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HIGH RESOLUTION MONSOON RECORDS FROM LAND AND OCEAN: WHAT HAVE WE LEARNT DURING THE LAST DECADE?

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ABSTRACT *Impact of global climate change on the future water resources of India is intimately coupled with the variation of monsoon wind and monsoon rainfall. In this paper, the differences in paleo records from different areas have been discussed. It is felt that the climate models which can reproduce observed spatial distribution of monsoon are required to further understand the recorded paleomonsoon fluctuations.*

Key words Climate change; water resources; water cycle; paleomonsoon

INTRODUCTION

In one sentence one can define water as "water is life"; all organisms require water for their survival. Water is used for drinking, irrigation, transport, and also in industries. Old civilizations such as the Indus Valley civilization and the Mesopotamian civilization thrived near major water channels. All the populous cities are situated on the banks of large rivers (e.g., Varanasi and Patna on the banks of the Ganga). All the Indian rivers depend mainly on the monsoon rains for water, besides a small contribution from snow-melt in the case of perennial rivers. Drastic changes in the past monsoon might have lead to the extinction of civilizations. For example, some believe that the Indus valley civilization collapsed around 4200 years B.P. (Before Present, fixed at 1950 A.D.) due to the failure of the south Asian monsoon (Staubwasser et al., 2003).

More than ~70% of the area of the earth is under water. The earth is known as the 'blue planet', but only 2.55% of it is fresh water and the rest, saline. Most of the fresh water is locked in glaciers and deep underground. We draw water for our use from rivers, lakes and ground aquifers. The total amount of water present in world's rivers i.e. 2000 km^3 (Oki and Kanae, 2006) is much less than the annual water withdrawn by humans i.e. 3800 km^3. It is estimated that 7600 km^3 of evapotranspiration occurs from cropland and an additional 14,400 km^3 from permanent grazing land every year (Oki and Kanae, 2006). However unlike oil reserves, water reserves may not dwindle with geological time because of earth's closed water cycle. Water has its own cycle; it evaporates using the solar energy and the vapors condense to liquid during convection in the troposphere. Water is continuously supplied to rivers from either glacier melt or rainfall. Therefore the management of naturally flowing water with the help of dams is of prime concern. It is a very critical societal issue in modern times to sustain a fresh water supply and to increase the availability of water to match the demands of the ever growing population.

The International Hydrological Decade promoted studies on world water balances, and initial estimates were published in the 1970s (Lvovitch, 1973; Baumgartner and Reichel, 1975; Korzun, 1978). Recent advances in information technology have enabled the estimation of global water-balance at finer spatial resolutions (Oki et al., 2001; Vorosmarty et al., 2000; Alcamo et al., 2003). In the era of the 'Anthropocene'' (Paul and Crutzen, 2002; Crutzen, 2004), the era in which human impacts on natural processes are large and widespread, it no longer makes sense to study only the natural hydrological cycle: some studies have been initiated to assess the impact of human intervention on the hydrological cycle, thereby simulating more realistically the modern hydrological cycle on a global scale. In such studies, human withdrawals are subtracted from river flow (Alcamo et al., 2003), and the regulation of flow regime by major reservoirs is also incorporated (Hanasaki et al., 2006).

The demand for high-quality drinking water is limited to a few liters per person per day and can be met through international trade or by desalination. However, other demands for water for households, industry, and agriculture require up to one metric ton of water per day per person in developing countries and considerably more in developed countries. Problems of water, food, health, and poverty are interlinked in many developing countries, particularly in the regions where freshwater resources are scarce, the local economy is too weak to allow the import of food from outside on a large scale, and desalination plants are impractical to implement (Oki and Kanae, 2006).

According to the World Health Organization, 1 billion people do not have access to fresh clean water. A World Resources Institute analysis adds that 2.3 billion people (41% of the world population) live in water-stressed areas, a number expected to go up to 3.5 billion by 2025. Global population is rising by 80 million a year and the demand for new sources of fresh water also keep rising.

GLOBAL CHANGE AND THE WATER CYCLE

A number of factors affect the hydrological cycle and global climate change is one of them. Any change in temperature affects the atmospheric moisture, precipitation and circulation pattern of the atmosphere, e.g., changes in the rate of evaporation affects the hydrological cycle. Higher temperatures turn some part of snowfall into rainfall; the snowmelt season occurs earlier, consequently the timing and volume of spring flood changes substantially (IPCC, 2001). Nearly half of the world's population depends on groundwater sources for drinking and other domestic use; the ground water recharge depends on the seasonal distribution and rate of precipitation. Sea level rise causes salt water intrusion into freshwater aquifers near the coasts and decreases the available groundwater resources. Using IPCC reports on the future emission scenario of green house gases and multi-model ensemble techniques the uncertainties in the projections of future hydrological cycle (Milly et al., 2005) have been considerably reduced.

Any change in precipitation affects the fresh water reserves; changes in the intensity, amount, frequency and the type of precipitation have been observed. This behavior of precipitation makes us unable to sustain our water resources. The intensity of precipitation is influenced by increase in the heating due to green house effect, El Nino, etc. Increase in heat provides more surface moisture by enhanced evaporation. The heat used for evaporation acts to moisten the air rather than warm it. According to the Clausius Clapeyron relation, the water holding capacity of atmosphere increases by 7% with every 1^0C rise in temperature. Thus increase in temperature results in enhanced precipitation and risk of heavy rainfall. A warmer climate increases the risk of droughts and floods but at different places and at different times (e.g., in 2005, Mumbai, Chennai and Bangalore were flooded while the rest of India faced near-droughts). The amount of water vapor and cloud cover also control the amount of precipitation. Water vapor in the troposphere is the main source of moisture for monsoon, which plays an important role in our hydrological cycle. Monsoon occurs over India, China, SE Asian countries, Western Australia, Africa and to some extent in America. Most of these areas receive an intense summer precipitation and winter (dry) winds and this has a strong socio-economic impact in the region (Webster et al., 1998; Wang, 1994).

The Indian subcontinent experiences two different types of monsoons i.e. South West (or summer) monsoon (SWM) and North East (or winter) monsoon (NEM). SWM provides for ~80% of the total rain fall over the Indian subcontinent. SWM occurs between June-September (see Gadgil, 2003 for a review). The Indian monsoon is influenced by El Nino, La Nina, Indian Ocean Dipole (IOD) and the Walker circulation in the equatorial Pacific (Kumar et al., 1999; Krishnamurthy and Goswami 2000; Sarkar et al., 2004). Change in Walker circulation and the Southern Oscillation Index (SOI) i.e., difference in air pressure at sea surface between the eastern and western Pacific oceans are related: positive SOI leads to the La Nina and negative SIO leads the El Nino. Ashok et al. (2001) observed that sea surface temperatures has an important effect (i.e., a see-saw pattern between the central equatorial Indian Ocean and Indonesian water, known as IOD) on the intensity of the Indian monsoon. Climate change is expected to accelerate the global hydrological cycle, and the average precipitation is expected to increase (Pant and Rupakumar, 1997; Rupakumar et al., 2002). Evapotranspiration may not increase as much as precipitation globally because elevated CO_2 concentration induces stomata closure and is likely to reduce transpiration (Gedney et al., 2006), and river discharge is likely to increase on global scale because of the increased precipitation and the reduced transpiration (Milly et al., 2005).

India is a tropical country gifted with a large number of the river channels and the big rivers are perennial and they continuously receive water from glacier melt and monsoon runoff. But due to recent rapid industrialization and consequent urbanization some of our fresh water resources have become contaminated by chemical effluents and the sewage from urban localities. Superimposed on this is the threat of Global warming and its consequences. Therefore it is very important for us to understand the pattern of monsoonal rainfall and the effect of changing climate. A concerted effort between geologists, climatologists, engineers, chemists and toxicologist is needed to manage our dwindling water resources.

The changes observed so far, which may be linked to global warming are as under. It has been recorded that a large scale precipitation change (which includes rainfall, snow fall and other forms such as frozen or liquid water falls from clouds) has occurred during the last century. Around 2.17 mm per decade increase has been observed over land. The urban areas experienced higher precipitation mostly during summer than the nearby rural areas due to change in the gradient of human energy production (20-70 w/m^2). This destabilizes the environment, thermal perturbation of the boundary layer results in the downstream translation of Urban Heat Island or UHI generated clouds (Shepherd et al., 2002; Shepherd and Burian, 2003; Shepherd, 2005). The excess rainfall in the Indian metros recorded in 2005 could be due to such a phenomenon.

All the water resources of the world depend on precipitation. For the last century the precipitation change (% per century or decade) was calculated spatially by using Global Historical Climatology Network (GHCN) station data. It was observed that the rainfall over higher latitudes increased during the last century whereas some of the tropical provinces such as South America, Chile and Mexico were affected by droughts. At the same time Sahel in Africa faced the largest negative trend in precipitation.

PROXY RECORDS OF PALEOMONSOON

Prior knowledge about extremes of climate such as droughts, floods and precipitations is very important for the society; therefore it is crucial to understand the variability of climate on larger spatial and longer time scales (Ramesh, 2001; Ramesh and Yadava, 2005). Paleomonsoon reconstruction using stable isotopes of oxygen from marine sediments and speleothems is useful in addressing this issue.

It is known from previous studies that the South Asian monsoon exhibits variance at different time scales viz. decadal, centennial, and millennial. Decadal scale variations can be studied using recorded meteorological data, which is available for the last century or so, but limited to the four metros where weather stations are located (Parthasarathy et al., 1995). For longer time scales we must take recourse to various paleoclimatic proxies, such as sediments deposited in world oceans (e.g. Sinha, 2007). Recently with the advent of AMS (Accelerator Mass Spectrometry) one can obtain highly accurate chronologies because, here, instead of dating bulk sediments, planktonic foraminifera are dated (no contamination from detrital carbonate material). If suitable cores from appropriate regions (such as continental margins) are available then we can explore paleomonsoon variations on centennial to decadal time scales (comparable to human lifetime). The AMS method also offers a higher sample throughput; thus more layers of the sediment can be dated, providing a high time resolution up to about 40,000 years back in time (e.g., Tiwari et al., 2005).

Three distinct regimes exist as far as paleomonsoon reconstruction is concerned: The western Arabian Sea, off the Somalian coast experiences intense upwelling during southwest monsoon resulting in increased organic/inorganic productivity and negligible fresh water run off due to meager precipitation over adjoining landmass. As the western Arabian Sea is well known for the upwelling induced by the monsoon winds, leading to a large reduction of Sea Surface Temperature (SST) by at least 4°C, the monsoon signal is more easily detectable here. A large number of previous studies (e.g., Naidu, 2006; Staubwasser, 2006;

Tiwari et al., 2006; Naidu and Malmgren, 1996, 2005) concentrated here, reconstructing the monsoon winds over the past several millennia, using the variation in the abundance of a particular surface-dwelling (planktonic) foraminifera, called *G. bulloides*. This species is cold loving and has a natural preference to temperate rather than tropical waters. However as SSTs can be quite low in the upwelling regions of the northwest Arabian Sea (off Gulf of Aden and Somalia), this species proliferates here and has become the marker for pale monsoon reconstruction. It is now clearly established that the monsoon was weak during the Last Glacial Maximum (21ka). The Holocene (i.e., the past 10 ka) records indicate a decade to century scale variability superimposed on the orbit ally driven change in the monsoon: due to changes in the sun-earth geometry, i.e., eccentricity (periodicity - 100,000 years), tilt or obliquity (41,000 years) and the precession of equinoxes (19 to 23 thousand years), the earth goes through warm (interglacial) and cold (glacial) periods. Warmer climates favour better monsoons, while they weaken during colder periods in general - see Korisettar and Ramesh (2002) for a review. Another important finding is that the Asian monsoon has recently reversed its millennia –long orbit ally driven low frequency trend towards less rainfall in the drier areas of its influence, coinciding with the synchronous increase in the inferred monsoon winds over the Western Arabian Sea (Anderson et al., 2002); this leads to enhanced upwelling and hence increase in productivity (Gupta et al., 2003). This change could be related to an increased summer heating over and around the Tibetan Plateau (Braining and Man twill, 2004; Morrill et al., 2005) or with persistent interglacial millennial scale mode of monsoon variability (Gupta et al. 2003). (ii) The eastern Arabian Sea off the Western Indian coast experiences moderate upwelling along the coastal regions of western India and copious fresh water runoff due to intense precipitation (1000-4000 mm/yr) on adjoining land (between Mumbai and Cochin, Sarkar et al., 2000). Results from this region (Sarkar et al., 2000; Thamban et al., 2001; Tiwari et al., 2006) show that there was no decline in the monsoon run-off during the Holocene, in contrast to observations from the western Arabian Sea. (iii) the northern Bay of Bengal, which receives an enormous amount of freshwater discharge due to the monsoon rains on the hinterland; the significant decrease in the surface salinity and the stable oxygen isotopic composition ($\delta^{18}O$) of the surface water, well preserved in the $\delta^{18}O$ of the $CaCO_3$ shells on plank tonic foraminifer such as *G. sacculifer and G. ruber*, is the monsoon signal to look for. Three detailed records available for this region are not concordant for the late Holocene (Chauhan et al., 1993; Weber et al., 1997; Kudrass et al., 2001).

As better dated high-resolution pale monsoon records become available, important questions such as the following could be addressed: How good is the correlation between the wind and rain records from the western and the eastern Indian Ocean? Tiwari et al. (2006) reported precipitation proxies, i.e., high-resolution stable oxygen isotope variations of two different species of plank tonic Foraminifer viz. *Gs. sacculifer* and *Gs. rubber* in an AMS ^{14}C dated sediment core from the monsoon-runoff-dominated eastern Arabian Sea. Their data sets reveal that during the past ~1800 years (and perhaps up to ~2800 years, resolution of wind record was too low for comparison) reductions in monsoon wind strength in the western Arabian Sea appear to be persistently accompanied by aridity over India. It clarifies that past fluctuations in SWM precipitation over the Indian subcontinent followed the wind intensity records from the western Arabian Sea on centennial time scales. Figure 1 compares the $\delta^{18}O$ record of *G. ruber* from an equatorial Indian Ocean sediment core (data from Tiwari et al., 2005) and the salinity reconstructed from *G. ruber* $\delta^{18}O$ in a sediment core from the Bay of Bengal (data from Kudrass et al., 2001). Both show similar long term trends, while the fluctuations in the equatorial Indian Ocean appear to be of somewhat higher magnitude. This is because we know that ~1‰ change in salinity changes $\delta^{18}O$ only by 0.3‰.

Until this question is fully settled, we must take recourse to continental and ocean records of precipitation rather than wind to learn about the paleo-hydrology of India. The variation of $\delta^{18}O$ in the ice accumulating in Himalayan/Tibetan glaciers, which are free from the problems of melting and refreezing that affect preservation of original isotopic signatures with fidelity, has been used as a qualitative pale monsoon indicator (Thompson et al., 1997, 2000). The records from land, such as sand dunes suffer from imperfect chronology, with large uncertainties. It appears that widespread deposition of sediments by rivers and dune building activity was associated with cool and dry phases, while large scale fluvial erosion, incision and sediment transport in the rivers and extensive slope failures were dominant during the warm, wet, monsoonal phases (see Kale, 2004). In the background of such uncertain qualitative inferences, the first quantitative reconstruction of monsoon rainfall in India with decadal to annual resolution has been obtained from speleothems, including samples from the Gupteswar cave, Orissa (Yadava et al., 2004; Yadava and Ramesh, 2005, 2006). This has been possible by using what is known as the 'amount effect' in the precipitation.

In the deeper parts of a cave in karstic regions, where air circulation is poor and high humidity prevails, carbonate precipitates slowly, maintaining isotopic equilibrium between different ionic species. In such a case isotope ratios of oxygen ($^{18}O/^{16}O$) of the ions in the dripping water, which are influenced by the ambient environment of the cave, are preserved in the growing speleothem lamina and can be used to reconstruct the past environment (e.g. Gascoyne, 1992; Lauritzen, 1995; McDermott, 2004).

For mid latitude and semiarid climatic zones $\delta^{18}O$ of precipitation decreases with increasing rain amount (Dansgaard, 1964; Bar-Matthews and Ayalon, 1997; Fricke and O'Neil, 1999) and the temperature dependence is very weak. In tropical locations any obvious temperature correlation is not observed for the modern rainfall. The $\delta^{18}O$ of rain is dependent on the amount of rainfall (Yurtesever and Gat, 1981): more rainfall is associated with less of ^{18}O content in the precipitation; this is termed as the 'amount effect' (Dansgaard, 1964). Hence, in tropical caves the $\delta^{18}O$ of freshly deposited calcite layers on a growing speleothem is a good proxy for the past variations in $\delta^{18}O$ of meteoric water, and thus the amount of rain.

For island stations a linear relationship (Yurtsever and Gat, 1981) between the mean monthly $\delta^{18}O_m$ of precipitation and the mean monthly rainfall was observed:

$$\delta^{18}O_m = (-0.015 \pm 0.002) \times P_m - (0.47 \pm 0.42) \qquad (1)$$

P_m is the mean monthly rainfall, with correlation coefficient r = 0.87 for 14 island stations (each has at least 40 monthly observations). Average rate of depletion is found to be -1.5 ± 0.2‰ for a 100mm increase in the monthly rainfall. This depletion rate should be applicable to those locations where annual temperature fluctuations remain within a narrow range. During the monsoon season in 1999, precipitation samples collected at Jharsuguda (22°N, 84°E), which receives majority of the annual rainfall during southwest monsoon, monthly depletion rate for 100 mm increase in the monsoon rainfall was found to be –2.2 ± 0.8‰ (Yadava and Ramesh, 2005), based on daily samples collected during three successive months: July, August and September. This agrees well with the depletion rate observed at the island stations. It suggests that during the monsoon months when the vast continental land area cools down and attains a moderate temperature till the monsoon is active, the amount effect at the inland sites is the same as what is observed at the island stations. The $\delta^{18}O$ of speleothem from Gupteswar, Orissa, could thus be converted into amount of rainfall, assuming that changes in the speleothem $\delta^{18}O$ had been solely due to variations in the $\delta^{18}O$ of the annual rain, and that the depletion rate experienced at the island stations is also applicable at the cave site.

The Gupteswar stalactite $\delta^{18}O$ record (Fig. 1(c)) is compared in Fig.1 with two other, similar, high resolution (comparable with that of speleothems; i.e. ~1yr to ~15 yr) paleomonsoon records from southern Asia, spanning the last 3400 yrs: (i) the stable oxygen isotope variations from stalagmites in southern Oman (Fig. 1(a); data from Fleitmann et al., 2003); (ii) a high resolution (~7yr) record of von Rod etal.,(1999), varve thickness of sediments collected from off Karachi, Pakistan (northeastern Arabian Sea): here, precipitation and hence the river runoff control varve thickness. The precipitation at the sampling site (von Rod et al., 1999) occurs both during the summer (Jun.-Sep) and winter monsoons (Nov to Mar.) forming an annual couplet of alternate dark and light colored sediment sequences. The precipitation may have fluctuated (Lückge et al., 2001) due to variations in the extreme positions of the ITCZ (inter-tropical convergence zone) and hence, the variability in varve thickness was interpreted as a proxy for past rainfall variations (Fig. 1(b)). The Gupteswar and Oman speleothem records seem to agree well. The monsoon was stronger around 3000 B.P. as indicated by more depleted $\delta^{18}O$ values and also by the increased growth rate (higher sampling density). The increasing trend between 1200 yr B.P. to 400 yr B.P. is seen in both the records. Also during the extremely low rainfall epochs of 1700 and 2000 yr B.P. shown by the Gupteswar speleothem, the Oman stalagmites show a hiatus, probably due to the complete lack of rain. It must be noted that Oman is more like a desert relative to eastern India and therefore growth of stalagmites there is more sensitive to rainfall fluctuations. The decreasing trend of rainfall from 3400 to 1900 yr B.P. is reflected both in the varve and Gupteswar records (Lückge et al., 2001, have reported lowest Ti/Al ratio that reflects terrigenous matter and hence runoff) in the same core around 2000 yr. B.P. But the two records differ significantly during the last 1500 years. Pollen records from different lakes in Rajasthan (Singh et al., 1974; Bryson and Swain 1981; Swain et al., 1983) show high monsoon rainfall around 600 yr B.P., similar to the Gupteswar data. As the speleothem records separated by a larger distance agree very well, it is likely that the varve thickness response to the monsoon is nonlinear.

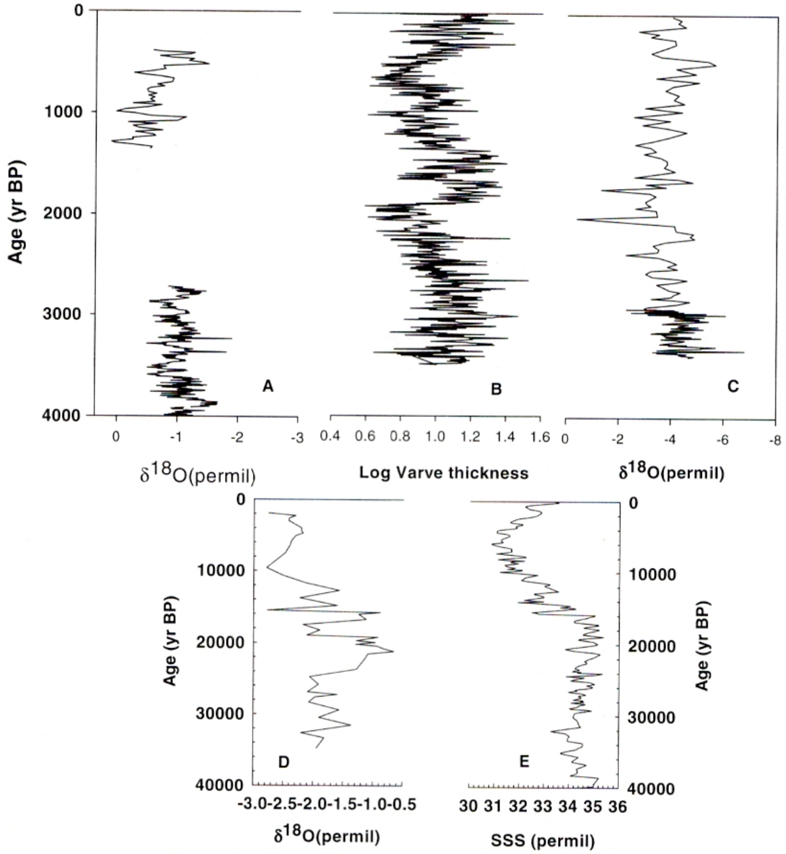

Fig. 1 Comparison of high resolution paleomonsoon records for the last four millennia from (a) Oman stalagmites (data from Fleitmann et al., 2003) (b) varved sediments (northern Arabian Sea, data from von Rad et al., 1999) (c) Gupteswar stalactite (data from Yadava and Ramesh, 2005); and the last forty millennia from (d) *G. ruber* $\delta^{18}O$ from the eastern Arabian Sea (data from Tiwari et al., 2006) and (e) salinity data derived from *G. ruber* $\delta^{18}O$, from Bay of Bengal (data from Kudrass et al., 2001).

Staubwasser (2006) compiled the available paleo monsoon records from the different localities from the Arabian sea (see Table 1; ODP site 723A,Gupta et al., 2003; southern coastal Oman, Q5, Fleitmann et al., 2003, site 56KA, from NE Arabian sea, von Rod et al., 1999 and 63KA, Staubwasser et al., 2002, 2003, and 3268G5 from the eastern Arabian Sea, Sarkar et al., 2000) and from Bay of Bengal (BOB, Site 126KL, Kudrass et al., 2001). The above studies involved different types of proxies e.g. abundance of planktic foraminifera (Gupta et al., 2003), speleothem stable oxygen isotope ratios, varve thickness (von Rod et al., 1999) and the stable oxygen isotope record of water mass sensitive planktic foraminifera (Kudrass et al., 2001; Sarkar et al., 2000). Staubwasser (2006) observed that the

Table 1 Summary of high resolution monsoon records from the south Asian region.

Number		Locality/ Sites	Latitude	Longitude	Proxy Used	Indicator of climatic condition	Duration	Dating method and time resolution
1		63KA	24^{0}50'N	65^{0}55'E	*Gs.ruber* $\delta^{18}O$ records	Depletion in $\delta^{18}O$ shows strong SWM and high runoff	12,000 yrs B.P.	AMS ^{14}C 16.72 yrs
2		56KA	24^{0}50'N	65^{0}55'E	Varve thickness	Higher thickness suggests higher precipitation and increase runoff from the Indus river	5,000 yrs B.P.	AMS ^{14}C 7 yrs
3		3268 G	12.5^{0}N	73^0 E	$\delta^{18}O$ records of *Gs. sacculifer* & *Gr.menardii*	Depletion in $\delta^{18}O$ records shows strong SWM and high runoff from the Western Ghats	10,000 yrs B.P.	AMS ^{14}C 252.48 yrs
4		GC 5	10°23′N	75°34′E	$\delta^{18}O$ records of *Gs. ruber* & *Gs. sacculifer*	Increase in $\delta^{18}O$ records of *Gs. ruber* & *Gs. sacculifer* suggest low runoff from western Ghats and high salinity	18,000 yrs. B.P	AMS ^{14}C 473.68 yrs
5	Arabian Sea	ODP-723	18^{0}58.40' N	58^0 E	Abundance of *Gg.bulloides*	Higher abundance suggests strong SWM and upwelling	11,500 yrs B.P.	AMS ^{14}C 137.66 yrs
6		Q5	17^{0}10'N	54^{0}18E	$\delta^{18}O$ records of stalagmite	High $\delta^{18}O$ value suggests low rainfall i.e. weak SWM	10,000 yrs B.P.(With the hiatus in the growth between 2600-1400 yrs B.P.	*Th-U TIMS and MC-ICPMS
7		126KL	19^{0}58.40' N	90^{0}02.03' E	$\delta^{18}O$ records of *Gs.ruber*	Depletion in $\delta^{18}O$ records shows the strong SWM, decrease in salinity and high runoff from the Ganga Brahmaputra Discharge	80,000 yrs B.P.	AMS ^{14}C 100 yrs
8	**Bay Of Bengal**	31/11	15^{0}52.00' N	91^{0}10.00' E	$\delta^{18}O$ records of *Gs.ruber* & *Gs. sacculifer*	Higher value of $\delta^{18}O$ records shows the weak SWM	20,000 yrs B.P.	^{14}C (Bulk Sample) 371.70 yrs

Source: (1) Staubwasser et al., 2002, 2003; (2) Von rod et al., 1999; (3) Sarkar et al., 2004; (4) Thamban et al., 2001; (5) Gupta et al., 2003; (6) Fleitmann et al., 2003; (7) Kudrass et al., 2001; (8) Chauhan et al., 1993.

* TIMS-Thermal Ionization Mass Spectrometry and MC-ICPMS (Multi-Collector Inductively Coupled Plasma Mass Spectrometry).

Holocene paleo-archvies from these regions revealed significant spatial differences in paleomonsoon performance. The Oman margin (Q5 and ODP 723A) record suggested a strong South West Monsoon (SWM) during early to mid Holocene on the basis of very high abundances of upwelling indicator planktic foraminiferal species *Gg. bulloides,* a reduction in $\delta^{18}O$ of speleothems and lower $\delta^{18}O$ of *Gs.ruber* from NE Arabian Sea. Kudrass et al. 2001, (Site 126KL) also reported maximum river water discharge from the Ganga- Brahmaputra (G-B) to the BOB during the early to mid Holocene, these records also correlated with the enhanced humidity recorded in southern Tibet (Gasse and van Campo, 1994) . On the contrary, eastern Arabian Sea records show higher runoff from the western Ghats during the Late Holocene (Sarkar et al., 2000; Thamban et al., 2001) inferred on the basis of reduction in foraminiferal $\delta^{18}O$ due to the lowering in the salinity of sea water caused by runoff. Thus the Eastern Arabian Sea records show an inverse relationship with the western Arabian sea records. Stabwasser (2006) conjectured that the observed differences could be due to changes in the winter monsoon rains, affecting the run-off proxies. He proposed a model to explain the variability in the monsoon over time, which suggests that the dissimilarities in between the paleo records might be due to the difference in the distribution of active and break monsoon periods. Due to break in the monsoon, some of the monsoon rainfall could have shifted from the Indian peninsula to the Himalayan foothills, as deciphered from the higher river discharge in BOB during the Early Holocene.

Kale (2004) has summarized that in the early Holocene (9.5 to 5.5 ka) the southwest monsoon was stronger than today and in the late Holocene, around 3.5 ka, it weakened. Anderson et al. (2002) showed that during the past 4 centuries, it has improved again. The annual resolution record of monsoon from a speleothem in Karnataka (Yadava et al., 2004) has not shown any long term increasing trend. However, recent work (Yadava and Ramesh, 2007) has clearly shown that there is a solar periodicity of ~22 years present during a better part of the record. Based on the instrumental data of monsoon rain for the last century, it appears that the monsoon was never lower than 20% of the normal. This appears to be true for the ~330 years record as well. But when we look at the past ~3400 years, there were extended periods when the monsoon had really failed. The sun-monsoon link is also confirmed by the speleothem record from the Himalaya (Denniston et al., 2000, Sinha et al., 2005).

CONCLUSIONS

Predicting the future water resources of India in the context of global change is intimately coupled with the variation of monsoon wind and monsoon rainfall. The differences in the paleorecords from the different areas have been discussed. In general a warmer climate appears to strengthen the monsoon. While there were some years of deficient rain the paleo-records pertaining to the last few centuries, and no failures were observed, this is not true of the paleo-record of the last three millennia, which shows decadal failures of the monsoon. This appears to have some link with solar irradiance variations, though the connecting physical mechanism is not clear. On longer time scales, warmer periods have in general witnessed higher monsoon rain than the colder episodes. This appears to be mainly controlled by the precession of the earth. Good climate models capable of reproducing the observed spatial distribution of modern monsoon are required to further understand the recorded paleomonsoon fluctuations.

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