Chapter 7

PALAEOCLIMATES

7. PALAEOCLIMATES

The stratigraphic record in Mainland Gujarat contains innumerable clues that help assess the nature of past climates in Gujarat. In the following sections these evidences are discussed in detail and an attempt is made to provide a coherent picture of climate change over the late Quaternary, aided by an intensive review of published data. It is probably obvious through the preceding chapter that different hydrological regimes prevailed during the three aggradation phases, which in turn are related to evolving climate regimes. To enable a one-to-one relation these climate regimes are termed here as Climate Phase 'n' (n=1,2,3) and are genetically related to Aggradation Phase 'n' (n=1,2,3). It is implicit that while discussing past climates and the control they exert on depositional systems, one is actually discussing the overtly dominant role of an allocyclic (syn. allogenic) parameter. Moreover the various forcing parameters that have determined the varying intensity of the southwest Indian monsoon are discussed.

The southwest Indian monsoon is propelled by temperature contrasts between land and ocean masses (Fig. 7.1). Such differences come about due to the greater heat capacity of the ocean relative to land. During the summer season, high solar radiation induces heating over the Tibetan plateau and northern India. This generates a pressure gradient; the warmer Asian continent becomes an area of low pressure while the Indian ocean represents a high pressure region (Madagascar high). This pressure gradient leads to low-level southwesterly summer monsson winds, i.e. the southwest Indian monsoon (Prell et al., 1992; Overpeck et al., 1996). In the winter months an oppositely directed wind current is generated whose genesis again lies in the differential heat capacities of land and ocean masses. These winds flow from northeastern India over much of Gujarat and

Rajasthan towards the Indian Ocean. The NE monsoon winds however do not yield rain over the state of Gujarat.

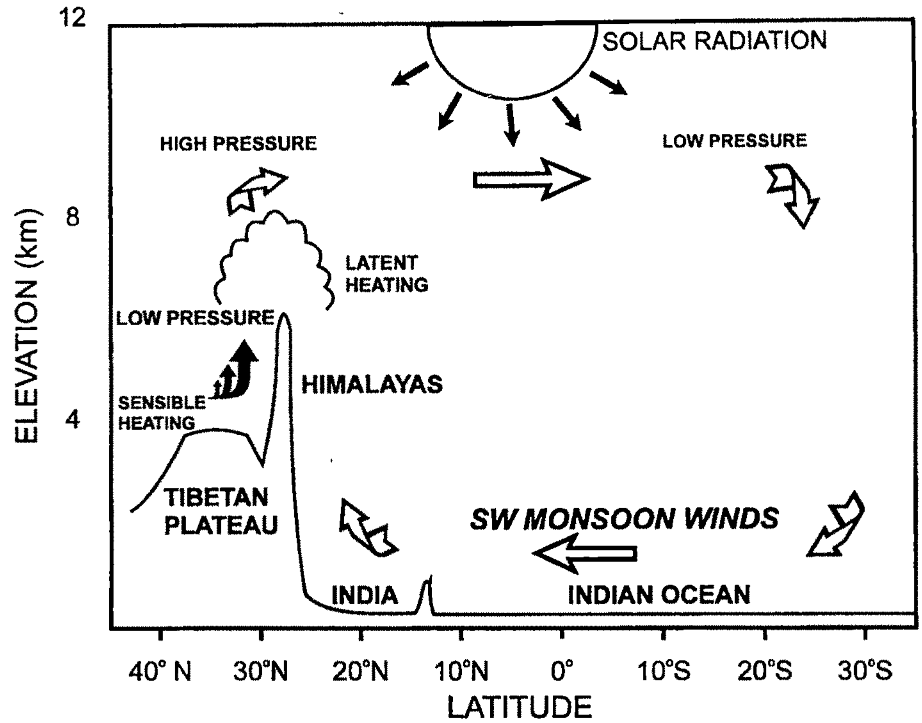

Figure 7.1: Structure of the SW Indian Monsoon (Prell et al., 1992)

In the recent past statistical analyses of rainfall time series have revealed certain trends. According to Pant & Hingane, (1988), regions peripheral to the Thar Desert show a net increase in regional rainfall. Singh et al., (1992) studied temporal changes in spatial variability of arid regions in northern India. Their studies again suggest a long term decrease in arid area with inter-annual variations of 34.6%. Figure 7.2 shows time series plots for the period 1871-1984, of mean annual rainfall in western Peninsular India (Sontakke & Singh, 1996), and area in India experiencing annual rainfall under 800 mm Singh et al., 1992).

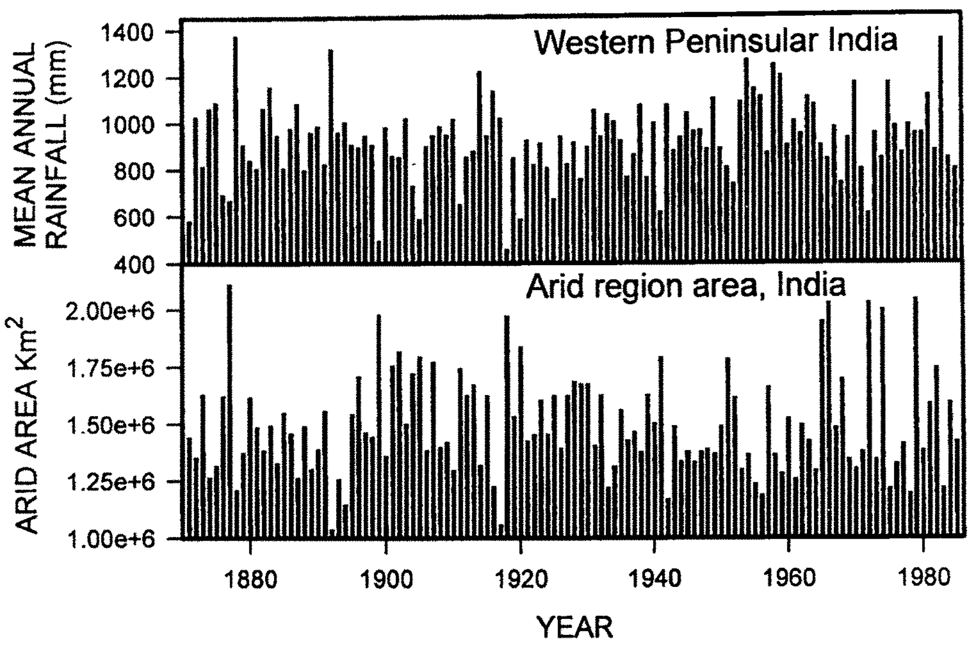

Figure 7.2 : Time series plot for the year 1871-1984 of mean annual rainfall in western Peninsular India and arid area in India.

It is interesting to note that long term perturbations in mean annual rainfall are coeval throughout India. Years during which India was largely covered by arid areas correlate with periods of extremely low mean annual rainfall in western Peninsular India. This implies that palaeoclimatic information derived from the Quaternary record of Gujarat may in actuality reflect regional climatic shifts over the Indian Peninsula.

7.1 Climate Phase 1

This climate phase covers the conglomerate-vertisol facies association of Aggradation Phase 1. The presence of vertisols suggests that a seasonal climate prevailed during this period. Periods of higher rainfall (monsoon months) were succeeded by dry climates which led to the formation of calcretes within the soils. Today the soil order 'vertisol' occurs typically in the sub-humid and humid tropics with calcic varieties dominating the former regions. The presence of conglomerates with structures indicating gravel dune migration again points to high discharges related to a wetter climate. Stable isotope analyses carried out on the calcretes indicates the presence of a woodland vegetation community (C3 plants). Since this photosynthetic pathway is effective in areas of high atmospheric pCO_2 it may be inferred that atmospheric carbon dioxide levels were high. A wetter climate is also indicated by the existence of a C3 vegetation biomass.

The time frame for this phase is not well established due to lack of absolute dates. The presence of Lower Palaeolithic tools within the deposits suggests an archaeological age between 250-100 ka BP. However the lower age limit largely remains conjectural. Thermoluminescence dates on the lowermost sediments at Vijapur give a minimum age of >300 ka BP. Hence it is only by circumstantial evidence that an estimate of the time frame can be arrived at. By comparison with the global climate records (references in following paragraphs) it is seen that a significantly wetter climate phase existed between 125-90 ka BP. Since in the stratigraphic record of Mainland Gujarat, vertisols are not seen at any other level it is conjectured that the vertisol-calcrete association represents a manifestation of an intense southwest Indian monsoon during the above time period.

Climate records from the Guliya ice cap on the Qinghai-Tibetan plateau, China provide a high resolution database to compare the present results. This record is based on a 308.6 m long ice core (Thompson et al., 1997). It shows low $\delta^{18}O$ values between -15 ‰ and -12 ‰ till 90 ka BP with a brief span of 5 ka during which values were around -18 ‰. The higher values (i.e. enrichment of the heavier isotope) represent interstadials while lower values mark the stadial periods. These values are higher than today by 2-3 ‰ suggesting that the monsoon was more intense. A similar result was obtained from the salt lake in northeastern Tibet wherein high carbon contents were observed between 123-90 ka BP (Yan & Petit-Maire, 1994).

Evidences of a more intense monsoon also comes from the Arabian Sea ocean core data. Ratios of di- and tri- unsaturated ketones termed alkenones are temperature controlled. The measurement of these prymnesiophyte-produced compounds provides a means of calculating ancient sea-surface temperatures. In this method the ratios are first calibrated under controlled conditions and the values extrapolated to those measured in the core. This approach was followed by Emeis et al., (1995), Rostek et al., (1993) and Bard et al., (1997). This approach suggests that during the Climate Phase 1 sea-surface temperatures off the Arabian coast ranged between 26 °C and 27 °C. Sea-surface temperatures measured from the $\delta^{18}O$ record of the planktonic foraminifer *P. obliquiloculata* differ from the alkenone approach by 1-2 °C. However the general trend of both curves matches (Emeis et al., 1995).

Past atmospheric CO_2 levels were determined in the Vostok ice core by Barnola et al., (1987) which shows a atmospheric carbon dioxide concentration ranging between 296 ppmv and 226 ppmv during the period 135 ka BP and 90 ka BP (Fig. 7.3). A marked fall occurs at 111 ka BP and is associated with a concomitant decrease in the Vostok isotopic temperatures by ~2 °C (Barnola et al., 1987). Such high atmospheric carbon dioxide levels led to the development of a vegetation ecosystem dominated by C3 plants and is in agreement with the isotopic $\delta^{13}C$ values of around -7 ‰ in calcretes. It is well known that the C3 photosynthetic pathway is more efficient under high CO_2 concentrations. Moreover experiments conducted by Polley et al., (1993) to monitor the response of vegetation biomass to changing levels of atmospheric CO_2 indicate that with increasing concentrations the efficiency of C3 plants towards utilising water increases considerably. This consequently leads to increased productivity of the C3 biomass over C4 plants.

Insolation which is the measure of incoming solar radiation was high at around 125 ka BP to 100 ka BP, about 60 Wm^{-2} higher than the present day values (Berger, 1978).

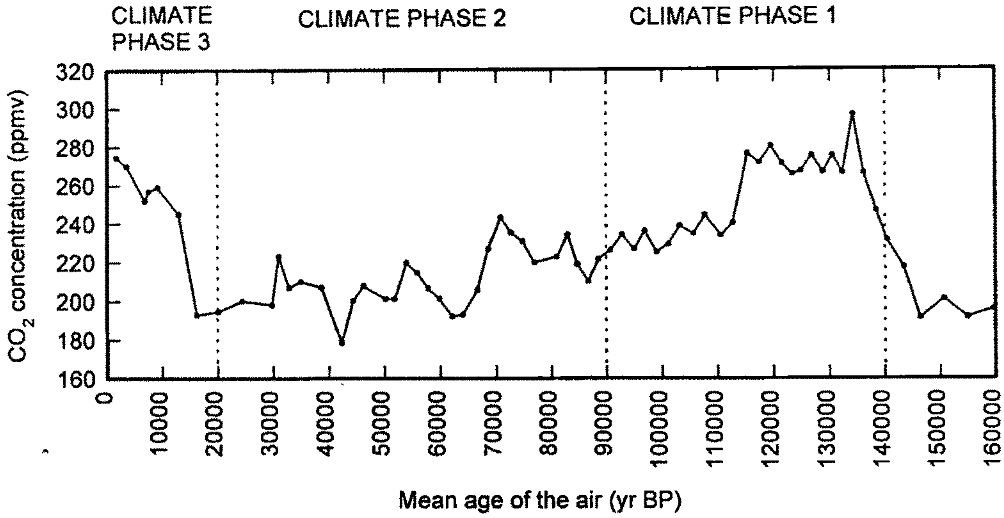

Figure 7.3: Carbon dioxide concentration in the atmosphere over the past 160 ka BP (Data from Barnola et al., 1987).

Using travertine deposits from the Western Desert in Egypt, Crombie et al., (1997) reported periods of extensive recharge between 100-200 ka BP which was related to increased monsoon strength. They compared the eustatic sea level curve with periods travertine formation. High sea levels at 100-125 ka BP and 200 ka BP, were also pluvial periods. This links continental palaeoclimatic records with interglacial signatures of the oceanic record. In the eastern Sahara during isotope stage 6-5e (140-130 ka BP) pluvial conditions are observed in the stable isotope record of lakes (McKenzie, 1993). Values as negative as -9.5 ‰ indicate presence of isotopically light water during this period. This is explained by the author in terms of intensification of the southwest Indian monsoon activity, during which moisture laden winds from the Atlantic produced isotopically light water through the continental effect (Rayleigh distillation). In the desert interiors of the Namib, detailed geochronology has shown a period of dune building activity between 115-95 ka BP (Stokes et al., 1997). The dunes accreted in response to changes in the southeastern Atlantic ocean sea surface temperatures which govern the northeast-southwest summer rainfall gradient.

7.2 Climate Phase 2

During this period ephemeral rivers covered the landscape which were characterised by flashy discharges. These rivers experienced infrequent rainfall and were dry during the major part of the year. The climate was semi-arid throughout this region. Vegetation densities were low enough to preclude deep soil development. Published thermoluminescence dates on this phase give an age grouping between 80 ka BP and 58 ka BP (Tandon et al., 1997). The latter date however is at a level much below the contact between Aggradation Phase 2 and 3. Additionally quartz and feldspar grains were dated using thermoluminescence (TL) and Optically stimulated luminescence (OSL) techniques. This material gives the age when the clock was reset to zero by sun bleaching and subsequent burial and effectively gives the age of sedimentation. Hence it does not reflect on the age of the red soil. The same holds true for the TL age of 48 ka BP measured by Prasad (1997) from Nal Sarovar, Gujarat which falls south of the study area. Radiocarbon dates on pedogenic calcrete nodules developed in the soil fall between 26 ka BP and 23 ka BP (Allchin et al., 1978). Petrographic examination of these nodules has shown them to be unaffected by repeated shrinkage processes that produce channel and planar voids through which fresh carbon may be introduced. Instead the bulk of the nodule is made of pristine low magnesium micritic calcite. This validates the credibility of these radiocarbon dates. As discussed in Chapter 3 no typological evolution is observe in the Palaeolithic Stone Industry, hence it is not possible to provide any archaeological age for this phase.

The dominance of kaolinite in the red soil along with montmorillonite indicates high mean annual rainfalls. The process of rubification itself necessitates a seasonal rainfall pattern with months of intense rainfall. Thus the red soil event represents a climatic amelioration event between 50 ka BP (as the TL date is below the upper boundary of the red soil, some time must be accommodated while taking into consideration the deposition of another 30-50 cm of sediment) and 23 ka BP. Capping the red soil again is 2-3 m succession of ephemeral river deposits. This succession is speculated to represent 3-5 ka of time and the boundary between Climate Phase 2 and Climate Phase 3 is kept in tandem with global climatic deterioration at the Last Glacial Stage (20-18 ka BP).

Stable isotopic values for the red soil calcrete nodules show a wide range in their $\delta^{13}C$ content between -7 ‰ to -1 ‰. Hence not much climatic interpretation is advisable. The vegetation could have been wooded grassland during this interlude, with wooded pockets giving the negative values and open grassland patches governing the soil carbon dioxide isotopic composition in other regions.

During this period sea surface temperatures were between 25.5 to 24 °C. A marked fall of around 21.5 to 23 °C occurs between 60-70 ka BP (Emeis et al., 1995). The general trend in cores taken from spots at either hemispheres in the Indian Ocean is a progressive fall in sea surface temperatures by about 1 °C (Bard et al., 1997). Carbon dioxide levels (Barnola et al., 1987) fell from 240 ppmv at the start of this phase to near 180 ppmv (Fig. 7.3). Insolation also dropped by approximately 40 Wm^{-2} (Berger, 1978). Oxygen isotope record of the Tibetan plateau (Thompson et al., 1997) does not exhibit any trends. Long stadial periods 20 kyrs in span represented by $\delta^{18}O$ values between -18 and -21 ‰ are punctuated by rapid less prolonged spurts of interstadials roughly 5 kyrs long. A significant signal is the presence of a long interstadial between 50 ka BP and 24 ka BP after which a clear stadial returns. This is in agreement with the Gujarat stratigraphic record. A similar red bed forming event between 40-20 ka BP has been observed on the African continent in the Maghreb desert also (Rognon, 1987). Dhir et al., (1994) also suggest that climates in the Thar Desert interior were much wetter with a reduction in aeolian dynamism between 25 and 50 ka BP. During this period hindered sand mobility helped in the secondary redistribution of carbonate and gypsum and also the formation of colluvial sediments. In the Namib desert (Africa) also wetter conditions prevailed between 40 and 23 ka BP (Vogel, 1989). Dune building episodes in the Namib were restricted between 46-41 ka BP and 26-20 ka BP (Stokes et al., 1997). High water tables and increased spring discharges led to pan formation between 40-20 ka BP in the southwestern Kalahari desert (Lancaster, 1989).

Carbon isotopic studies on tropical peats from the montane regions of the Nilgiris, southern India shows increase dominance of C3 plants implying a stronger monsoon between 40 ka BP to 18 ka BP (Rajagopalan et al., 1997).

7.3 Climate Phase 3

Rapid deterioration of climate led to the formation of sandy-loess blankets. Dry climates, low to negligible vegetation biomass densities favoured soil erosion. The soil particles in turn were mobilised and entrained via dry moisture-less winds or dust storms. Such dust storms (haboobs) moulded the loess into blankets and mounds. The winds appear to have been the southwest monsoon winds as is revealed from the foraminiferal record of these sediments. The change from Climate Phase 2 to Phase 3 is quite sudden. Radiocarbon dates on the red soil nodules between 26-23 ka BP along with thermoluminescence age on the contemporaneous Langhnaj dune of 20 ka BP (Wasson et al., 1983) puts an age bracket of origin between 20-18 ka BP. The upper limit corresponds to ages of dune stabilization. This age may be arrived at from various sources. First is the distribution of Mesolithic sites in Gujarat which is usually on the loessic mounds. The advent of Mesolithic man in Gujarat is synchronous with regional climate amelioration. The Mesolithic period in western India is dated to range between 8-4 ka BP (Agrawal et al., 1978) although Allchin et al., (1978) extend it back to 10 ka BP. In Maharashtra, rivers responded by aggrading their beds between 17 ka BP and 10 ka BP after which incision took place till 4.5 ka BP. This aggradation is explained by Kale & Rajaguru (1987) as an indication of a weakening of the Indian monsoon. Tropical peats of the Nilgiris demonstrate a weakening of the monsoon between 16 to 9 ka BP after which it intensifies once again, coinciding with the global Holocene optimum (Sukumar et al., 1993; Rajagopalan et al., 1997).

In Rajasthan and northern Gujarat initiation of dune stabilization varies between 6.5 ka BP to 5.8 ka BP (Wasson et al., 1983). Pollen analyses of lake sediments from Lunkaransar, Rajasthan (Swain et al., 1983) asserts to precipitation values 200 $mmyr^{-1}$ higher than modern values during 10.5 ka BP and 3.5 ka BP. At Didwana pollen analysis of a 5.8 m core provided a climate record extending back to the last glacial maxima (Singh et al., 1990). For the study the taxa were grouped into three ecologically significant categories shown in table 7.1.

ECOLOGICAL GROUP	TAXA
Desert steppe and shrubland	*Aerva, Ephedra, Calligonum, Chenopodiaceae, Amaranthaceae*
Savanna grassland	*Gramineae, Artemisia, Oldenlandia, Prosopis*
Wetland	*Cyperaceae, Typha, Cosmarium*

Table 7.1: Ecological groups and their taxa recorded from the Didwana core (Singh et al., 1990)

They distinguished 6 zones. Zone D1 spans the period from the last glacial maxima to 13 ka BP. During this period *Aerva*, *Chenopodiaceae/Amaranthaceae*, grasses, *Ephedra* and *Artemisia* covered the landscape. The presence of halite suggests hypersaline lake conditions. Zone D2 from 12.8 ka BP to 9.3 ka BP was a period in which shrub savanna grassland established itself. This is manifested in higher values of *Calligonum* and a drop in Aerva. Zone D3 (9.3-7.5 ka BP) was marked by a steep rise in sedges (*Cyperaceae*) and *Typha* (wetland taxa). From 7.5-6.2 ka BP mainly trees (*Prosopis*) and shrubs from the sub-humid zone moved in forming a tree savanna. In zone D4 (6.2-4.2 ka BP) *Calligonum* reappears and *Artemisia* and *Oldenlandia* values declined. *Typha* remained present indicating moderately deep freshwater conditions in the lake. Few samples from zone D6 suggest very low values for *Oldenlandia* and *Artemisia* by ca 4 ka BP. The surface samples from the site shows a return of high *Aerva* values similar to Late Pleistocene levels. This suggests that the climate from the last glacial to 13 ka BP was hyperarid. A trend towards increasing precipitation is maintained till 6 ka BP. At around 6 ka BP precipitation levels were considerably higher than present (twofold high). The present day climate regime established itself around 4.2 ka BP.

In the Tibetan ice core (Thompson et al., 1997), $\delta^{18}O$ values are low at -20 ‰ between 30 to 15 ka BP. After ca 15 ka BP the values range between -10 to -12 ‰. Global atmospheric carbon dioxide levels during climate phase 3 ranged between 194 ppmv and 252 ppmv (Fig. 7.3) with a progressive increase through time (Barnola et al., 1987). Insolation too was very high during this phase by 45 Wm^{-2} as compared to present values (Berger, 1978). Sea surface temperatures in the Indian Ocean shifted to as low as ~24 °C and remained the same till 15 ka BP (Bard et al., 1997). Oxygen isotopic data from the Arabian sea also reveals a weaker monsoon spanning between 20580 yr BP and 9900 yr BP after which there was a rapid shift towards an active monsoon (Sirocko et al., 1993). Stable isotope analysis of G. ruber also from the Indian Ocean away from the upwelling areas of the Arabian coast also support a weaker summer monsoon during the last Glacial maxima (Duplessy, 1982).

In the Kalahari desert groundwater tables fell between 24-16 ka BP and desiccation was most marked. A period of 2 kyrs between 19 and 17 ka BP was one of increased aeolian activity after which climates ameliorated for a period of 7 kyrs till 10 ka BP (Lancaster, 1989). In the Namib desert dunes accreted between 26-20 ka BP and 16-9 ka BP (Stokes et al., 1997).

7.4 Forcing mechanisms of the Indian summer monsoon

From the above account it is clear that the palaeoclimatic record of Gujarat matches well with global record of climate change, right from ice and ocean cores to continental records from desert margins and interiors. The southwest Indian monsoon is governed by numerous parameters and current general circulation models (GCM's) attempt to take into account as many as possible.

Here, various parameters thought to influence the intensity or the timing of the monsoon are reviewed. Insights are obtained both from global circulation models (GCM's) and coupled GCM's (CGCM's) as well as empirical data discussed in previous sections of this chapter.

The monsoonal system represents a dynamic interaction between the atmosphere, oceans and the continents (Clemens et al., 1991). For the present study the time scales of variability under consideration do not warrant a review of the effect of mountain uplift on the Indian monsoon. Numerical simulations (Prell & Kutzbach, 1992) show that a reduction by half of the Tibet-Himalaya orography is sufficient to maintain the monsoon.

This dates the origin of the monsoonal system to 7-8 Ma which coincides with the advent of C4 grasses over Asia (Quade et al., 1989).

The principal parameters that control the southwest Indian monsoon are:

1] Solar variability induced by variations in the sun spot activity,

2] Carbon dioxide, water vapour and dust concentrations in the atmosphere, and

3] Insolation changes induced by orbital forcing.

The energy emitted through radiation from the sun is not constant, and undergoes subtle changes during a full solar cycle of 1 Wm^{-2} (Diaz & Kiladis, 1995). These variations may potentially perturb global climate on decadal to millennial scales. Coupled with the earth's orbital parameters of eccentricity, precession and obliquity the effects of solar (sunspot) variability stand a chance of being amplified. Friis-Christensen & Lassen (1991) demonstrated a high degree of covariance between changes in the land-surface temperatures of the northern hemisphere and the changes in the length of the sunspot cycle (which in turn is connected to solar variability) over the past three decades. An interesting implication is the possibility of variations in solar radiation that are sufficiently large to govern the nature of global climates over longer time scales (Friis-Christinsen, 1993). Due to inherent differences in heat capacities of land and ocean masses, such hypothesized increases in solar activity would enhance the temperature (land-ocean) contrasts leading to a stronger southwest monsoon. In their simulations using the NCAR Community Climate Model, Prell & Kutzbach (1992) observed a linear increase in monsoon precipitation with solar radiation. For every 1% increase in solar radiation there was a concomitant 2% increase in precipitation.

Sea surface temperatures which are to a large extent controlled by the radiation budget affect the rate of ice-cap melting in the tropics. Due to an increase in the moisture content of the troposphere induced by a rise in sea surface temperature, water vapour induces warming over the Tibetan plateau leading to an acceleration in the melting (Diaz & Graham, 1996). Sea surface temperatures in the North Atlantic region also influence the behaviour of the Indian monsoon (Overpeck et al., 1996). Due to northward shifts in the North Atlantic polar front, their model shows that a warming over the Tibetan plateau takes place which could sufficiently reduce snow cover in early summer and permit a stronger monsoon.

One is all too aware of the detrimental role of increased carbon dioxide concentrations on the global climate and the resultant 'Greenhouse effect'. Atmospheric carbon dioxide concentrations in the past 160000 years (Barnola et al., 1987) have nowhere been near to the post-industrialization values of around 350 ppmv. The highest levels that were present during the last interglacial were 50 ppmv less than the modern values. However, climatologists have strived to determine the effects of doubled CO_2 concentrations on the global climate. Prell & Kutzbach (1992) report that in their simulations the sensitivity of global temperature to CO_2 concentrations decreases with increasing CO_2; but the initial response is linear till 1000 ppmv. Coupled ocean-atmosphere models (Meehl & Washington, 1993) show that area-averaged precipitation increases by 6.3 % due to faster warming of the land mass relative to oceans due to doubling of atmospheric CO_2. Forward-modelling using a similar approach (Manabe & Stouffer, 1993) reveals a increase in global air surface temperature by 3.5 °C over a period of 500 years that also results in a 1 m rise in sea level due to thermal expansion. Although the simulations predict global warming under doubled and quadrupled carbon dioxide concentrations which are not really relevant to climate changes over the past 125 kyrs., these models suggest a lowering of temperatures with drop in CO_2 levels as was the case for example during the last glacial stage in Mainland Gujarat.

Genthon et al., (1987) opine that climate changes triggered by changes in insolation are amplified by orbitally induced CO_2 changes, and the former in isolation may actually be a weak forcing parameter. Although atmospheric concentrations of CO_2 determined by Barnola et al., (1987) and Shackleton et al., (1992) exhibit a phase lag of 8 kyrs, numerical simulations by Berger et al., (1993) show that this does not affect the time of reaching maximum ice volumes. These results, they propose, show a dependence of ice volumes to insolation changes rather than CO_2 variations. Thus during Climate Phase 1 in Mainland Gujarat higher pCO_2 levels were instrumental in generating a more intense monsoon which amplified orbitally-controlled insolation changes.

Water vapour, which constitutes the most important greenhouse aerosol contributes in a major way to climate change. Lesser concentrations of water vapour induces a weakening of the greenhouse effect in such a way that it's inclusion as a determinative parameter in model simulations, changes the amount of cooling at the last glacial maxima by 66 % when coupled with albedo changes (Berger et al., 1993).

The role of dust in the troposphere has only recently been appreciated (Overpeck et al., 1996). Due to increased loading of dust in the atmosphere backscattering of solar radiation decreases when the aerosol overlies snow/ice covered areas (high albedo). An increase occurs in the backscattering when dust overlies dark land/ocean surfaces. Greenhouse warming is also effected by absorption of thermal radiation. In Mainland Gujarat such dust laden atmospheres during Aggradation Phase 3 (Climate phase 3) reduced land temperatures by backscattering thus contributing to a decrease in the land-ocean temperature gradient leading to a weakening of the Indian monsoon.

Spectral analyses of ice and ocean core records resolved three consistent periodicities. One was a 23000 year cycle, the other a 41000 year cycle and the last a 100000 year cycle. This consistency led palaeoclimatologists to look for external mechanisms for forcing such cyclic changes. This led to the revival of the Milankovitch theory (Berger, 1994). This theory was first proposed by Milutin Milankovitch a Yugoslavian astronomer. The central tenet of this theory is that orbital parameters of the earth exert significant influences on the net solar flux received on earth. The orbital parameters identified are eccentricity which is a measure of the ellipticity of the earth's orbit, obliquity, the angle between the earth's axis and the perpendicular to the ecliptic and lastly the longitude of the perihelion which reflects the earth-sun distance for a given season and is termed precession (Berger, 1994). Of these, the obliquity (23000 years) and precession (41000 years) cycles are more effective than the eccentricity (100000 years) cycle in controlling the radiation budget of the earth. In the stratigraphic record of Mainland Gujarat only the eccentricity cycle is readily observable which led to the deterioration of climate over the past 125 kyrs.

Through their influence on this budget these external parameters control the timing and duration of ice ages as first demonstrated by Hays et al., (1976). However the conundrum "whether the external orbital forcing drives the internal forcing, phase-locks the oscillations of an internally driven system, or acts as a pace-maker for the free oscillations of an internally driven system remains an open question" (Berger, 1994).

From the preceding discussion the complexity and feedback mechanisms involved in governing the timing and strength of the southwest Indian monsoon is all too evident. The connections between tropical and high latitude climates has been demonstrated by Sirocko et al. (1996), who state "The observation that not all abrupt climate changes are synchronous between Antarctica and Greenland but that the subtropical monsoon climate shares features with the climatic evolution in both polar hemispheres suggests that the cause of large-scale climatic changes might lie in the low latitudes possibly mediated to the high latitudes by means of water vapour content". The implied teleconnections between subtropical and high latitude climates may alter earlier views on the independence of the monsoon strength in relation to the global ice budget.

7.5 A multiple-state hypothesis of climate change

The stratigraphic record of Mainland Gujarat reveals three climate states. The insensitivity of continental ecosystems to millennial scales adds a new dimension in our understanding of how rapid changes in climate affect such ecosystems. The sharp stratigraphic boundaries between Aggradation Phases 1, 2 and 3 highlight the abruptness of these events. It is as if these three states were separated by certain thresholds, which when achieved altered the nature of the ecosystem. Further, high frequency climate-changes appear to be oscillations or reverberations constrained within these states. In the present study the states reflect progressively deteriorating climates (incremental weakening of the monsoon).

The present study suggests that continental ecosystems respond in a subdued manner. Within an inter-stadial, brief periods of colder climates/weaker monsoon do not affect either the nature of the vegetational community, or the hydrological budget of the area. A brief duration of weaker monsoon as seen in the Qinghai record (Thompson et al., 1997) of around 5 kyrs separated a return to conditions similar to those at the Eemian till 90 ka BP. This does not manifest itself in Aggradation Phase 1 (Climate Phase 1) during which vertisols represent the palaeosol record. The core record during Aggradation Phase 2 (Climate Phase 2) is marked by longer periods of a weaker monsoon (20 kyrs) punctuated by short wetter interludes of 5 kyrs. The latter again is not obvious in the stratigraphic record. However, a stronger monsoon between 50-25 ka BP is manifested as a red-bed forming event. This suggests that for climate shifts to effect durable changes in the continental ecosystem, a period between >5 kyrs and 20 kyrs is the minimum time required to permanently change the landscape of an area.

Recently, a mathematical model which incorporates the concept of threshold conditions has been put forward by Paillard (1997). The main facets of this model are as follows:

The model is conceived within the framework of the Milankovitch theory in which insolation changes effect climate change on earth. The model originates from a similar three tiered state in the oceanic thermohaline circulation. Three distinct states called **i** (interglacial), **g** (mild glacial) and **G** (full glacial) are identified. The climate system is allowed to enter into either of these regimes when ice volumes and insolation reach critical values. A **i-g** transition happens through a critical fall in insolation whereas a **g-G** change effects through a critical change in ice volume. A **G-i** transition occurs through insolation increases. These transitions occur in a specific path which begins with **i** and ends with **G** through **g**.

This multiple-state model predicts correctly the beginning of 100 kyrs glacial cycles at 0.8 to 1 Ma and also explains why in certain situations some stages are warmer although corresponding insolation and ice volumes suggest otherwise. The model shows the incongruence between insolation maxima and interglacials in isotopic records, the latter typically occurring after smaller peaks in insolation.

The cause of the natural system to jump between states appears to be mystifying. Paillard (1997) contemplates that the ocean may be responsible; in particular the north Atlantic thermohaline circulation. This in turn would also affect the intensity of the southwest Indian monsoon (Overpeck et al., 1996). This study shows that a similar three tiered system also occurs on the continents, each tier representing a different ecosystem.

Aggradation Phase 1 corresponds to the interglacial **i**, Aggradation Phase 2 to the mild glacial **g** and Aggradation Phase 3 to the full glacial **G**.

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