

# Chapter 14

## Evolution and Development of the Indian Monsoon



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**Abstract** The Indian monsoon is a complex oceanic-atmospheric-coupled mechanism of the tropics that plays a key role in inter-hemispheric heat transfer on Earth. The summer monsoon brings moisture to the highly populated South Asian countries and affects the livelihood of more than a billion people. The intensity of the monsoon significantly influences the ecological diversity and hydrological reservoirs across the South Asian region. However, the intensity of the monsoon greatly varies spatially and temporally, driven by both external and internal forcing factors. Modeling and palaeoclimatic studies indicate several phases of strong and weak summer monsoon rainfall caused by changes in solar insolation, snow accumulation in western Europe, El Niño-Southern Oscillation, North Atlantic Oscillations and sea surface temperature in the Indian and Pacific Oceans. The initiation and strengthening of the Indian monsoon during the middle-late Miocene are sometimes linked with phases of major surface uplift of the Himalayan and/or Tibetan Plateau. The Plio-Pleistocene glaciation prompted a strong winter monsoon and a weak summer monsoon. During the early Holocene, the summer monsoon strengthened and subsequently weakened with two major phases of sudden rainfall reduction at ~8.2 and ~4.2 kyr BP; the latter event caused significant societal impact including the migration of population of the Indus Valley Civilization. In the last millennium, the Indian summer monsoon (ISM) was strong during the Medieval Warm Period (MWP) now designated as Medieval Climate Anomaly (MCA) and Current Warm Period (CWP), punctuated by a weak phase during the Little Ice Age (LIA). Meteorological records indicate an increasing trend in the intensity and frequency of extreme rainfall events in the last few decades leading to widespread floods and droughts. High-resolution climatic records from marine as well as continental archives improve our

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understanding of Indian monsoon variability and its forcing factors on different time scales.

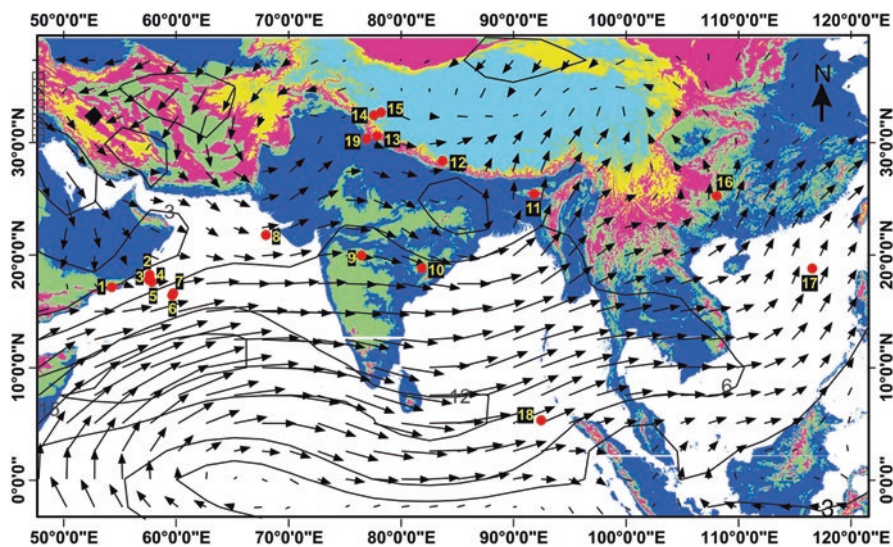
**Keywords** Indian monsoon · Marine and terrestrial archives · Early Miocene climate warmth · Plio-Pleistocene · Holocene

## 14.1 Introduction

The word monsoon is derived from an Arabic word *Mausim* meaning an inter-hemispheric, land-atmosphere-ocean-coupled mechanism in the tropics that shows seasonal reversal in wind direction, poleward westerly flow in summers and equatorward easterly flow in winters. The summers are generally associated with the wet season; and winters are relatively dry. The tropical monsoon system plays an important role in heat and energy distribution on Earth. The global monsoon comprises five major individual monsoon systems, i.e. the North American monsoon, the South American monsoon, the African monsoon, the Australian monsoon and the Asian monsoon. The Asian monsoon has two major components namely the South and East Asian monsoons. The monsoon winds bring moisture to ~63% of the global area and support the livelihoods of a great part of the population.

The Indian monsoon is one of the largest and most complex climatic phenomena on Earth. During the boreal summers, cross-equatorial southeasterlies flow from the Indian Ocean, entering the Indian subcontinent from the southwest, known as the southwest (SW) Indian monsoon or Indian summer monsoon (ISM) (Fig. 14.1). In boreal winters, the winds blow from the Indian subcontinent to the Indian Ocean from a northeast (NE) direction and are called the NE Indian or winter monsoon. The ISM supplies around 80% of the annual precipitation throughout the Indian subcontinent, except in a few regions of southern and western India (Gadgil 2003). The NE Indian monsoon winds are relatively dry but acquire some moisture while crossing the Bay of Bengal and bring precipitation to parts of southern India and Sri Lanka (Gadgil 2003).

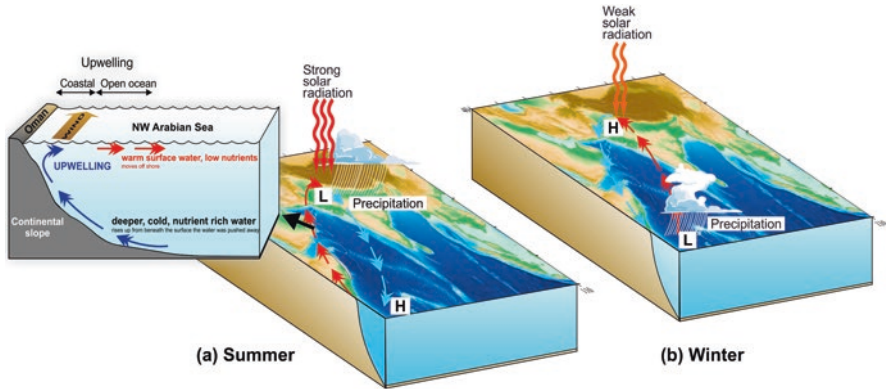
ISM precipitation plays a crucial role in the socio-economic wellbeing of the agrarian-based, densely populated countries in South Asia. In India, the agricultural output contributes ~20% of its Gross Domestic Product (GDP) which is largely dependent on summer monsoon rainfall (Gadgil 2003). A slight deviation of ~10% from normal ISM rainfall affects the agricultural production and water availability across the region. A variation of ~30% can cause serious disasters in the form of extreme floods and widespread droughts associated with famines (Shewale and Kumar 2005) such as the Kedarnath flood of 2013, the Kashmir flood of 2014, the Kerala flood of 2018 and drought in several regions in the country in 2002. Not only the rainfall amount but the timing of the ISM onset and retreat affects the agricul-



**Fig. 14.1** Map showing the Indian monsoon wind pattern in the month of July and mean 850 mb pressure level using the 1981–2010 climatological base period. (Source: [http://iridl.ldeo.columbia.edu/maproom/Global/Climatologies/Vector\\_Winds.html](http://iridl.ldeo.columbia.edu/maproom/Global/Climatologies/Vector_Winds.html)). Also shown is the location of study sites from the Indian and East Asian monsoon regimes. (1) Oman cave (Fleitmann et al. 2003), (2) ODP Hole 730A (Gupta et al. 2015), (3) ODP Hole 728B (Gupta et al. 2015), (4) ODP Hole 723A (Gupta et al. 2003), (5) R2730 (Anderson et al. 2002), (6) ODP Hole 731A (Gupta et al. 2015), (7) ODP Hole 722B (Gupta et al. 2015), (8) 2502G (Agnihotri et al. 2002), (9) Lonar Lake (Prasad et al. 2014), (10) Jhumar cave (Sinha et al. 2011), (11) Mawmluh cave (Berkelhammer et al. 2012; Dutt et al. 2015), (12) Dasuopu ice core (Thompson et al. 2000), (13) Sahiya cave (Sinha et al. 2015), (14) Tso Moriri Lake (Leipe et al. 2014), (15) Chandra peat bog (Rawat et al. 2015), (16) Dongge cave (Dykoski et al. 2005; Wang et al. 2005a, b), (17) ODP Site 1148 (Clift et al. 2008), (18) ODP Site 758 (Gupta and Thomas 2003), and (19) Indian Siwaliks (Sanyal et al. 2004)

tural output of the region. The ISM winds normally reach the west coast of India in the first week of June. A delayed onset of summer monsoon in 2002 caused widespread droughts and a 13% reduction in food grain production in India (Shewale and Kumar 2005).

Owing to its critical importance in the sustenance of society, the ISM has been studied extensively since the development of the earliest civilizations in this region. The ancient Indian text *Rig Veda* mentioned the concept of monsoon as a reversal in wind direction. Another Indian epic *Meghaduta* described the movement of clouds and the onset of monsoon rain in different regions of India with great precision regarding the present meteorological timings. Scientific study of the Indian monsoon in modern times began with the work of Sir Edmund Halley. Halley (1686) described the Indian monsoon as a ‘giant land sea breeze’ that arises due to differential heating of the Indian landmass and the Indian Ocean. In the Northern Hemisphere summers, the differential heating of the ocean and continental area develops a pressure gradient between the Indian Ocean and South Asia, with low pressure over central and western India and high pressure in the equatorial Indian



**Fig. 14.2** Schematic diagram illustrating the Indian monsoon circulation during (a) summer and (b) winter and coastal upwelling off the Oman margin and northwest Arabian Sea during the summer. L represents low pressure and H represents high-pressure zones

Ocean (Fig. 14.2a) (Fein and Stephens 1987; Quade et al. 1995). As a result, a low atmospheric flow of moisture laden, southwesterly winds originate in the Indian Ocean and move towards the Indian landmass before being deflected in a right direction, after crossing the equator under the influence of the ‘Coriolis effect’ (Fig. 14.1) (Webster et al. 1998; Gadgil 2003). On the contrary, during Northern Hemisphere winters, the pressure gradient reverses leading to the development of weak and variable northeasterly winds blowing from the landmass to the ocean (Fig. 14.2b). The geometry of the Indian landmass facilitates bifurcation of the ISM winds into two branches, the Arabian Sea and the Bay of Bengal branches (Fig. 14.1). The Arabian Sea branch first strikes the Western Ghats of India and then moves further towards the low-pressure areas of the Indian Peninsula and Ganga Plain. The Bay of Bengal branch initially strikes NE India and moves westward following the strike of the Himalaya along the Ganga Plain (Gadgil 2003).

The differential heating of the land and ocean associated with the solar insolation has been suggested as the primary causative factor for deriving the monsoonal circulation in south Asia three centuries ago (Halley 1686; Charney 1969). But this hypothesis has been criticized by some authors because there is no direct correlation between land-sea thermal contrast and ISM precipitation. The ISM winds are strongest during July and not in the hottest month of June. Furthermore, the hottest region of northwest India has desert conditions and receives very little monsoonal rainfall. An alternative hypothesis has been proposed that the latitudinal movement of the Inter Tropical Convergence Zone (ITCZ) (a narrow latitudinal belt of clouds) is a major forcing factor in controlling the strength of the ISM (Gadgil 2003). The seasonal latitudinal migration of the ITCZ determines the onset and withdrawal periods, frequency and intensity of the rainy season in the tropics (Gadgil 2003). During boreal summers, the maximum solar insolation lies at  $23.5^{\circ}\text{N}$  (Tropic of Cancer), so that the ITCZ moves towards the low-pressure areas in these regions, resulting in the movement of moisture laden winds from the Indian Ocean towards the Indian

subcontinent. During boreal winters, the maximum insolation lies at 10°S, and the ITCZ moves towards the low-pressure areas of the southern hemisphere, resulting in the retreat of the monsoon and winds blowing from the NE to the SW (Gadgil 2003).

The heating of the Tibetan Plateau has also been suggested as a major forcing factor in the modulation of the ISM (Molnar et al. 1993). However, recent modeling studies indicate that the Himalayan and Tibetan Plateau provide a mechanical barrier to the ISM winds so that this monsoon system would be predicted to have existed even in the absence of the Tibetan Plateau (Boos and Kuang 2010, 2013). It is argued that the Tibetan Plateau helps in holding the monsoonal winds in the Indian subcontinent which otherwise would have moved towards Central Asia. The heating of the low elevation areas of Central India and the Ganga Plain is more effective in driving the Indian monsoon than the heating of the Tibetan Plateau (Boos and Kuang 2013).

Although the ISM is a regular mechanism, its intensity, time of onset and retreat, as well as its spatial and temporal strength, are highly variable. Due to its importance and variability, a lot of scientific research has been undertaken to understand the mechanism and forcing factors of the ISM. The instrumental record of the last ~150 years depicts several phases of strong and weak summer monsoon in India (Shewale and Kumar 2005). This also indicates the influence of solar insolation, snow cover in Western Europe, the El Niño-Southern Oscillation (ENSO) and the Indian Ocean Dipole (IOD) on the ISM at interannual to decadal time scales. Nonetheless, long-term climatic changes cannot be determined with short-scale meteorological investigations. Therefore, climatic variability prior to the instrumental period is derived from natural archives (ocean, river, lake, glacier, speleothems, etc.) which preserve climatic signals in the form of different physical, chemical and biological proxies.

## 14.2 Forcing Factors and Global Teleconnections

Due to critical importance of the ISM to socio-economic conditions and ecological controls in South Asia, intense efforts have been made over the last three centuries to understand the controlling mechanisms and forcing factors of this inter-hemispheric atmospheric feature. Multiple archives and multi-proxy palaeoclimatic reconstructions, as well as modeling studies, indicate the role of various external (solar forcing) and internal (e.g. ocean-atmospheric teleconnections, El Niño-Southern Oscillation) factors that influence ISM variability on different time scales. Whereas factors such as solar variability, ENSO, IOD and Himalayan snow cover lead to interannual and multidecadal or centennial scale changes, Himalayan-Tibetan tectonics and orography drove longer-scale (million-year) changes in the Indian monsoon. Earth's orbital parameters such as precession, obliquity and eccentricity, on the other hand, have controlled millennial or astronomical scale changes in the Indian monsoon.

### ***14.2.1 Solar Variability***

Solar variability or changes in solar insolation reaching the Earth's surface play a significant role in controlling ISM strength. Solar variability has two major variables, one related to orbital forcing and the other to solar irradiance (Imbrie et al. 1984). Variations due to orbital forcing occur because of changes in the Earth's orbit around the Sun (eccentricity) and its own axis (obliquity and precession), which operate at 100 kyr, 41 kyr and 23 kyr time scales, respectively (Laskar et al. 2004). These changes caused by variability in eccentricity, obliquity and precession affect hemispheric distribution of solar insolation on Earth's surface and relative strength of the seasons. Periods of low/high solar insolation in the Northern Hemisphere cause expansion/melting of ice sheets in polar and high-altitude regions of the Himalaya, and/or glacial/interglacial cycles on Earth, which affect the land sea thermal contrast, ITCZ migration, evaporation and convection over the Indian Ocean and ultimately the strength of ISM winds (Clemens and Prell 2003). The ISM becomes weak during times of glacial boundary conditions and strong during interglacial periods (Prell and Kutzbach 1992). Another important factor in solar variability is changes in solar irradiance. With the present advanced technology, changes in solar activity can be measured directly. However, geological proxy records of  $^{10}\text{Be}$  and  $^{14}\text{C}$  radionuclides have been used for reconstructing past solar activity (Wanner et al. 2008). Several high-resolution proxy records of ISM variability suggest that solar forcing has been one of the strongest factors controlling the ISM on decadal to millennial time scales (Agnihotri et al. 2002; Gupta et al. 2005, 2013a).

### ***14.2.2 Eurasian and Himalayan Snow Cover***

Blanford (1886) was the first to suggest an inverse relationship between snow cover in the Himalaya and Eurasia and ISM rainfall over India—a hypothesis that has been further supported by later workers. Blanford (1886) suggested that the greater the winter snowfall, the weaker is the ISM in the following summer and vice versa. A thick snow cover in the Himalaya and Eurasia increases the albedo of land surface, and a larger part of the solar radiation is then used in melting snow rather than heating the land so that a low-pressure trough fails to develop in Northern and Central India. As a result, high snow deposition in the Himalaya and Eurasia would be expected to lead to reduced rainfall in India in the succeeding summer monsoon season. However, modeling studies indicate that only the snow cover in western Eurasia has an inverse correlation with the summer monsoon rain in India (Bamzai and Shukla 1999).

### ***14.2.3 El Niño-Southern Oscillation (ENSO)***

The El Niño-Southern Oscillation or ENSO is an anomalous condition in sea surface temperature coupled with atmospheric circulation in the tropical eastern Pacific Ocean, which appears over intervals of 3–7 years (Sikka 1980; Gadgil 2003). The oceanic parts of this system are El Niño (warm) and La Niña (cold), marked by sea surface temperature anomalies in the tropical eastern Pacific Ocean greater than 0.5 °C. The Southern Oscillation is the accompanying atmospheric component, coupled with the sea temperature change. El Niño is accompanied by low and La Niña by high atmospheric pressure in the tropical eastern Pacific Ocean. The ENSO is one of the strongest factors affecting the global climate on interannual scales. The El Niño events are known to cause changes in the intensity and amount of rainfall in those regions which are influenced by the Asian summer monsoon (Berkelhammer et al. 2012). In the Indian subcontinent, several episodes of prevalent droughts and famines have occurred during the El Niño years (Sinha et al. 2011). However, this relationship is not maintained throughout (Ashok et al. 2001), and normal monsoon rainfall has been observed even during intense El Niño years (e.g. in years 1963 and 1997). This was probably due to the influence of the positive IOD mode in the Indian Ocean (Gadgil 2014; Krishnaswamy et al. 2014).

### ***14.2.4 The Indian Ocean Dipole***

The Indian Ocean Dipole or IOD is an anomalous state of the sea surface temperature like ENSO in which an east-west temperature gradient prevails in the Indian Ocean (Saji et al. 1999; Ashok et al. 2001; Krishnaswamy et al. 2014). A positive IOD is characterized by higher sea surface temperatures over the western Indian Ocean and lower temperatures in the east. A positive IOD is associated with higher rainfall, while negative IODs correlate with lower precipitation over the Indian subcontinent and the Arabian Sea (Ashok et al. 2001). A positive IOD phase acts as a modulator to the ISM, driven by the ENSO events. Modeling studies indicate that frequent occurrence of the IOD during the last few decades has weakened the linkage between the ISM and ENSO (Saji et al. 1999; Ashok et al. 2001; Kumar et al. 2006).

### ***14.2.5 The Midlatitude Westerlies***

The interaction between midlatitude westerlies and ISM winds leads to subseasonal or intraseasonal variations in monsoon precipitation (Krishnan et al. 2009; Rajeevan et al. 2010). This variability occurs with the ‘active and break cycles’ of the Indian monsoon (Sikka 1980). In Central India, during the ‘active cycle’, the monsoonal

rainfall increases, whereas during 'break cycle', weak rainfall is observed even during the peak monsoon period. The phenomenon of 'breaks' has become more important because of its relationship with the intraseasonal variability of monsoonal precipitation. During the break phase, the interaction between westerlies and the ISM pushes the low-pressure trough towards the Himalayan foothills, resulting into reduced rainfall all over South Asia except for in the Himalayan region (Gadgil 2003; Rajeevan et al. 2010).

### ***14.2.6 The North Atlantic Oscillation***

Several studies have documented the impact of North Atlantic Oscillation (NAO) on the global climate on different time scales. Records of ISM variability also demonstrate marked changes in ISM intensity on interdecadal, multidecadal and centennial time scales. Cooling events in the North Atlantic (Heinrich events) have been linked with marked decreases in monsoonal rainfall in the Indian Ocean and Indian subcontinent (Gupta et al. 2003; Dutt et al. 2015). North Atlantic cooling events influence the ISM by increasing Eurasian snow cover through the westerlies' mediated mechanism or by weakening the land-sea thermal contrast (Gupta et al. 2003; Fleitmann et al. 2003; Dutt et al. 2015).

### ***14.2.7 Changes in Atmospheric Composition***

Both natural (e.g. volcanism, CO<sub>2</sub> degassing from land and oceans, forest fires) and anthropogenic factors (e.g. industrial, fossil fuel burning) can bring pronounced changes in the atmospheric composition, which can affect strength of the ISM. The concentration of CO<sub>2</sub> in the atmosphere critically affects the global climate by changing the atmospheric temperature. An increase in atmospheric CO<sub>2</sub> concentrations raises the atmospheric temperature, causing higher evaporation and convective activities in the Indian Ocean. This in turn leads to a stronger ISM (Officer and Drake 1983; Prell and Kutzbach 1992), although such changes are temporary and may hold for one or a few seasons.



## 14.3 Palaeoclimatic Archives in the Reconstruction of ISM Variability

### 14.3.1 Marine Archives

The rivers from South and Southeast Asia together with winds from the Arabian Peninsula carry and deposit large quantities of sediment in the Arabian Sea and the Bay of Bengal every year. The fossil skeletons of calcareous and siliceous microorganisms thriving in the ocean water are deposited on the seafloor along with clastic sediments, and in situ chemical precipitates. These sediments continue to be extensively studied by taking cores from the Arabian Sea and the Bay of Bengal. The type, abundance and geochemical character of these sediments provide much information concerning the climatic conditions that have prevailed in the adjacent continental regions as well as the ocean surface at the time of sedimentation (Clemens et al. 1996; Govil and Naidu 2011).

A close relationship has been established between ISM wind intensity and abundance of cold water, subpolar planktic foraminifer species *Globigerina bulloides*, which is deposited in sediments offshore the coast of Oman in the Arabian Sea during intervals of monsoon-driven intense upwelling (Kroon et al. 1991; Prell and Kutzbach 1992). The summer monsoon winds displace surface water offshore Oman, resulting in upwelling of nutrient-rich, cold sub-surface water to the surface, which promotes the growth of phytoplanktons as well as zooplanktons such as an opportunist planktic foraminifer *Globigerina bulloides* (Kroon et al. 1991; Prell and Kutzbach 1992). Thus, abundance of *Globigerina bulloides* in the sediments reflects upwelling intensity in the Arabian Sea and in turn the strength of the summer monsoon winds (Overpeck et al. 1996; Gupta et al. 2003; Curry et al. 1992). The isotopic ratios (oxygen and strontium isotopes) and elemental abundances (Mg/Ca, Ba/Ca ratios, etc.) of the calcareous shells of foraminifera further provide clues to climatic variability on different time scales (Saraswat et al. 2013).

### 14.3.2 Continental Archives

#### 14.3.2.1 River Sediments

River-derived sediments are one of the sensitive terrestrial archives for understanding both climatic and tectonic changes. The grain size and sedimentation patterns of fluvial sediments provide information about changes in the flow regimes and river discharge during sedimentation. In the absence of either tectonic influences or sea level changes, phases of aggradation in river valley flood plains generally indicate high sediment load and high water discharge from the source mountains associated with strong ISM precipitation. The slack water deposits in the river (sedimentary feature forms during intrusion of water from the river trunk stream to a tributary)

indicate large floods at that time (Kale 2007). In addition, the floral and faunal remains deposited in the floodplains are indicative of the climatic conditions in the region.

Numerous studies have been carried out on the Indian rivers indicating the role of climate and tectonics on their sedimentation pattern (Gibling et al. 2005, 2008; Juyal et al. 2010; Ray and Srivastava 2010; Scherler et al. 2015; Kumar and Srivastava 2017). Late Quaternary valley aggradation has extensively been observed in the Himalayan rivers, but it is still debated as what was the role of strong/weak southwest monsoon phases on these rivers (Juyal et al. 2010; Ray and Srivastava 2010; Scherler et al. 2015; Kumar and Srivastava 2017). Singh et al. (2017) observed channel avulsion of the Sutlej River in the Himalayan foreland after 8 ka in the weak monsoon phase. Other studies from the Mahi, Sabarmati and Luni rivers implicate climate as a dominating factor for valley aggradation and incision (Jain and Tandon 2003; Jain et al. 2004). Jain et al. (2004) also suggested that the rivers in the desertic region are more sensitive to climate variability than rivers in the monsoon margin regions.

#### 14.3.2.2 Speleothems

Speleothems are the secondary mineral deposits formed in caves by the precipitation of calcium carbonate from water percolating through the cave. They may be deposited in different forms such as stalagmites, stalactites and flowstone that capture the signals of regional precipitation via lamina thickness, fluorescence, elemental abundance (Sr, U, etc.) and isotopic signatures (Fairchild et al. 2006). Speleothems have proved to be the best continental proxy providing very high-resolution record of climate change (Wang et al. 2005a, b; Sinha et al. 2011; Dutt et al. 2015). The oxygen isotopic values of speleothems in tropical regions are derived from the meteoric precipitation, which is controlled by changes in the moisture source, transport pathways and fractionation at the vapour source (Dansgaard 1964; Breitenbach et al. 2010). All these factors result into relatively more negative oxygen isotope ratios ( $\delta^{18}\text{O}$  values) in the precipitation during periods of strong monsoon compared with rainfall during weak phases. These signals are preserved in the speleothem carbonate precipitation (Breitenbach et al. 2010; Dutt et al. 2015). Furthermore, the speleothems can be dated precisely with U-Th series dating methods that help in retrieving annual and multiannual scale information of climate change. The millennial scale variability of the Indian monsoon has been reconstructed using speleothems from the Himalayan regions (Dutt et al. 2015; Kathayat et al. 2016). Recent studies by Kathayat et al. (2018) have also reconstructed the structure and timings of 4.2 kyr BP event at the annual resolution that marks the beginning of a long arid interlude in the Indian subcontinent.

### 14.3.2.3 Lake Sediments

Lakes are highly sensitive to climate change and are the most widely used terrestrial archives for the reconstruction of regional climatic variability on different time scales and resolution. Lake sediments can be dated by varve counting, AMS  $^{14}\text{C}$  dating techniques, U-Th series dating and optically stimulated luminescence (OSL) dating techniques. Changes in the sediment grain size, varve thickness, abundance of organic and inorganic chemical concentrations, environmental magnetism, mineral assemblages, faunal and floral remains, isotopic characteristics of lake sediments and contemporary floral and faunal remains reflect the variability in lake level, regional hydrology and temperature, rainfall amount, moisture source, vegetation type and intensity (Leipe et al. 2014; Dixit et al. 2014a, b, 2018; Prasad et al. 2014; Singh et al. 2015; Dutt et al. 2018).

### 14.3.2.4 Glaciers

The Himalayan region has the maximum snow accumulation outside the North and South Poles; hence it is also known as the ‘Third Pole of the World’. The expansion and retreat of the glaciers indicate changes in the regional and global climate on longer (centennial to millennial) time scales. Glaciers retreat during warm climates and expand in cold climatic conditions, which can be understood by analysing isotopic signatures in glacial ice cores (Stuiver and Grootes 2000). Glaciers archive the climatic signal in geomorphological features, relict lakes, fluid inclusions trapped in the ice, dust transported from lower latitudes and isotopic signatures (Thompson et al. 2000; Hedrick et al. 2011). Glacial moraines are the geomorphological indicators of former glacier expansion (Hedrick et al. 2011; Bisht et al. 2015). Debris cover over the glaciers also affects the response of glaciers to climate change (Scherler et al. 2011). Changes in oxygen isotope compositions in ice cores from the Himalaya have been used to indicate variability of atmospheric temperature and the moisture sources, i.e. the ISM or midlatitude westerlies (Thompson et al. 2000).

### 14.3.2.5 Tree Rings

Tree rings are one of the most precise archives to develop high-resolution climate variables for the last few millennia, as growth ring sequences could be dated exactly to the calendar year of their formation. Trees which grow in the seasonal climate such as low-latitude monsoon or mid- to high-latitude temperate regions produce distinct annual growth rings, consisting of early and late woods formed in early and late growing season, respectively. The ring widths, wood density and chemical constituents (stable isotopes of carbon, hydrogen, oxygen) of precisely dated growth ring sequences from several geographic locations in India and adjoining countries under the influence of variable degrees of summer monsoon strength have revealed the importance of tree ring data to reconstruct precipitation and drought indices for

the past few centuries and millennia (Ramesh et al. 1985, 1989; Treydte et al. 2006; Shah et al. 2007; Cook et al. 2010; Managave et al. 2010, 2011, 2017; Sano et al. 2012, 2013; Shi et al. 2014; Yadava et al. 2016; Yadav et al. 2017). However, thus far reconstruction of monsoon season precipitation/drought indices based on tree ring parameters is limited but essentially required to understand its pre-instrumental natural variability. The annually resolved summer monsoon precipitation reconstructions are usually short, limited to the past few centuries (Shah et al. 2007; Sano et al. 2012, 2013) except that of Cook et al. (2010) and Shi et al. (2014) spanning over the past several centuries and millennia. These reconstructions (Shah et al. 2007; Cook et al. 2010; Sano et al. 2012, 2013; Shi et al. 2014), consistent with the observed summer monsoon precipitation from sites close to respective proxy locations, revealed subdued precipitation in recent decades consistent with the instrumental records (Sontakke et al. 2008). The centennial components in the South Asia summer monsoon index (Shi et al. 2014) show a strong summer monsoon phase during the Medieval Climate Anomaly (MCA, AD ~950–1250) and a weak summer monsoon during the Little Ice Age (LIA) (AD ~1400–1700). The drought indices (Palmer Drought Severity Index) for the summer (June–July–August) derived using tree ring chronologies from over 300 sites of monsoon Asia (Cook et al. 2010) have shown that the megadroughts are linked to tropical Pacific sea surface temperature (SST) forcing. However, the South Asia summer monsoon index (Shi et al. 2014) is found to be directly correlated with total solar irradiance (TSI) on multidecadal time scales implying that solar activity has strong control on the South Asia summer monsoon.

#### 14.3.2.6 Peat Bogs

Peat bogs are common in high-altitude cold regions of the Himalaya. Changes in pollen abundance, as well as isotopic and geochemical characteristics observed in peat deposits, reflect variability in temperature, as well as vegetation type in the region. These in turn are the results of changes in ecological and climatic factors (Phadtare 2000; Rawat et al. 2015). Phadtare (2000), Rawat et al. (2015) and Srivastava et al. (2017) have reconstructed climatic changes in the NW Himalaya during the Quaternary using peat records and identified intervals of wet and dry phases in the ISM. Late Quaternary climate variability in southern India has been reconstructed using carbon isotope signatures of peat deposit from the Nilgiri Hills (Sukumar et al. 1993).

#### 14.3.2.7 Aeolian Deposits and Palaeosols

Aeolian sediments are deposited as loess and sand dunes in arid and semi-arid regions typically where annual rainfall is less than 250 mm. In India, aeolian deposits are common in the *Thar Desert* which have been studied intensely to understand climatic variability in NW India (Thomas et al. 1999; Juyal et al. 2006). The

development of sand dunes and loess takes place in arid environment where evaporation is higher than precipitation. Phases of higher rainfall might be expected leading to vegetation growth that binds the soil, ceasing dune activity and stabilizing the dunes to allow formation of soil. However, provenance analysis of sand dunes in the *Thar Desert* reveals them to be largely derived from the Indus Delta during wetter interglacial phases (East et al. 2015). While this may seem counterintuitive, these are times of strong SW monsoon winds that facilitate transport of material from the delta into the continental interior, which is still relatively dry. The chronology of active and dormant phases of dune activity, vegetation growth and loess-palaeosol cycles represent the climatic changes in the region at that time interval. Thomas et al. (1999) suggest that the *Thar Desert* was once a region of high rainfall and high vegetation density. Similarly, the loess-palaeosol sequence of the Tibetan Plateau provides a record of climatic variability, including past ISM-westerlies interactions (Maher and Hu 2006).

#### 14.4 Evolution of the Indian Monsoon System and Driving Mechanisms

Earth's climate has undergone a significant change from being relatively warm and equable to colder with nearly freezing temperatures at the poles during the Miocene (Kennett and Barker 1990; Lear et al. 2000; Zachos et al. 2001). It has been argued that the ocean basins had essentially assumed their modern shapes by the Early Miocene with relatively modest changes since that time (Kennett and Barker 1990). A permanent thermal gradient was established between the poles and the tropics that allowed the formation of the modern wind regimes and surface water masses during the Middle Miocene (Kennett 1977). The opening and closing of different seaways including the final closure of the Tethyan Seaway (Lawver and Gahagan 1998), which caused meridional turnovers in the ocean circulation, were also observed during the Miocene (Verma et al. 2013). The Early Miocene climatic warmth (Weissert et al. 1984), Middle Miocene Antarctic ice volume expansion (Kennett and Barker 1990; Zachos et al. 2001) and changes in ocean circulation (Verma et al. 2013) influenced the deep and surface oceans in numerous ways. These changes had significant influence on regional and global climates and ecology, including the ISM (Lear et al. 2000; Barker 2001; Zachos et al. 2001; Gupta and Thomas 2003; Lawver and Gahagan 2003; Barker and Thomas 2004; Singh and Gupta 2004, 2005, 2010; Singh et al. 2012; Gupta et al. 2013b; Verma et al. 2013).

Inception and development of the Asian monsoon system, which comprise Indian or South Asian and East Asian monsoons, was one of the most important climatic events to affect the tropics during the Neogene. Whereas the Indian monsoon is driven by the thermal contrast between land in the north and ocean in the south, the East Asian Monsoon is regulated by land in the north and south, Tibet in the west and open ocean in the east (Wang et al. 2005a, b). During the last few decades,

numerous studies on the Asian Monsoon have been undertaken, with major international programs such as INDEX (Indian Monsoon Expedition, 1975–1976) and the MONEX (International Winter/Summer Monsoon Experiment, 1978–1979) in the Indian Ocean (Krishnamurti 1985), South China Sea Monsoon Experiment (SCSMEX, 1998–2002; Lau et al. 2000; Ding and Liu 2001) and International Ocean Discovery Program (IODP) Expedition 346, which were exclusively focused on the Asian Monsoon around the Sea of Japan (Tada et al. 2015). IODP expeditions 353/354 and 355 collected marine sediments from the Bay of Bengal and Arabian Sea, respectively, and were exclusively focused on the Indian monsoon and reconstruction of history of Himalayan and Tibetan Plateau (H-TP) erosion and weathering (Clemens et al. 2015; Pandey et al. 2016).

The evolution and long-term changes in the ISM have been linked to the phased uplift of the H-TP (Rea 1992; Prell et al. 1992; Molnar et al. 1993; Prell and Kutzbach 1997; Filippelli 1997; An et al. 2001), uplift of the H-TP combined with Antarctic glaciations and global climate changes (Gupta et al. 2004) and movement of Inter Tropical Convergence Zone (ITCZ) with respect to the Indian Plate (Armstrong and Allen 2011; Allen and Armstrong 2012). The collision of the Eurasian and Indian plates uplifted the H-TP, which has been proposed to result in the onset or major intensification of the ISM (Hahn and Manabe 1975), as well as the present-day climate system over the Indian Ocean region during the Miocene (Ramstein et al. 1997; Blisniuk et al. 2001; Clift and Gaedicke 2002; Rowley and Currie 2006; Clift et al. 2008). The uplift of the H-TP strengthened the land-sea thermal contrast that causes the summer monsoon with heavy rainfall over the Indian subcontinent (e.g. Hastenrath 1991; Boos and Kuang 2010) with cold dry winters. Those areas of rapid exhumation in the Himalaya have been correlated with zones of the most intense summer monsoon (Thiede et al. 2004), which led to the transportation of large volume of sediments to the ocean. However, the timing of the Himalayan uplift is poorly constrained in relation to the climate change (Gupta et al. 2004). Furthermore, links between precipitation and erosional exhumation have been questioned in several parts of the Himalayan front (Burbank et al. 2003).

The elevated heat source of the Himalaya and the Tibetan Plateau was earlier suggested to be of vital importance for the establishment and stability of Indian summer monsoon circulation through mechanical and thermal factors (Molnar et al. 1993). Recent climate models suggest that Tibet's role is limited, at least with regard to the ISM, and by removing Tibet, the ISM largely remains unaffected provided the narrow orographic barrier of the Himalaya is preserved (Boos and Kuang 2010). It is suggested that monsoon is also sensitive to changes in surface heat fluxes from non-elevated regions of the Indian landmass as well as to changes in heat fluxes from adjacent elevated regions (Boos and Kuang 2013).

## 14.5 Timing of Initiation and Intensification of the Indian Monsoon

Considerable efforts have been made towards high-resolution reconstruction (via high density sampling of marine cores) of proxy records of the Indian monsoon that have helped in understanding the monsoon evolution, its variability over various time scales and forcing factors that drive the monsoon on orbital and suborbital time scales. However, there are still unresolved questions as to the timing of the advent of the modern monsoon and driving mechanisms that control monsoon variability. The elevated heat source of the H-TP is of vital importance, but there are different models concerning the timing of the H-TP attaining the critical elevation in order to drive the Indian monsoon. These range from 55 to 7.5 Ma (Table 14.1). While the majority of marine records indicate a major shift in the monsoon system between 13 and 8 Ma, as well as earlier after 24 Ma, continental records suggest a range from 55 to 7.5 Ma during which time the monsoon may have evolved (Table 14.1). However, some of these studies make simplistic links between H-TP elevation or chemical weathering and the monsoon. Climate modeling, on the other hand, predicts the origin at ~35 Ma (Rowley and Currie 2006). Studies from China suggest a wet phase in the Early Miocene and beginning of an arid phase (weakening of the summer monsoon) across 13–11 Ma (Jiang and Ding 2008).

Changes in benthic foraminiferal assemblages suggest that deep-sea conditions gradually became less oxygenated with sustained flux of organic matter since the Middle Miocene (Singh and Gupta 2004). The increase in benthic foraminifer *Uvigerina proboscidea* abundances during 9–8 Ma in the eastern Indian Ocean roughly corresponds to CaCO<sub>3</sub> MAR and a negative shift in  $\delta^{13}\text{C}$  coinciding with the major intensification of the Indian monsoon system, as observed in several land and marine proxies, mostly linked to wind intensity (Quade et al. 1989; Kroon et al. 1991; Gupta and Srinivasan 1992; Prell et al. 1992). The high productivity during this time probably depressed the carbonate compensation depth (CCD) in the Indian Ocean (Peterson and Backman 1990), and mixed layer started thickening (Singh and Mohan 2007). The continuous occurrence of *U. proboscidea* between ~8.4 and 7.6 Ma and 6.5–5 Ma at Deep Sea Drilling Project (DSDP) Site 214 and Ocean Drilling Program (ODP) Site 757 suggests a major increase of nutrients in deep waters (Gupta and Srinivasan 1992; Gupta et al. 2004; Singh and Gupta 2004) during the Miocene. Burbank et al. (1993), however, observed a decline in physical weathering in the Himalaya correlated with the proposed intensification of the monsoon. These authors explain this unexpected negative correlation to be due to reduced tectonic activity, decreased Himalayan glaciations or stabilization of slope from dense plant cover. Later workers have, however, proposed that Late Miocene reductions in erosion rate are linked to reduced and not increased rainfall (Clift et al. 2008). After 5 Ma, monsoon began to weaken with fluctuations and gradually attained the modern day conditions during the Pliocene (Sanyal et al. 2004).

A few studies have suggested a major intensification of the Indian monsoon roughly at the same time as the ‘biogenic bloom’ in the late Miocene (Chen et al.

**Table 14.1** Evidence of evolution of the Indian monsoon system and Himalayan uplift during the Cenozoic (modified from Gupta 2010)

Reference	Evidence	Event	Timing (Ma)
Shukla et al. (2014)	Terrestrial record	South Asian monsoon	55–52
Licht et al. (2014)	Oxygen isotopes, Myanmar	Monsoon	40
Rowley and Currie (2006)	Oxygen isotope	Tibetan Plateau	35
Ramstein et al. (1997)	Modeling	Monsoons and Paratethys retreat	~30
Srivastava et al. (2012)	Megaflora	South Asian monsoon	~28–23
Betzler et al. (2016)	Marine sediments	Proto monsoon to monsoon	~25–20
Guo et al. (2002)	Terrestrial record	Asian desertification	22
Wang (1990)	Sediments in China	Monsoons	20
Clift et al. (2002)	Marine sediments	Himalayan exhumation and Asian monsoon intensity	15.5
Clift et al. (2008)	Weathering South China Sea and Indus Fan	Monsoons	24
Spicer et al. (2003)	Fossil flora	Himalayan elevation and monsoons	>15
Blisniuk et al. (2001), Coleman and Hodges (1995)	Tectonics	Himalayan uplift and upwelling	>14
Gupta et al. (2015)	Multi-proxy oceanic records	Monsoon wind and upwelling	12.9
Chen et al. (2003)	Oceanic microfossils	Monsoons and upwelling	12–11
Sanyal et al. (2010)	Carbon isotopes from land record	Monsoons	~11
Zhuang et al. (2017)	Biomarker	Monsoonal upwelling	11–10
Dettman et al. (2001)	Isotopes and land	Monsoons	~10.7
An et al. (2001)	Land and marine sediments	Uplift and onset of evolution	9–8
Kroon et al. (1991)	Planktic foraminifera	Monsoons and upwelling	8.6
Filippelli (1997)	Weathering and sediments	Monsoons	~8
Quade et al. (1989)	Carbon isotopes from land record	Monsoons	8–7.5

2003; Gupta et al. 2004). However, several studies proposed the inception of the Asian monsoon across the Oligocene-Miocene boundary (Ramstein et al. 1997; Guo et al. 2002; Jia et al. 2003; Sun and Wang 2005; Clift et al. 2008). Studies based on the pollen records from China suggested enhanced aridity and monsoon establishment around 24 Ma, followed by intensification at 15–13 Ma and ~8 Ma (Sun and Wang 2005). Wan et al. (2006) argued that the monsoon possibly initiated at about 12 Ma or even as early as 22 Ma with intensification until about 8 Ma, but most of the marine proxy records from the Indian Ocean have indicated that the

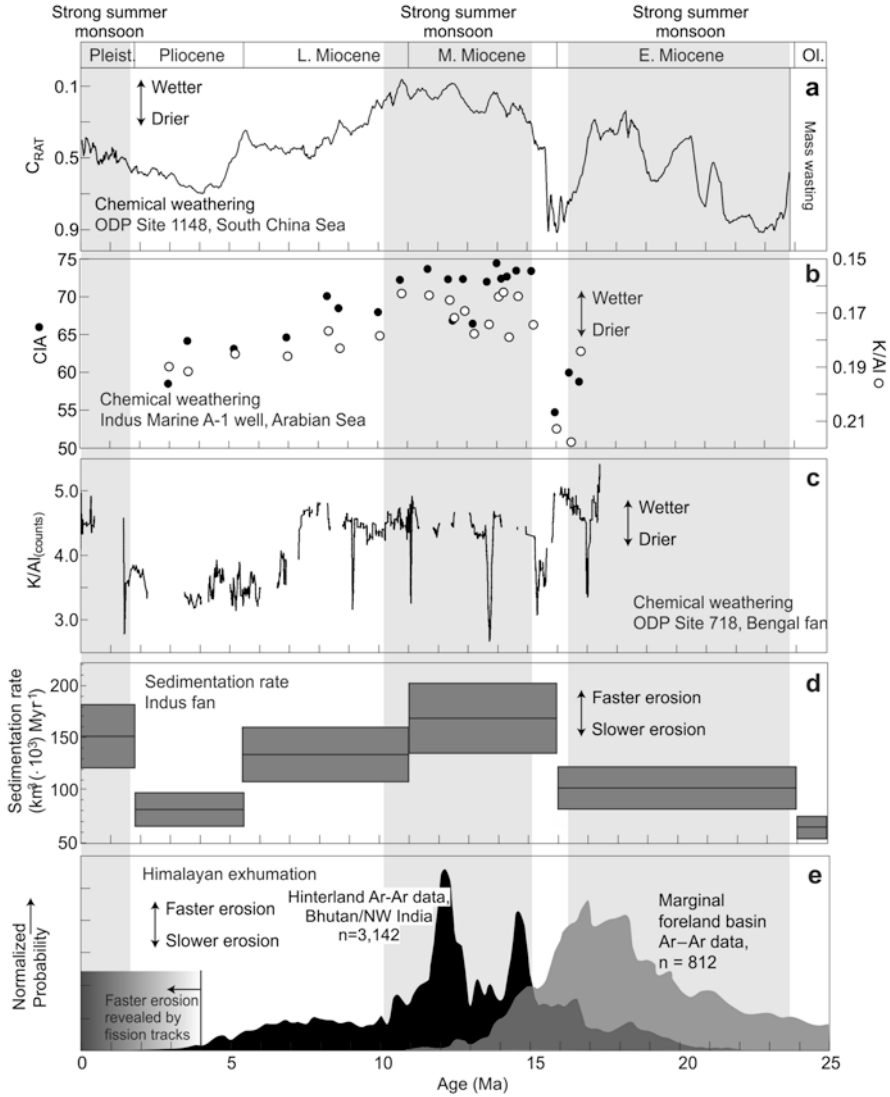


monsoon-related winds and upwelling developed or intensified during 10–8 Ma (Kroon et al. 1991; Gupta et al. 2004). Studies based on chemical and mineralogical indices from South China Sea ODP Site 1148, Indus Marine A-1 well, Arabian Sea and southern bay of Bengal ODP Site 718 suggested that the monsoon initiated across the Oligocene-Miocene boundary (~24 Ma) and intensified until ~10 Ma with brief tropical wet period between 16.5 and 15 Ma, coinciding with the Middle Miocene Climatic Optimum (Clift et al. 2008; Fig. 14.3a–c). A recent study based on *Globigerina bulloides* and stable isotope data from the Oman margin and Owen Ridge indicated a major increase in Indian monsoon winds between 12 and 11 Ma and onset of the Indian monsoon system at about 12.9 Ma (Gupta et al. 2015). These authors further suggest that it was not the H-TP uplift alone that contributed to the inception of the Indian monsoon but the attainment of present-day land-sea geometry of the Indian Ocean, and Asian landmass also played a pivotal role in the initiation of the Indian monsoon wind system (Gupta et al. 2015).

### 14.5.1 Miocene

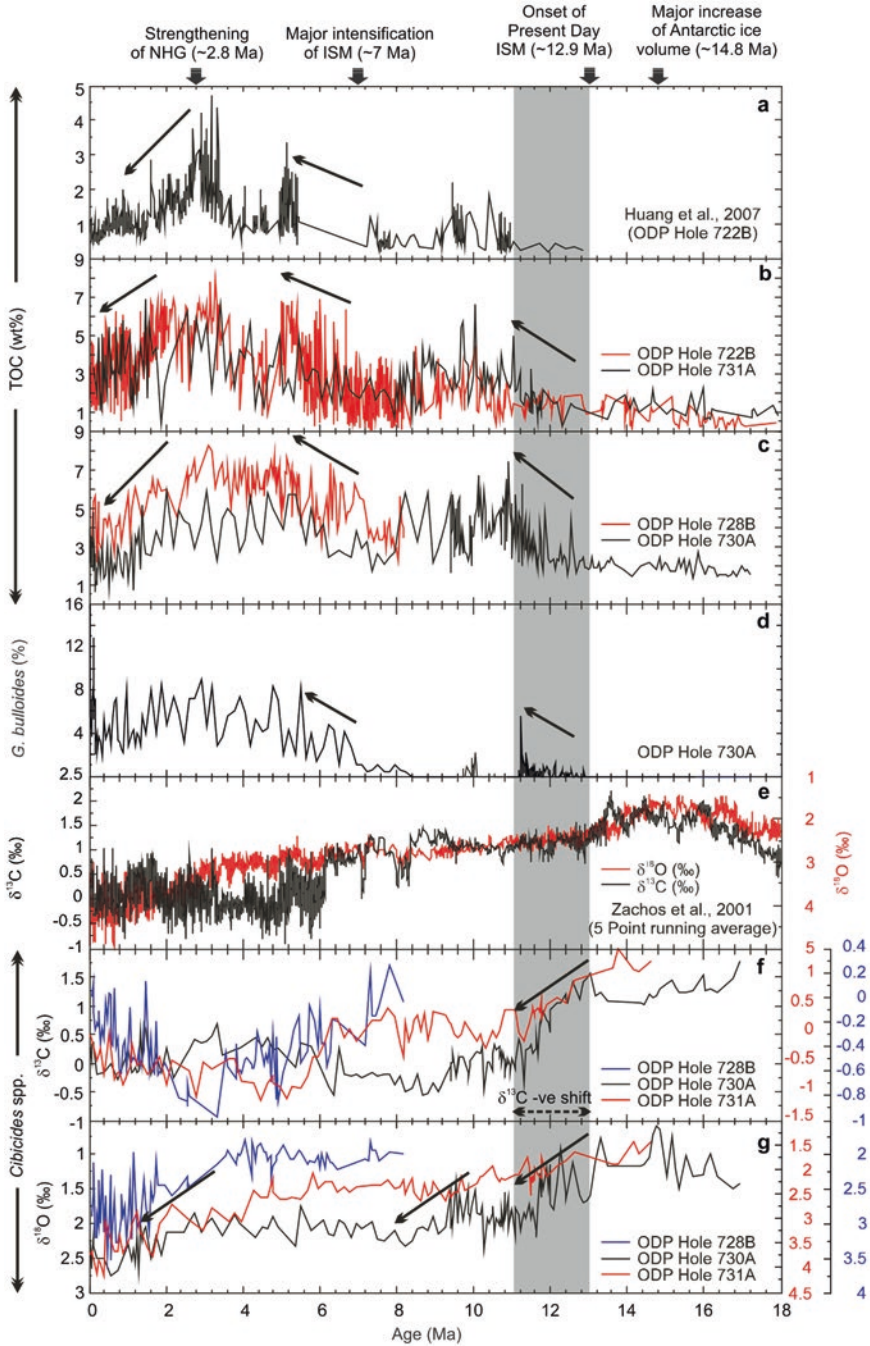
The Miocene epoch is the oldest division of the Neogene period, extending from 23.03 to 5.33 Ma (Gradstein et al. 2012). This is an important epoch for understanding Earth's large-scale climatic and tectonic features such as the Indian monsoon and its linkages with the Himalayan uplift. A long marine record (~23.8 Ma) was used by Clift et al. (2008) to understand the relationship between the Himalayan exhumation rates and the Asian monsoon intensity from the Indus Fan, Bay of Bengal (ODP Site 718) and South China Sea (ODP Site 1148) (Fig. 14.3d, e). These workers mainly used chemical weathering-based proxies such as chemical weathering index ( $C_{\text{RAT}}$ ), K/Al ratio and Chemical Index of Alteration (CIA) to reconstruct past humidity (Fig. 14.3a–c). They also used sedimentation rates as a proxy for erosion rates onshore, arguing for a positive relationship between rainfall and mass flux. The results of  $C_{\text{RAT}}$  from this study suggest five phases of changes in the monsoon strength: intensification during ~23.08–16.5 Ma, 15–10 Ma and 3.5 Ma to Recent. Weakening of the monsoon occurred at 16.5–15 Ma and 10–3.5 Ma. The Arabian Sea and Bay of Bengal sediment proxy records of K/Al ratio and CIA indicate a stronger monsoon from 16 to 10 Ma and a weak monsoon since 10 Ma, which is like the South China Sea records (Clift et al. 2008). The results of sedimentary record from the Himalayan foreland basin shows higher erosion rates from 24 to 18 Ma (Fig. 14.3d), which was interpreted to signify strengthening of the Indian monsoon (Clift et al. 2008).

Recent study by Gupta et al. (2015) has changed the perception regarding the evolution of the South Asian monsoon among palaeoclimatologists. This study used multi-proxy and high-resolution climatic records from the main zone of the Indian monsoon activity in the northwestern (NW) Arabian Sea (ODP Holes 722B, 728B, 730A and 731A). These proxies include planktic foraminifer *Globigerina bulloides* (Figs. 14.4d and 14.5a–e), stable isotopes of benthic foraminifera ( $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ )

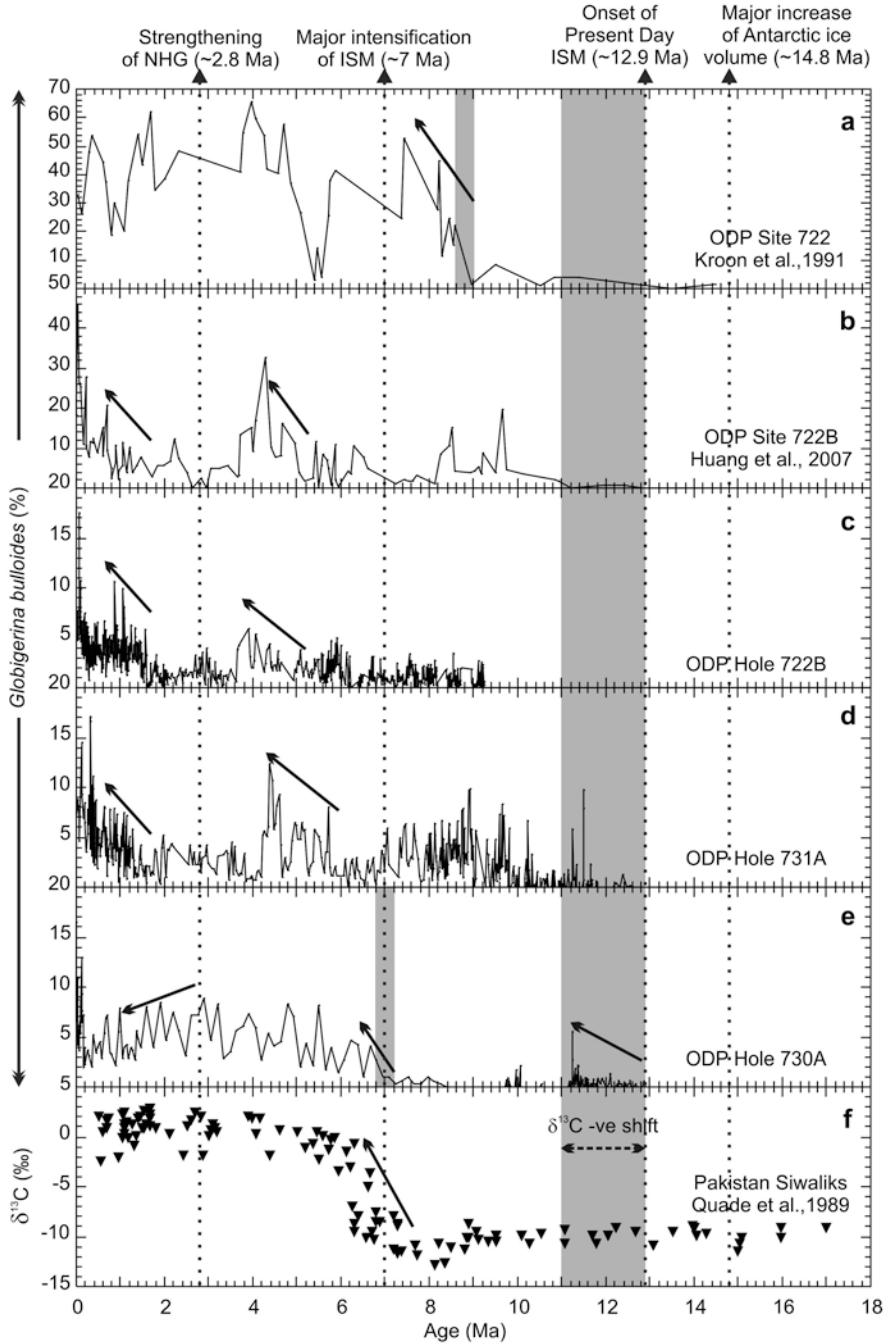


**Fig. 14.3** Correlation of erosional and depositional histories from ODP Site 1148, South China Sea (a), Indus marine A-1 well, Arabian Sea (b and d), ODP site 718, Bengal Fan, Bay of Bengal (c), and Himalayan foreland basin (e). (Adopted from Clift et al. 2008). Light-grey bars indicate strong summer monsoon intervals

(Fig. 14.4f, g) and total organic carbon (TOC) (Wt%) from the NW Arabian Sea (Fig. 14.4a–c). The upwelling proxy planktic foraminifer *G. bulloides* (Anderson and Prell 1993; Gupta et al. 2003) first appears at ~12.9 Ma at Hole 730A on the Oman margin and concurrently at Hole 731A on the Owen Ridge, with a short-lived increase at ~11 Ma (Fig. 14.4d). From 11 to 7 Ma, this species nearly disappeared



**Fig. 14.4** Multi-proxy records of Indian monsoon from ODP Holes 722B, 731A, 728B, 730A, northwestern Arabia Sea correlated with global isotope data (Zachos et al. 2001) (adopted from Gupta et al. 2015). (a–c) Total organic carbon (TOC, Wt%) and (d) *Globigerina bulloides* from ODP Sites 722, 728 and 730, NW Arabian Sea, (e) global isotope record (Zachos et al. 2001) and (f, g) stable carbon and oxygen isotope ratios of benthic foraminifer species *Cibicides* spp. from ODP Sites 728, 730 and 731. Vertical grey bar indicates negative carbon shift; arrows indicate high and low strengths of the Indian monsoon



**Fig. 14.5** Indian monsoon proxy records of *Globigerina bulloides* from Oman margin (ODP Hole 730A) and Owen Ridge (ODP Holes 731A, 722, 722B), northwestern Arabian Sea (panel a–e), and stable carbon isotope record of carbonate deposits from the Potwar Plateau, Pakistan (Quade et al. 1989) (f). (Adopted from Gupta et al. 2015). Vertical dashed lines and grey bars indicate major climatic events

except for a short-lived, minor increase at ~10 Ma on the Oman margin but remains abundant at Owen Ridge ODP Hole 731A. *Globigerina bulloides* shows a pronounced increase at ~7 Ma corresponding to a shift from C<sub>3</sub> to C<sub>4</sub> type vegetation in South Asia (Quade et al. 1989) and remains abundant and variable in the interval younger than 7 Ma at Hole 730A (Figs. 14.4d and 14.5f). The results of total organic carbon (Wt%) show a higher value at ~12.5 Ma in the studied holes almost coinciding with the appearance of *G. bulloides* (Fig. 14.5a–e).

The carbon isotope ( $\delta^{13}\text{C}_{\text{org}}$ ) record of benthic foraminifera from the Oman margin (ODP Hole 730A) shows a major shift towards negative values beginning at ~12.9 Ma, postdating the global  $\delta^{13}\text{C}$  shift by 0.7 Ma, whereas on the Owen Ridge (Hole 731A), the negative  $\delta^{13}\text{C}$  shift coincides with the global negative carbon shift (Zachos et al. 2001) (Fig. 14.4e–g). The coincidence of first appearance of *G. bulloides*, major increase in TOC (wt.%), and  $\delta^{13}\text{C}$  negative shift at Oman margin Hole 730A all suggest intense ISM winds and high organic production during the Middle Miocene. These changes in the proxy records argue for the onset of the Indian monsoon wind system at about 12.9 Ma during the Middle Miocene (Gupta et al. 2015). The inter-site variations in the *G. bulloides* population between Oman Margin and Owen Ridge provide a clue to the shifting wind stresses and thus upwelling zones in the NW Arabian Sea.

A palaeoclimatic evidence of monsoon shifts is also preserved on land, but the preservation of continuous, long climatic records on land is poor as compared to the marine sediments and generally has poorer age control. In the Himalayan foreland basin, the fluvial sediments of the Siwalik succession are deposited from Burma (East) to Afghanistan (West). These deposits represent the most significant terrestrial record that can be used to study palaeoclimatic changes from the Late Miocene to Pleistocene. A stable isotope study of Neogene soil carbonate was carried out by Quade et al. (1989) on the Potwar Plateau (Fig. 14.6), which mirrors vegetation changes on land over the time period as recorded in the Indian, Nepal and Pakistan Siwaliks. Another notable study has been carried out by Sanyal et al. (2004) to understand the Indian monsoon evolution and vegetational changes since the late Miocene (Fig. 14.6). The results were initially interpreted to indicate that the Indian monsoon intensified at ~10.5 Ma and 5.5 Ma. More recently the shift from C<sub>3</sub> to C<sub>4</sub> plants after 8 Ma is taken to indicate the onset of a drier, more seasonal climate. The monsoon strength decreased since 5.5 Ma, transforming it into the modern-day condition. This study also suggests that C<sub>3</sub> plants dominated in the region from 10.5 to 6 Ma, whereas C<sub>4</sub> plants have largely appeared since 6 Ma, which also coincides with the monsoon intensification. The vegetation record from 6 to 1.8 Ma suggests that convergence of C<sub>3</sub> and C<sub>4</sub> plants occurred during this time period.

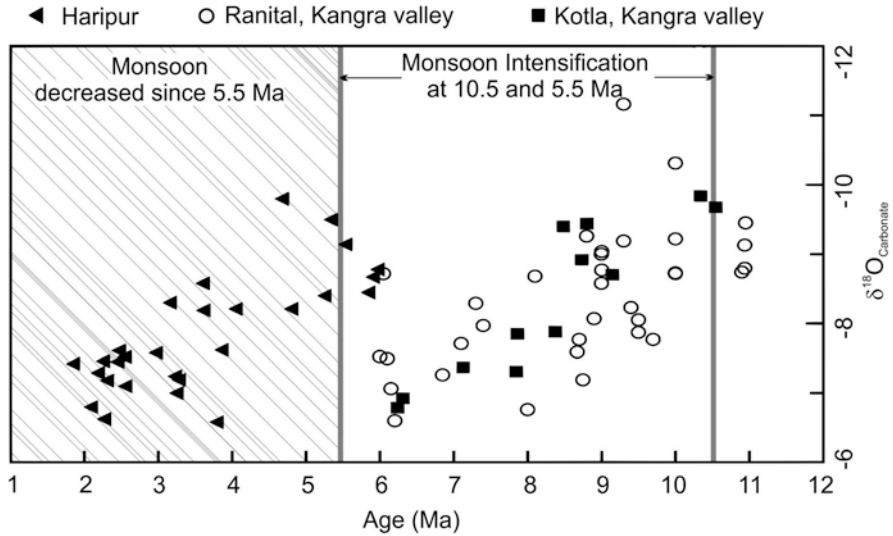
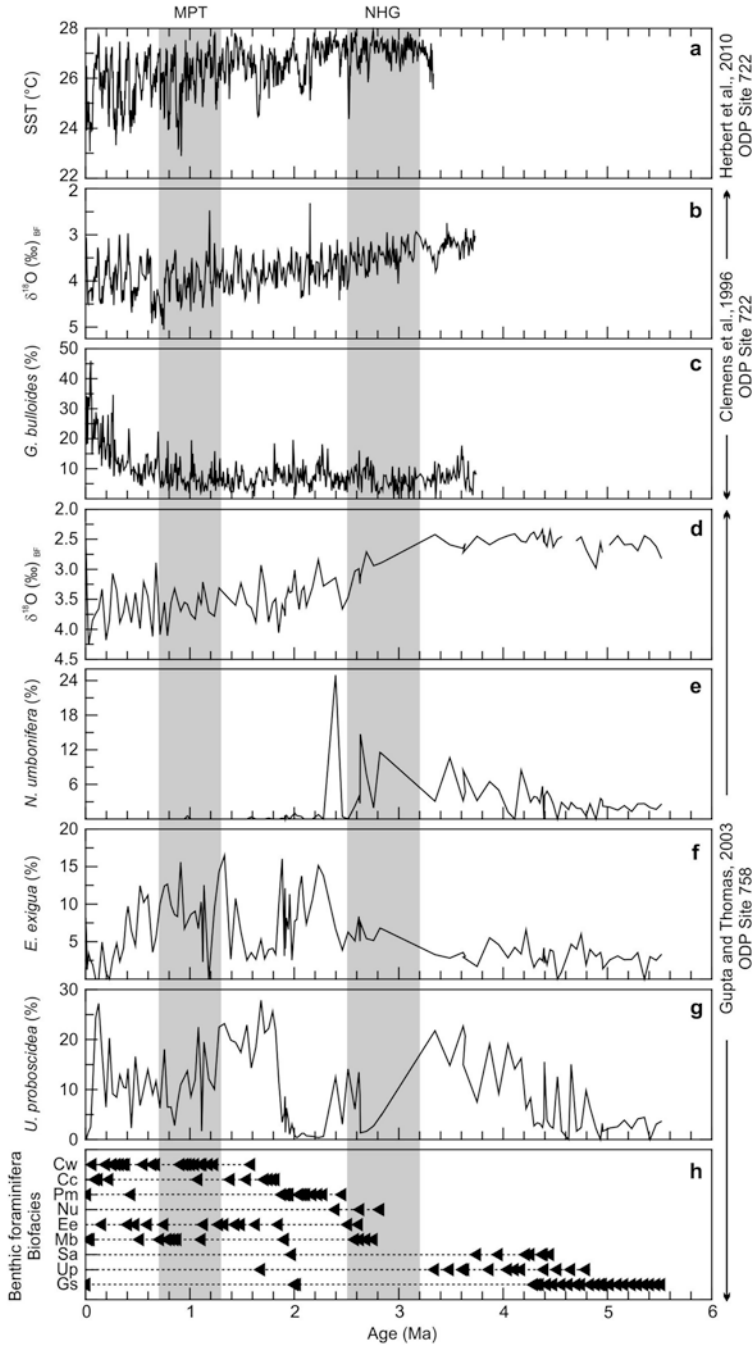


Fig. 14.6 Stable oxygen isotope proxy record of Indian monsoon intensification from Siwalik deposits in India. (Adopted from Sanyal et al. 2004)

### 14.5.2 The Plio-Pleistocene Period

During this period the Earth's climate underwent remarkable changes, i.e. mid-Pliocene warm period (MPWP) (5–3 Ma), onset of the Northern Hemisphere glaciation (NHG) (~3.2–2.5 Ma) and the Mid-Pleistocene Transition (MPT) (~1.25–0.7 Ma). These events are recorded in the ocean and terrestrial archives. In this section we present different proxy records from the study of Gupta and Thomas (2003, benthic foraminifera,  $\delta^{18}\text{O}$  records) from ODP Site 758, Bay of Bengal, (Clemens et al. 1996, *G. bulloides* % and  $\delta^{18}\text{O}$ ) and Herbert et al. (2010, sea surface temperature) from ODP Site 722, NW Arabian Sea to understand Indian monsoon variability during the Plio-Pleistocene interval (Fig. 14.7a–h).

Gupta and Thomas (2003) mainly studied benthic foraminiferal faunal and isotope records (Fig. 14.7d–g), which have earlier been used for palaeoproductivity reconstructions (Smart et al. 1994; Gupta and Thomas 1999). Gupta and Thomas (2003) identified nine biofacies from ODP Site 758, Bay of Bengal, which show significant palaeoenvironmental shifts since 5.5 Ma (Fig. 14.7h). The results of the benthic biofacies indicate low to moderate seasonality and high food and primary productivity during 5.5–3.2 Ma. Thereafter the other six biofacies show intermediate to high seasonality and low to intermediate organic matter contents, together with high seasonality since 3.2 Ma. A major change was observed between 3.2 and 2.5 Ma which suggests that the strength and duration of the winter monsoon increased in this region. This change is synchronous with the initiation of the NHG and expanding East African aridity (deMenocal 1995). Gupta and Thomas (2003)



**Fig. 14.7** Plio-Pleistocene proxy records of Indian monsoon variability in the Arabian Sea. (a) ODP Site 722 (Herbert et al. 2010), (b, c) ODP Site 722 (Clemens et al. 1996) and (d–h) ODP Site 758 (Gupta and Thomas 2003). Grey bars indicate Northern Hemisphere glaciation (NHG) and Mid-Pleistocene Transition (MPT), respectively

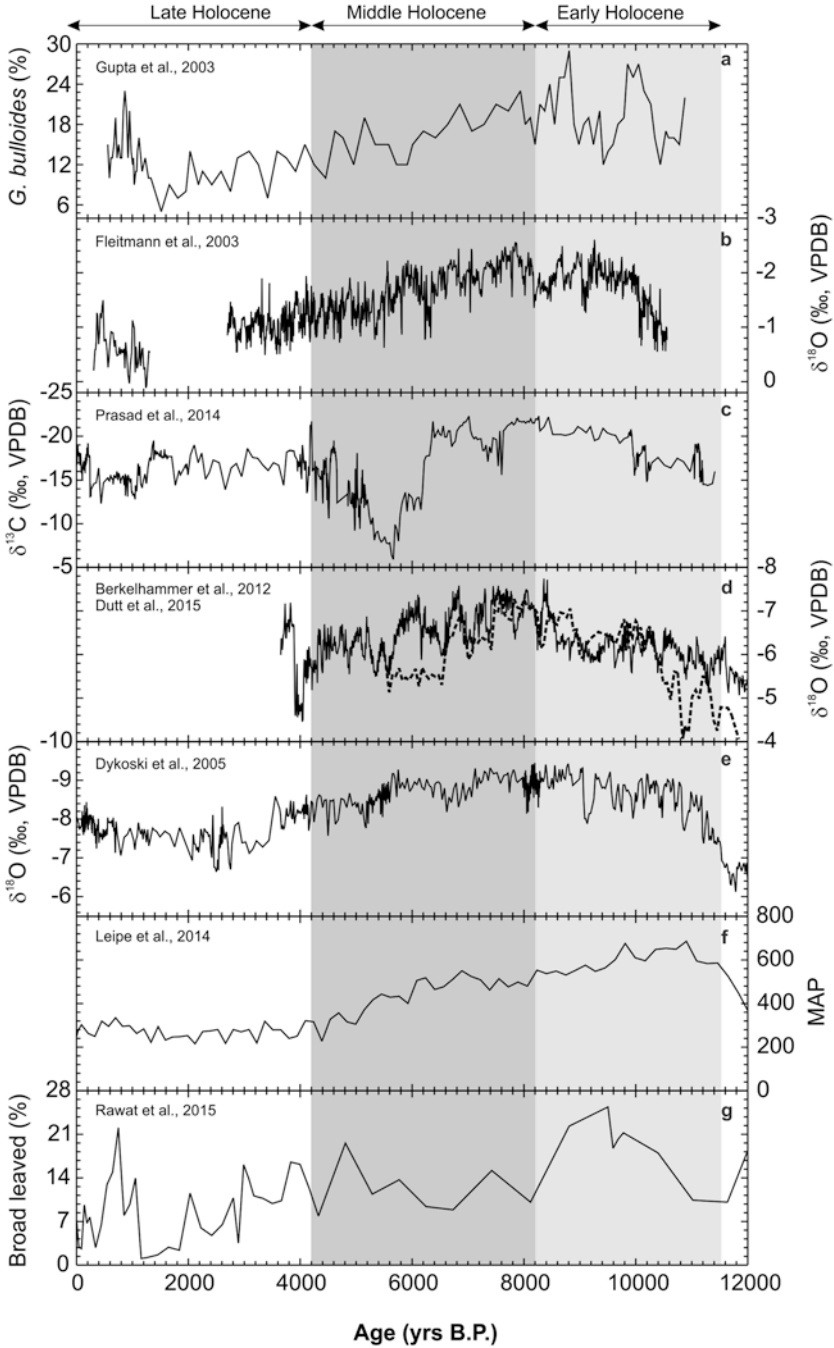
provided an explanation for the strengthening of the winter monsoon, which may have been linked to the shifts in the latitudinal position of the westerly winds and the Siberian high-pressure system (Chen and Huang 1998).

We also used *G. bulloides* (%) and  $\delta^{18}\text{O}$  (‰) data from Clemens et al. (1996) and sea surface temperature (SST) ( $^{\circ}\text{C}$ ) data from Herbert et al. (2010) to understand monsoon strength and phase relationship since 3.5 Ma (Fig. 14.7a–c). During the SW monsoon, the percentage of *G. bulloides* and SST are inversely related in this region, i.e. higher *G. bulloides* populations correspond with lower SST in summers, both caused by stronger upwelling. Clemens et al. (1996) suggest that with the initiation and growth of the NHG, the phase of strong monsoon moves away from the phase maximum ice volume, systematically shifting by  $\sim 83$  and  $\sim 124$  kyr at the precession and obliquity bands, respectively. During the Late Pleistocene, the ice sheets further expanded, triggering another phase of global climate change, and at the same time, the Asian monsoon weakened. Clemens et al. (1996) have argued that the Asian monsoon had a nonstationary phase during the Plio-Pleistocene.

### 14.5.3 The Holocene Epoch

It was earlier believed that the Earth's climate has been relatively stable during the Holocene epoch (last 10,000 years). However, recent studies have indicated that the Holocene was marked by repeated occurrences of abrupt centennial to millennial scale climatic shifts (Fleitmann et al. 2003; Gupta et al. 2003; Dutt et al. 2015) (Fig. 14.8a–g). During the early Holocene, the ISM was strong throughout Asia. High *G. bulloides* production in the Arabian Sea (Gupta et al. 2003); more negative oxygen isotopic values in speleothems from Oman (Fleitmann et al. 2003), NE India (Berkelhammer et al. 2012; Dutt et al. 2015) and China (Dykoski et al. 2005); more negative stable carbon isotopic values from Central India (Prasad et al. 2014); high organic production in the NW Himalaya (Leipe et al. 2014; Rawat et al. 2015); and high discharge to the Bay of Bengal (Govil and Naidu 2011) all indicate a warm and wet climate during the early Holocene caused by strengthening of the ISM (Fig. 14.8a–g). The increase in the ISM in the early Holocene is attributed to increased solar insolation and associated enhanced evaporation in the oceans and northward migration of the ITCZ (Fleitmann et al. 2003; Dutt et al. 2015). The boundary of early and Middle Holocene is marked by a cold and dry climatic event (8.2 kyr BP) (Alley et al. 1997; Dixit et al. 2014b). However, this event is not so prominent in the palaeoclimatic records from the Indian subcontinent. The Middle Holocene (8.2–4.2 kyr BP) climate records indicate asynchronous behaviour of the ISM across Asia. Palaeoclimatic records from the Arabian Sea (Gupta et al. 2003), Oman (Fleitmann et al. 2003), NW Himalaya (Leipe et al. 2014; Rawat et al. 2015) and China (Dykoski et al. 2005; Hu et al. 2013) indicate a consistent weakening of the ISM during the Middle Holocene (Fig. 14.8). However, high summer monsoon intensity has been observed in Central India (Prasad et al. 2014) and NE India (Dutt et al. 2015) followed by a sudden weakening at  $\sim 6.4$  kyr BP (Fig. 14.8).





**Fig. 14.8** Multi-proxy and multi-archive records of Indian summer monsoon variability during the Holocene from (a) ODP Hole 723A, Arabian Sea (Gupta et al. 2003), (b) Oman cave, Oman (Fleitmann et al. 2003), (c) Lonar Lake, Central India (Prasad et al. 2014), (d) Mawmluh cave, northeast India (Berkelhammer et al. 2012; Dutt et al. 2015), (e) Dongge cave, China (Dykoski et al. 2005), (f) Tso Moriri Lake, NW Himalaya (Leipe et al. 2014) and (g) Chandra peat bog, NW Himalaya (Rawat et al. 2015). Different shades indicate different time spans of Early, Middle and Late Holocene

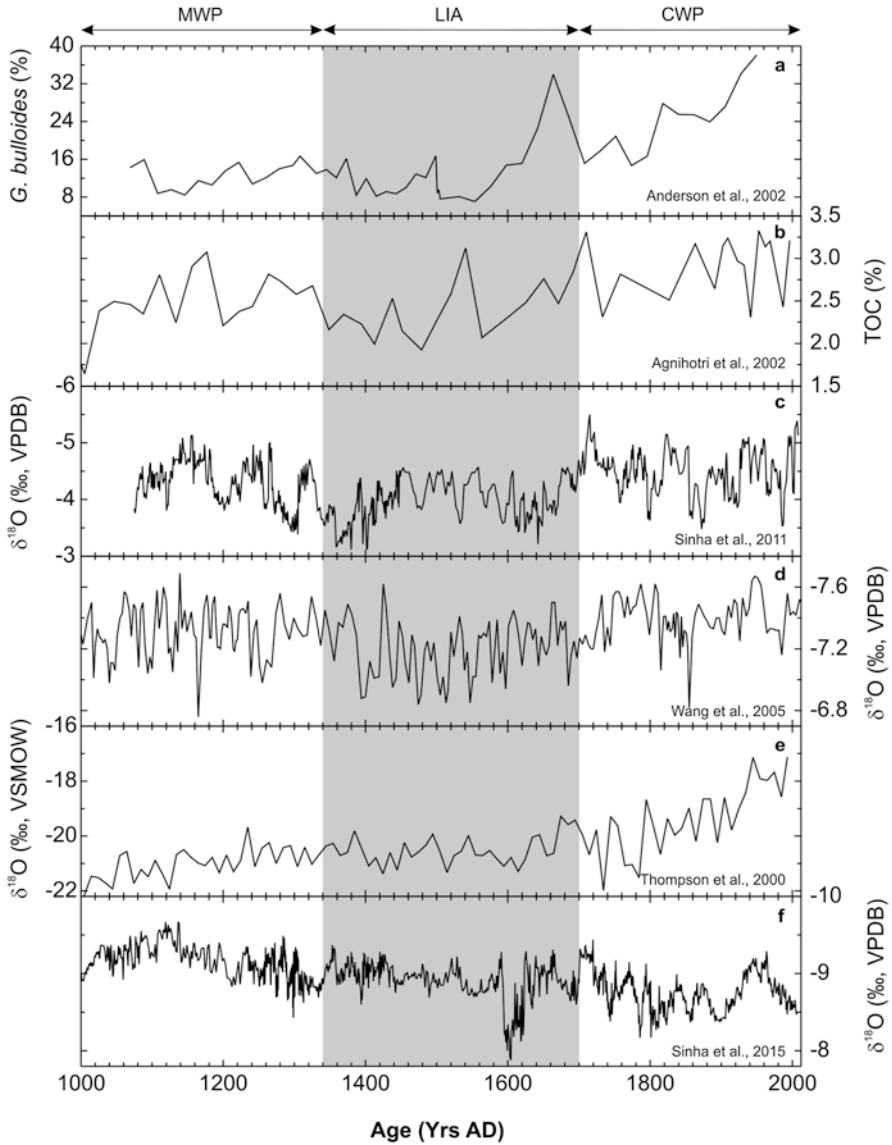
The asynchronicity in these records may be due to the different responses of two branches of the ISM (the Bay of Bengal branch and the Arabian Sea branch) and to changes in the moisture source and position of the ITCZ (Fleitmann et al. 2003; Prasad et al. 2014). The transition from middle to late Holocene is typified by another abrupt cold-arid phase at 4.2 kyr BP which was widespread throughout the globe and marked by reducing precipitation in monsoon regions, including the ISM in the Indian subcontinent (Berkelhammer et al. 2012; Dixit et al. 2014a; Leipe et al. 2014; Dutt et al. 2018; Kathayat et al. 2018). This weakening is evident throughout the region except in Central India (Prasad et al. 2014). The 4.2 kyr BP event had great bearing on the migration of Indus-Harappan population. The reduced monsoon strength decreased agricultural productivity of the Harappan settlements as this culture was more dependent on indentation agriculture (Berkelhammer et al. 2012; Dixit et al. 2014a, 2018; Dutt et al. 2018). The people of the Indus culture migrated to new areas to the east where there was more water availability for domestic and agricultural use, as reliable seasonal flooding of the river systems around the edge of the Thar Desert disappeared (Giosan et al. 2012).

During the Holocene, solar insolation and migration of the ITCZ, NAO and ENSO were major factors for controlling variability of the ISM (Fleitmann et al. 2003; Gupta et al. 2003; Dutt et al. 2015). Gupta et al. (2003) established a relationship between weak events in the ISM with the cold phases of the North Atlantic. The solar insolation and movement of the ITCZ had been the major factors in continuous weakening of the ISM since the early Holocene Maximum (Fleitmann et al. 2003; Gupta et al. 2003).

#### ***14.5.4 The Last Millennium***

The last millennium also witnessed several decadal to centennial scale changes in the ISM intensity as summarized in Dixit and Tandon (2016). Beginning about AD 950 and until AD 1250, the global climate was marked by warmer conditions of the MCA. The ISM experienced a strong monsoon phase and higher precipitation during the MCA (Wang et al. 2005a, b; Sinha et al. 2011, 2015) followed by a weak phase during the LIA cold period from AD ~1400 to 1700 (Anderson et al. 2002; Agnihotri et al. 2002; Sinha et al. 2011) (Fig. 14.9a–f). The interval from AD 1800 to Recent is known as the Current Warm Period (CWP) which witnessed a strong ISM.

This strengthening of the monsoon during the CWP is due to an increase in global temperature coinciding with the industrial revolution and other anthropogenic factors as well as the associated increase in surface evaporation and convection in the Indian Ocean (Anderson et al. 2002; Agnihotri et al. 2002; Wang et al. 2005a, b; Sinha et al. 2011). During the last 1300 years, solar insolation and El Niño intensity have been the major forcing factors influencing the ISM strength (Cook et al. 2010; Sinha et al. 2011). The increase in the ISM strength during the MCA and CWP and reduced rainfall during the LIA are all typically interpreted to be driven



**Fig. 14.9** Marine and continental proxy records of Indian summer monsoon variability during the last millennium. (a) RC2730, Arabian Sea (Anderson et al. 2002), (b) Site 2502G, Arabian Sea (Agnihotri et al. 2002), (c) Jhumar cave, Central India (Sinha et al. 2011), (d) Dongge cave, China (Wang et al. 2005a, b), (e) Dasuopu ice core (Thompson et al. 2000) and (f) Sahiya cave, northwest India (Sinha et al. 2015). *MWP* Medieval Warm Period (currently named as Medieval Climate Anomaly, MCA), *LIA* Little Ice Age, *CWP* Current Warm Period

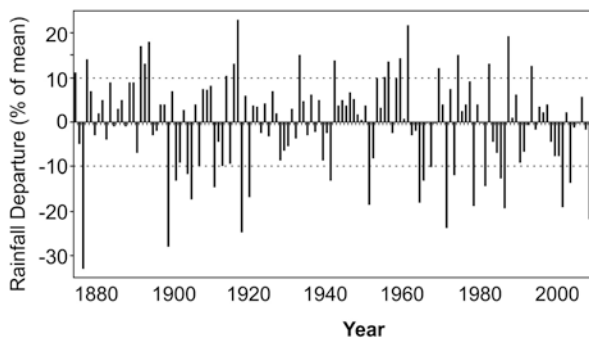
by changes in the incoming solar insolation (Fig. 14.9). ENSO events have been more frequent in the warm intervals, which are also reflected in more frequent low precipitation events during the warm phases. ENSO restricts the northward movement of the ITCZ that delays the onset of the summer monsoon causing a reduction in the rainfall (Sinha et al. 2011).

### 14.5.5 Instrumental Records

The ISM rainfall has also been highly variable during the last 130 years of meteorological observations. The average summer monsoon rainfall has shown affinity to an 11-year cycle of solar insolation (Gadgil 2003; Shewale and Kumar 2005). Several incidences of severe droughts and floods have been observed. The years 1877, 1899, 1918, 1972 and 2002 were years of weakest summer monsoon with 33%, 29%, 25%, 24% and 19% negative departure from the average summer monsoon rainfall (Fig. 14.10). Drought affected area increased between 1894 and 1908 and from 1964 to 1975 but decreased during 1908–1964 and 1985–1998 as a result of the changing rainfall conditions in India (Shewale and Kumar 2005). An increase in the intensity and frequency of sudden rainfall changes has been observed in India especially in the Himalayan region. Modeling studies indicate a future rise in the number of extreme rainfall events with a rise in global temperatures (IPCC 2014).

## 14.6 Summary and Future Aspects

Long-term changes in the Indian monsoon have been linked to the uplift of the Himalayan-Tibetan Plateau (H-TP). However, timing of its inception and intensification is still debatable. The H-TP uplift is associated with the regional climate, but



**Fig. 14.10** Percentage rainfall departure from the average all India summer monsoon rainfall between year 1875 and 2010. (Source: IITM, Pune)

consensus over timing of H-TP uplift and monsoon initiation/intensification has been elusive. Therefore, the palaeoclimate researchers and modelers still debate on the timing of the inception of the Indian monsoon and whether it really can be linked to surface uplift as suggested by climate modelers. Many proxies support monsoon initiation across the Oligocene-Miocene boundary (~22 Ma) or earlier. However, most recent studies fix the timing of beginning of the modern day Indian monsoon in the late Miocene.

During the Plio-Pleistocene, short-term changes in the Indian monsoon were found to be linked with several factors including the seasonal distribution of solar insolation, solar cycles, global ice volume, sea surface temperatures, albedo, H-TP heat budget, El Nino-Southern Oscillation and variability in the Indonesian Throughflow (Prell 1984; Kutzbach 1983; Quade et al. 1989; Clemens and Prell 1990; Raymo and Ruddiman 1992; Harrison et al. 1993; Kutzbach et al. 1993; Molnar et al. 1993; Li and Yanai 1996; Gordon et al. 1999; An et al. 2001; Gupta et al. 2003, 2004, 2013b). High-resolution benthic foraminiferal  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  records from the central equatorial Pacific captured significant imprints of periodic variations in the Earth's orbital parameters, particularly the obliquity (40 kyr) and the long eccentricity (400 kyr) cycles (Holbourn et al. 2013) during the Miocene. Monsoon studies from the Indian Ocean region lack high-resolution data for reconstructing monsoon strength and variability in response to the forcing factors. Very recently, new data from the NW Arabian Sea (Oman Margin and Owen Ridge) indicate the beginning of monsoon type wind system at about 12.9 Ma (Gupta et al. 2015), consistent with wind-related data from the Maldives (Betzler et al. 2016). Furthermore, studies based on materials from recent IODP expeditions in the Bay of Bengal and Arabian Sea may provide high-resolution data that can significantly contribute to the understanding of the evolution of the Indian monsoon system. Therefore, to resolve the timing of inception of the Indian monsoon, a coordinated effort is required to analyse and compare high-resolution records from marine cores located in high sedimentation areas of the Arabian Sea and the Bay of Bengal, as well as continental records with greatest degrees of continuity.

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