, and c obtained by different investigators. It can be seen that most of the equations have similar slopes indicating a temperature coefficient, $\frac{\partial \delta^{18}O}{\partial T}$, of about $-0.2$ $^{\circ}/oo$ per $^{\circ}$C and hence any one of them can be used for calculating relative temperature change.

Emiliani (1955) first analysed $\delta^{18}O$ in planktonic foraminifera separated from different depths in a core from the Caribbean sea. He obtained a saw-tooth pattern in the value of $\delta^{18}O$ as a function of depth (or time). A straightforward interpretation of this signal in terms of sea surface temperature (SST) variation is not possible since the $\delta^{18}O$ of foraminiferal CaCO$_3$ not only depends on temperature (T) but also on isotopic composition of water in which the organisms grow. It is known that the $\delta^{18}O$ of sea water depends on global and local climatic effects. Extraction of a large amount of water from the oceans as ice on the continents during colder periods change the isotopic composition of the water in the oceans and constitute the global effect whereas evaporation, river discharge, etc. are the local components. The difference in $\delta^{18}O$ between the glacial and interglacial, the glacial-interglacial amplitude (GIA) was ~ $1.8$ $^{\circ}/oo$ in the Caribbean sea. Of this, Emiliani attributed about $0.4$ $^{\circ}/oo$ to water-$\delta^{18}O$ change, while the remainder i.e. $1.4$ $^{\circ}/oo$ of the signal was explained by glacial-interglacial SST change of about $6^{\circ}$C. However, Shackleton (1967) based on simultaneous
Table 1.1
Empirical Equations for CaCO$_3$-H$_2$O $\delta^{18}O$ Equilibrium

<table>
<thead>
<tr>
<th>Nature of CaCO$_3$</th>
<th>a</th>
<th>b</th>
<th>c</th>
<th>Investigators</th>
</tr>
</thead>
<tbody>
<tr>
<td>Inorganic</td>
<td>16.0</td>
<td>-5.1</td>
<td>0.09</td>
<td>McCrea (1950)</td>
</tr>
<tr>
<td>Mollusk</td>
<td>16.5</td>
<td>-4.3</td>
<td>0.14</td>
<td>Epstein et al. (1953)</td>
</tr>
<tr>
<td>Mollusk</td>
<td>16.9</td>
<td>-4.2</td>
<td>0.13</td>
<td>Craig (1965)</td>
</tr>
<tr>
<td>Inorganic</td>
<td>16.9</td>
<td>-4.68</td>
<td>0.10</td>
<td>O'Neil et al. (1969)</td>
</tr>
<tr>
<td>Mollusk</td>
<td>17.04</td>
<td>-4.34</td>
<td>0.16</td>
<td>Horibe and Oba (1972)</td>
</tr>
<tr>
<td>Core top benthics</td>
<td>16.9</td>
<td>-4.0</td>
<td>0.00</td>
<td>Shackleton (1974)</td>
</tr>
<tr>
<td>$&lt; 16.9^\circC$</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Extension of O'Neil et al. (1969) in $&gt; 16.9^\circC$ range</td>
<td>16.9</td>
<td>-4.38</td>
<td>0.10</td>
<td>Shackleton (1974)*</td>
</tr>
<tr>
<td>G. sacculifer</td>
<td>16.998</td>
<td>-4.52</td>
<td>0.028</td>
<td>Erez and Luz (1983)</td>
</tr>
</tbody>
</table>

These slightly differing relationships yield a range in temperature coefficients of $\delta^{18}O$ from -0.2 to -0.25 °/oo per degree centigrade.

* For the present work, the relationship of Shackleton (1974) i.e. $T = 16.9 - 4.38 (\delta_c - \delta_w) + 0.1 (\delta_c - \delta_w)^2$ has been used. This yields temperature coefficient for CaCO$_3$-$\delta^{18}O$ as -0.2 °/oo/°C.
measurements of planktonic and benthic foraminifera from the same core, reinterpreted Emiliani’s GIA of 1.8 °/oo to correspond mainly with the global ice volume effect. Several other studies have brought out further complications in the interpretation of $\delta^{18}O$ of foraminifera, e.g., different depth habitats and vertical migration of forams (Emiliani 1954), seasonal variation in abundances (Williams et. al., 1981; Ganssen and Sarnthein 1983), disequilibrium fractionation (Shackleton et. al., 1973; Vergnaud-Grazzini 1976; Fairbanks et. al., 1982), gametogenic calcification (Duplessy et. al., 1981a), dissolution (Erez 1979) and bioturbation (Mix 1987).

Till these controversies are resolved, it seems difficult to accurately delineate the glacial-interglacial temperature change and the global ice volume effect. A possible solution is to look for other sources of climatological data to find out the temperature and/or ice volume effect. One such source is the $\delta^{18}O$ composition of ancient ice sheets and their correlation with the past sea level change. But controversies still exist among these various estimates of ice volume effect (Mix 1987).

Apart from these dominant global climatic effects of ice volume and temperature, down core $\delta^{18}O$ is considerably modified by small scale, local oceanographic processes, some of which are described below.

I.1 Local Effects in the $\delta^{18}O$ of Foraminifera

In Gulf of Mexico, a negative $\delta^{18}O$ spike (upto ~
1.5 °/oo) during the transition from the Last Glacial Maximum (LGM) to Holocene, superimposed on normal GIA, was observed. This was explained by the presence of a less saline water that originated from the melting of the Laurentide ice sheet (Kennett and Shackleton 1975) and was considered to be a local phenomenon. Berger et al. (1977a) and Berger (1978), however, proposed that such a meltwater lid during last deglaciation could have existed globally. Subsequent studies indicated presence of freshwater spike from melting of Fennoscandian and Barents ice sheets (Grosswald 1980; Jones and Keigwin 1988). If such an effect contributes to the total δ¹⁸O signal, it becomes difficult to assess the global ice volume effect.

Inspite of these aforementioned problems, down core δ¹⁸O analysis on various species of foraminifera from different oceans have yielded more or less similar signals. A large number of studies from various ocean cores suggest that about ~ 1/3 of the GIA is due to temperature variation while majority (~ 2/3) of the signal is due to ice volume effect. This makes the δ¹⁸O stratigraphy as a proxy ice volume indicator. After about a decade of the initial work of Shackleton and Opdyke (1973, 1976), it is now firmly established that the entire pleistocene experienced more than 20 ice ages and not four (e.g. Wisconsin, Illinoian, Kansas, Nebraskan) as was thought by Quaternary geologists. The following section will address the cause of these ice ages.
I.2. $\delta^{18}O$ Cycle in the Ocean Cores: Causative Mechanism

Since oxygen isotopes in deep sea cores provide continuous palaeoclimatic record for a long period, much effort has been made to establish an absolute $\delta^{18}O$ chronology for the entire pleistocene. Initially the chronology was based on palaeomagnetic boundaries coupled with the assumption of constant sedimentation rates in between (Shackleton and Opdyke 1973). Subsequently Broecker and van Donk (1970) made the correlation between isotopic stages and coral terraces dated by U-series isotopes.

But Hays et al. (1976), for the first time, made a combined effort to produce a time scale tuned to orbital variations, which established the link between terrestrial climatic changes and astronomical forces, the so-called Milankovitch hypothesis. The currently available SPECMAP (Imbrie et al., 1984) time scale is the culmination of Hays et al's effort, which demonstrated that isotopic signatures are phase locked and are strongly coherent with orbital variations, in all the frequencies namely 19 and 23 kyr. (precessional), 41 kyr. (obliquity) and 100 kyr. (eccentricity). According to Imbrie et al. (1984), "85 percent of the observed isotopic variance is linearly related to orbital forcing". However, a detailed investigation of the last deglaciation i.e. 20 kyr. to the present, shows that the isotopic record is not strictly a linear function of orbital forcing.

Now we turn our attention to the significance of carbon
isotope changes in the foraminifera.

1.3 Carbon Isotope Records of Climatic Change

The carbon isotopic composition of calcium carbonate precipitated from sea water is controlled by (Faure, 1986):

1. $\delta^{13}C$ value of the dissolved $CO_2$ gas, in equilibrium with the carbonate and bicarbonate ions in solution.

2. Fractionation of carbon isotopes between $CO_2$ gas, the carbonate and bicarbonate ions in the solution and solid $CaCO_3$.

3. Temperature of isotopic equilibration.

4. pH and eH values of the medium

where $\delta^{13}C = \left( \frac{^{13}C/^{12}C}_{\text{sample}} - 1 \right) \times 10^3 \text{ o/oo}$

The $CO_2$ gas is nothing but atmospheric $CO_2$ dissolved in ocean water, which produces several components thereafter. The total $CO_2$ can be represented as:

$$\sum CO_2 = CO_3^{(aq.)} + CO_2^{(dissolved)} + HCO_3^{(aq.)}$$

Consequently, the exchange reactions among the various components can be written as:

$$^{13}C_3O_2^{(dissolved)} + H^{12}CO_3^{(aq.)} \leftrightarrow ^{12}CO_2^{(diss.)} + H^{13}CO_3^{(aq.)}$$
Among these steps, the bicarbonate-carbonate exchange reaction is the most important, because the HCO$_3^-$ is the major component of the $\Sigma$CO$_2$. Hence by approximating $\Sigma$CO$_2$ by HCO$_3^-$, the relation between the $\delta^{13}$C of CaCO$_3$ and HCO$_3^-$ can be written as follows (Emrich et al., 1970):

$$\delta^{13}\text{C calcite} = \delta^{13}\text{C(HCO}_3^-,\text{aq.}) + 1.85 + 0.035 \; (T-20)$$

where $\alpha^{13}\text{C}_\text{HCO}_3^-$ = 1.00185, T = temperature in degree centigrade. The temperature coefficient of calcite-$\delta^{13}\text{C}$, i.e. $\partial \delta^{13}\text{C}/\partial T$ ~ 0.035 °C/°C, is quite small. Hence any change in $\delta^{13}$C of foraminifera can be interpreted in terms of change in the $\Sigma$CO$_2$ reservoir alone.

In general, $\delta^{13}$C of $\Sigma$CO$_2$ shows a stratification profile with depth in open ocean similar to those of nutrients like PO$_4^3-$ and NO$_3^-$ (Kroopnick 1974). This profile is controlled by organic productivity in the ocean. In the euphotic zone, organic productivity is high. The preferential uptake of $^{12}$C by organic matter through metabolic processes makes them depleted in $^{13}$C and consequently enriches surface water in $\delta^{13}$C. When the organic matter sinks through the water column, it gets oxidised and releases $^{13}$C depleted CO$_2$.
in the deeper water. As a result, deep waters are depleted in $\delta^{13}C$ relative to surface. This $\delta^{13}C$ profile can be recorded in foraminiferal calcite-$\delta^{13}C$ and can be used as palaeo-oceanographic tracer, provided different species grow at different depths in isotopic equilibrium with the ambient. Departure from this equilibrium and its effect on $\delta^{18}O$ has been mentioned earlier. Such disequilibrium may affect the $\delta^{13}C$ values too. Disequilibrium precipitation may occur, if apart from $\text{HCO}_3^-$ ($\Sigma\text{CO}_2$), some amount of respiratory $\text{CO}_2$ or $\text{CO}_2$ derived from oxidation of organic matter, takes part in the $\text{CaCO}_3$ shell building process. Hence it has to be assessed how reliable foraminiferal $\delta^{13}C$ are as palaeoclimatic indicator.

Evaluation of Foraminiferal $\delta^{13}C$

Majority of the work on $\delta^{13}C$ has been done on benthic foraminifera. Most of the benthics precipitate calcite out of isotopic equilibrium. The disequilibrium at a particular place is not constant but depends on taxonomic groups (Grossman 1984). For a particular species, however, this disequilibrium probably remains constant with time and can be approximately corrected; for example:

$$\delta^{13}C_{\text{Cibicides}} = 0.954 \delta^{13}C_{\text{T}CO_2} + 0.095$$
$$\delta^{13}C_{\text{Uvigerina}} = 0.904 \delta^{13}C_{\text{T}CO_2} - 0.805$$

where $\text{T}CO_2$ is total dissolved $\text{CO}_2$ ($\Sigma\text{CO}_2$) (Duplessy et al.,
In general, $\delta^{13}C$ of these commonly used species like Cibicides, Uvigerina, Planulina, etc. record $\Sigma CO_2$ of bottom water quite reliably (Grossman 1984). Planktonic foraminifera in most cases, show a higher scatter in downcore $\delta^{13}C$ values than benthics (Williams et al., 1977; Broecker and Peng 1982). However many species from the open ocean (e.g., N. dutertrei, G. bulloides etc.) have yielded very similar $\delta^{13}C$ variation, indicating that they too record surface water $\Sigma CO_2$ property (Shackleton and Pisias 1985).

Summarising, it is clear that stable isotope palaeoclimatology has experienced multifarious developments in the last three decades. Our present understanding of the global isotope palaeoclimatology reveals the following important features:

1. There exists a cyclic $\delta^{18}O$ stratigraphy in the ocean sediments for at least the last 2 million years.

2. Such fluctuations in $\delta^{18}O$ are mainly controlled by the waxing and waning of continental ice sheets, which in turn is controlled by the sun-earth orbital geometry changes namely changes in eccentricity, obliquity and precession of the earth.

3. The average glacial-interglacial change in $\delta^{18}O$ (GIA) is between 1.1 °/oo and 1.6 °/oo.

4. $\delta^{18}O$ signal and sea-level estimates show a fairly good agreement.
5. $\delta^{13}C$ difference between planktonics and benthics was maximum during LGM and is perhaps controlled by the change in the oceanic nutrient cycles, coupled with the deep sea circulation change.

Superimposed on these global effects, many local climatic changes modify the isotope patterns. By appropriately subtracting the global signals, one can decipher the local climatic variation. For example, in the Mediterranean and the Red sea GIA is $\sim 4^\circ/oo$ against the global GIA of $\sim 1.6^\circ/oo$ (Vergnaud-Grazzini, 1975; Thunell et al., 1988) and is explained by increased local evaporation in these marginal seas during LGM. Variations in coastal upwelling systems also show very complex local phenomena (Prell 1984a, b; Suess and Thiede, Thiede and Suess 1983). Such phenomena if present on a large scale ought to have influenced global climate as well.

We now discuss the problems and prospects of palaeoclimatic studies in the northern Indian ocean and the objectives of the present study in this context.

1.4. Review of Earlier Work in the Northern Indian Ocean

One of the early systematic studies of stable isotopes in foraminifera from the northern Indian Ocean was carried out by Oba (1969). His study focussed on estimating the palaeotemperature of the ocean water in relation to faunal geography and identifying the depth habitats of different foram species. $\delta^{18}O$ stratigraphy was also tested against the
global climatic cycles. Subsequent to the work of CLIMAP group (1976) on world oceans, which included the northern Indian Ocean also, increasing attention was drawn towards this region. This is because, the Arabian sea, the Bay of Bengal and the adjacent oceans respond to a unique feature in the regional climate namely, "monsoon", which introduces important physico-chemical changes in these oceans. Characteristic long term changes of the Indian monsoon, their causes and effects have been the subject of numerous studies (Miller and Keshavamurty 1968; Lighthill and Pearce 1981; Fein and Stephens 1987). Broadly, the monsoon system has two dominant effects in the oceanic realm:

1. Heavy rainfall during June to September over the Indian subcontinent. This results in a large river water discharge to the Bay of Bengal and the Arabian sea.

2. Intense upwelling due to the prevailing wind system takes place in the coastal Arabia and the west coast of India.

Both these phenomena bring about significant changes in the sea water salinity and sea surface temperature, which, in principle, can be traced spatially and temporally by the stable isotope ratios of oxygen and carbon (by the productivity changes).

A pioneering work in this regard was done by Duplessy
(1982) who showed that apart from the general global climatic cycles, the $\delta^{18}O$ stratigraphy of foraminifera in the Bay of Bengal and the Arabian sea records important information about local oceanographic changes. His studies, based on Holocene core tops and LGM levels of several cores in the Arabian sea and Bay of Bengal, showed significant changes in water properties in the past.

At present the Arabian sea is dominated by strong evaporation. On the contrary the Bay of Bengal experiences large river discharges from the Ganges, Brahmaputra, Irrawady rivers. Consequently the mean annual surface salinity in the Arabian sea increases from south to north by 1.5 °/oo and decreases in the Bay of Bengal by 3 °/oo. Modern core top G. ruber species record this salinity gradient in their $\delta^{18}O$ composition; this shows a south to north increase in $\delta^{18}O$ by 0.6 °/oo in Arabian sea and decrease by 1.1 °/oo in the Bay of Bengal. Duplessy (1982) showed that this gradient in $\delta^{18}O$ during LGM time was higher in the Arabian sea (~1 °/oo) and considerably smaller in Bay of Bengal (~0.2 °/oo). This has been explained by enhanced evaporation and low river discharge during LGM, in these two basins respectively. Zonal nature of the iso-$\delta^{18}O$ contours and their similarities along particular latitude, indicated decreased runoff from west-bound peninsular rivers as well as weak upwelling conditions in the western Arabian sea (op. cit.). These conclusions supported the earlier observations of Prell et al (1980) in this region. Based on planktonic foraminiferal biogeography, Prell et al. showed
that the Arabian sea was characterised by warmer SST and weak-upwelling condition during LGM while increased salinity was prevailing in the Bay of Bengal. Thus glacial-interglacial changes in $\delta^{18}O$ (a function of only ice volume and temperature) are considerably modified by local climatic effects in the northern Indian ocean.

Since upwelling off the Arabian coast is coupled with monsoon, its variation through time has been the subject of many investigations. During summer months, upwelling brings up cold, nutrient-rich water from the deeper levels and creates a distinct temperature anomaly. The centre of this upwelling lies between 17°N and 21°N and is 4°C cooler than the mid-ocean SST (Prell and Streeter 1982). If foraminifera, growing in this region, record such a temperature change, a total of ~0.8 °/oo change in the $\delta^{18}O$ values should be observed in their shells across this upwelling zone (considering the CaCO$_3$ temperature coefficient ~0.2 °/oo/°C). The study by Prell and Curry (1981) on Holocene core top planktonic foraminifera collected across this upwelling zone, indeed showed such a range in the $\delta^{18}O$. $\delta^{18}O$ compositions of any surface dwelling foraminifera such as G. sacculifer, G. ruber, G. bulloides, G. menardii show a total 0.8 °/oo change across the upwelling zone same as the predicted value. This prompted Prell and Curry to suggest that synoptic mapping of these shallow dwelling species across the upwelling zone can be a potential tool for studying its variation from LGM to the present. Due to the peculiar topographic configuration (Fig. 2.1), however,
obtaining undisturbed cores is quite difficult especially near the coast of Arabia, where the upwelling is maximum.

During the same investigation, Prell and Curry (1981) noticed that the percentage (%) of modern *G. bulloides* is directly correlated with PO₄ content of the surficial water (an index of upwelling) and inversely correlated with SST. Therefore, it was concluded that the abundance of *G. bulloides* can be used as an index of the intensity of upwelling. Subsequent investigation by Prell (1984a, b) on the down-core variation of *G. bulloides* from this region, showed that its abundance decreased during 18 kyr B.P. indicating a weak upwelling, while it increased around 9 kyr B.P., suggesting a stronger upwelling than today. This is consistent with the earlier observation of Duplessy (1982) which showed a weak SW-monsoon during LGM and a probable less intense upwelling off Arabia.

In addition, Duplessy proposed that a stronger NE monsoon existed during the same period but no oxygen isotope evidence was found for it. If true, such a stronger NE monsoon should presumably increase the evaporation in the northern Arabian sea due to the presence of dry NE winds over this region (op. cit.) during LGM.

Apart from the oxygen isotope studies, evidence for a weak monsoon during LGM and extreme aridity over south Asia comes from a number of other sources. Pollen studies in conjunction with δ¹⁸O stratigraphy from the western Arabian sea have shown the prevalence of an extreme arid condition around 20 kyr B.P. and a wet condition around 11 kyr B.P.
Increased glacial aridity is also indicated by high quartz content in Arabian sea (Kolla and Biscaye 1977), increase in sand dune activity (Sarnthein 1978) and stronger loess storms in the cold deserts of inner Asia (Huang 1973). Dessication took place in the large lakes from Rajasthan around 20 kyr. B.P. and recovered only at 13.6 kyr. B.P. (Singh et al., 1974; Wasson et al., 1983).

Study of the fluvial deposits from the Deccan upland region shows river aggradation and drier climate around LGM whereas early Holocene was characterised by strong incisional stages of these rivers, when discharge was high (Kale and Rajaguru, 1987). Based on the abundance of planktonic foraminifera Globoquadrina dutertrei, as well as total planktonic assemblages, increased salinity in the Bay of Bengal during LGM has been proposed (Cullen 1981). This study also indicates a strong salinity reduction in the early Holocene. Such salinity changes have been explained by varying river run-off coupled with monsoon fluctuations for last 20 kyr.

It is of interest to know the chronological evolution of monsoon from a glacial weak condition to its Holocene strength. Pollen study by Van Campo (1986) suggests that the evolution of the monsoon was gradual. After LGM, rainfall started increasing along the west coast as early as 16 kyr. at 15°N latitude. This produced extensive development of mangrove forests which culminated at 11 kyr. B.P. Monsoon reached its maximum strength around 10 kyr. B.P. at 10°N latitude and 8 kyr. B.P. at 28°N. Such an evolution of
monsoon was explained by progressive northward movement of intertropical convergence zone (ITCZ) (op. cit). In the western Arabian sea (north of 10°N) G. bulloides reached its maximum abundance around 9 kyr. B.P. indicating the strongest upwelling which is consistent with the above proposition.

Unlike oxygen, carbon isotope studies in carbonates from these sediments are meagre. Prell and Curry (1981) attempted to study the effect of upwelling on $\delta^{13}$C of modern planktonic foraminifera, based on the fact that the deeper water brought up by upwelling contains $\Sigma CO_2$ depleted in $\delta^{13}$C arising from oxidation of organic matter. Consequently one would expect a correlation between $\delta^{13}$C and various indices of upwelling. But $\delta^{13}$C values showed very poor correlation with both PO$_4$ and SST (two strong upwelling indices). Prell and Curry explained the lack of such correlations by one or more of the following reasons:

1. Disequilibrium fractionation of carbon isotopes.
2. Dilution of upwelled water by the surrounding reservoir.
3. Quick equilibration between atmospheric CO$_2$ and dissolved $\Sigma CO_2$.

However, $\delta^{13}$C of benthic foraminifera has been successfully used to reconstruct the hydrographic condition in the northern Indian ocean during LGM (Kallel et al 1988). Analysis of benthic Cibicides reveals significant
differences between the characteristics of intermediate and deep water masses during LGM. In the modern situation, the vertical decrease in $S^{13}C$ from 2000m to 2300m is less than 0.1 °/oo. During LGM, a sharp discontinuity was found around this depth separating intermediate and the deep water masses. The deep water was depleted in $S^{13}C$ by as much as 1 °/oo against the intermediate water value of ~ 0 °/oo. Similar discontinuity was observed in the $S^{18}O$ value also. GIA in the cores raised from the intermediate water depth, are ~ 1.1 °/oo whereas, in the deep waters GIA has values upto 1.5 °/oo (LGM levels being heavier). This was explained by the presence of a ventilated intermediate water mass (with same temperature as of today) and a cooler ($S^{18}O$ enriched), poorly ventilated (CO$_2$ rich and hence $S^{13}C$ depleted) deep water during LGM. Today's intermediate water originates from the Red Sea and the Persian Gulf region. During LGM, sea-level in this region was lower thus, significantly reducing their contribution to intermediate water. Consequently three probable sources for this ventilated intermediate water have been suggested:

1. Dense water formed by increased evaporation and reduced SST in the northern Arabian sea during LGM.

2. Expansion of Antartic intermediate water to the north of 10°S.

3. Increased flow of intermediate water from the Pacific Ocean through the Indonesian Archipelago.
Change in the $^{13}\text{C}$ of the deep water has been explained by a strong stratification and development of a deep thermocline during the same period (Kallel et al., 1988).

A recent work on palaeomonsoon circulation which is of great significance has been reported by Fontugne and Duplessy (1986). They used the $^{13}\text{C}$ of organic matter as a tracer. $^{13}\text{C}$ in sedimentary organic matter is a mixture of contributions from marine and terrestrial sources. Since the average $^{13}\text{C}$ values for these sources are $-26$ and $-20$‰ (w.r.t. PDB) respectively, one can calculate the fraction of terrestrial and marine organic carbon in these sediments, provided the organic matter is not diagenetically altered. Their reconstruction showed a higher terrigenous carbon input to the Bay of Bengal along a NE-SW axis, indicating an increased NE monsoon activity during LGM. This was supported by the total organic carbon (TOC) content of the sediment, which increased in the eastern Bay of Bengal. In the western Arabian sea, TOC decreased during LGM, indicating a weak upwelling. However, the increased NE monsoon activity was not indicated by any of the earlier $^{18}\text{O}$ studies cited above. An enhanced input of terrestrial carbon, deep into the Bay of Bengal during LGM, is also supported by the measurement of sediment accumulation rate. In the Bay of Bengal this rate was higher during LGM (Foucalt and Fang 1987; J.C. Duplessy and R. Chesselet, preprint) while that in the Andaman sea and the Arabian sea was much lower (Van Campo 1986; Fontugne).
and Duplessy 1986). A higher terrigenous flux during the glacial period is perhaps due to the direct pathways of rivers to the ocean cutting through the newly exposed shelf. Though terrigenous input controls the sedimentation rate in near coastal regions, in the deep ocean other factors such as productivity, eolian transport and dissolution play an important role.

The above discussion on the available isotopic and palaeontological data shows that significant changes in past climatic conditions have taken place in the northern Indian ocean and the adjacent regions.

In addition to these experimental studies, a significant amount of work has also been done to simulate the past climatic conditions based on theoretical modelling. These models take into account different boundary conditions provided by the geological data. The major boundary conditions, which control the climate over the northern Indian ocean and adjoining continents are SST, size and elevation of the Asian continent, surface albedo and the seasonal distribution of solar radiation over the Asian continent (Prell 1984b). A change in any of these parameters is potentially capable of perturbing the local climate. By using the GFDL general circulation model (GCM), Hahn and Manabe (1975) showed that the high elevation of the Tibetan plateau produces in part, an intensified monsoon. Absence of the mountains, in the simulation, produced a much weaker monsoon. However, the elevation of the Tibetan plateau is almost constant when one is concerned about the climatic
Another GCM experiment by Manabe and Hahn (1977) demonstrated the effect of surface albedo and produced a weak monsoonal circulation during LGM due to the higher albedo over Asia. Presence of an extensive snow cover over the Tibetan plateau perhaps was responsible for such a higher albedo during the LGM, which reduced the land-ocean temperature contrast. CLIMAP's (1976) simulation with the modern terrestrial boundary conditions and the glacial SST produced a monsoon intensity similar to that of today. This indicates that the SST change is perhaps not very significant for the monsoon performance. Recent studies by Kutzbach and Guetter (1986) indicate that the response of the monsoon circulation and tropical precipitation to change in the solar radiation is much larger than the response to the changes of glacial-age boundary conditions. Their model predicts a higher aridity during LGM and 10-20% increase in the summer monsoon precipitation between 12-16 kyr. ago.

Relation between the terrestrial climatic change and the astronomical forcing (Milankovitch theory) has been mentioned earlier. Spectra of time series of $^{18}O$ and G. bulloides (index of upwelling) from the Arabian sea show distinct peaks around 12 kyr. and 23 kyr. respectively (Prell 1984a). Both these signals are coherent with the insolation curve over the orbital frequencies. This indicates that monsoonal upwelling is linearly correlated with the solar radiation over time scales of the precessional frequency (23 kyr). However $^{18}O$ (ice volume)
and $G.\text{ bulloides}$ (upwelling) lag insolation change by ~5.5 kyr. Such a delay of monsoonal circulation is probably due to the persistence of seasonal snow cover into the summer and its wide areal coverage over the Tibetan plateau which prevented low-level heating and the consequent pressure gradient (op. cit.). A latest review by Prell and Kutzbach (1987) using the GCM simulation based on diverse arrays of data (also see COHMAP members 1988) on the monsoon variability shows four distinct monsoon maxima during interglacials for the last 150 kyr. Such maxima are caused by changes in the northern hemispheric summer radiation controlled by the precession of the earth. In this simulation, the glacial age boundary conditions produced an increased precipitation on the western Indian ocean along with glacial aridity and a weak SW-monsoon over Asia. However, glacial age boundary conditions have much less effect on the monsoon intensity compared to the insolation changes.

Summarising all the geological evidences (ocean sediments, pollen, lake level data, etc.) and model simulations, several important conclusions can be drawn about the past climatic changes in this region:

1. Presence of a weak SW-monsoon and weak coastal upwelling and indications of a strong NE monsoon during LGM.

2. Extreme aridity over southern Asia during LGM.
3. Presence of an $\text{O}_2$-poor, $\text{CO}_2$ rich deep water and more ventilated intermediate water front in the northern Indian ocean.

4. Increased precipitation and SW-monsoon circulation around 12-6 kyr. B.P.

I.5. Plan of the Present Study

In the foregoing section it has been shown that important information on palaeoclimatic changes have been obtained by using stable isotope tracers and foraminifera from the northern Indian ocean. However, several questions remain. The actual geological data base in this ocean on which the abovementioned conclusions are based is sparse. There is no real evidence from the foraminiferal $\delta^{18}$O data that the NE monsoon was stronger during LGM. Since most of the evidences lack absolute chronologies, it has not yet been possible to precisely date the glacial-interglacial transitions in the northern Indian ocean. In many cases oxygen-isotope stratigraphy itself is used to date the sediments. However, concurrence between the $\delta^{18}$O based chronology and that obtained by the absolute dating techniques may not be always achieved (Prell et al 1980; Keigwin et al 1984). Even among the various absolute dating techniques serious discrepancy may arise (Ku et al., 1968; Goldberg 1968). Such discordances arise because different isotopes used to date the sediments, have different sources.
and geochemical pathways. In addition their distributions are likely to be affected by various post depositional oceanographic processes viz. bioturbation, winnowing of bottom sediments, variation in redox condition etc. For example, earlier attempts of dating sediments of the Arabian sea by U-Th series isotopes did not prove very successful (Sarin et al., 1979; Sarin, M.M., personal communication, 1988, unpublished data; Borole 1980), while in the other oceans it proved to be a good technique (Goldberg and Koide 1962). Hence an intercomparison of these various methods can be made to ascertain their validity.

In the GCM simulation of Prell and Kutzbach (1987), influence of $\text{CO}_2$ change has not been taken into account and hence the question is, what role does the change in carbon cycle play in this ocean? Finally, the periodicities and the interrelationship of different climatic indices and their phase relation to the orbitally induced solar insolation should be tested. One such test has already been done for $\%$ of foraminifera, an upwelling index (Prell 1984a, b) and should be extended to other parameters e.g. $\delta^{18}O$, CaCO$_3$, etc.

The present work is an attempt to answer some of the queries cited above and has been focussed to:

1. study the geochronology of the sediments and put an absolute time bracket on the glacial-interglacial transitions by using $^{14}$C (upto 30 kyr.) and $^{230}$Th (beyond the range of $^{14}$C and upto ~ 300 kyr) dating
2. find out the nature and extent of local climatic variations in the northern Indian ocean vis-a-vis global changes, based on the long term $\delta^{18}O$-stratigraphy.

3. find an oxygen isotope evidence of a stronger NE-monsoon during LGM, based on high resolution $\delta^{18}O$ stratigraphy.

4. find out $\delta^{13}C$ variation from LGM to present, vis-a-vis $\delta^{18}O$ stratigraphy, in relation to changes in upwelling and organic productivity.

5. test the periodicities, coherence and phase relationship between $\delta^{18}O$, an ice volume index and CaCO$_3$, a productivity index.