Climate and water resources of India

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We attempt to synthesize available quantitative, precisely dated and high-resolution palaeorecords of the South Asian summer monsoon from different natural archives, highlighting their similarities and differences. We distinguish between the palaeorecords of monsoon winds and monsoon rainfall and underscore the importance of quantitative rainfall reconstruction using the amount effect in monsoon rainfall, which has been demonstrated based on actual measurements. Predicting the future of water resources of India in the context of Global Change, intimately coupled with the variations of monsoon, depends on how well we understand the palaeomonsoon.

Keywords: Climate, global change, palaeoclimate, water resources.

FOR a country like India, which is still largely dependent upon rain-fed agriculture, availability of freshwater is one of the foremost concerns for the future. Most of Indian plains receive about 80% of their annual quota of rain from the southwest monsoon during the four months, June to September[1]. The coastal areas in peninsular India receive rain from the northeast monsoon during October to December, which includes cyclonic storms. A small addition is made by the western disturbances to the northern parts of India in winter. In the context of anticipated global warming[2] due to increasing atmospheric greenhouse gases, it is necessary to evaluate the possible impact on freshwater resources of the country. A pre-requisite for undertaking this is good high-resolution (both in space and time) palaeomonsoon data from different geographical regions influenced by the monsoon. In this article we discuss what we know from well-dated, high-resolution palaeomonsoon records about past changes in the Indian monsoon rainfall, the main source of freshwater for the country, and how reliable are such palaeorecords.

Global change

Over the past decade, evidence is mounting for global warming and accompanying changes in the earth[2]. Over the twentieth century, it is estimated that the average surface temperature of the earth has increased by 0.6°C[3], and has led to possible decrease in the snow cover and ice extent, and an increase in the sea level. It is estimated that precipitation has increased by 0.5–1% per decade during the last century, over most mid- and high latitudes of the Northern Hemisphere continents; rainfall is likely to have increased by 0.2–0.3% over the tropical (10°S to 10°N) land areas. In the period 1900 to 1995 AD, it was observed that in parts of Asia and Africa, the frequency and intensity of droughts increased[2]. As greenhouse gases continue to build up in the atmosphere, global average water vapour concentration in the atmosphere and precipitation are likely to increase during the current century, as predicted by a wide range of climate models, based on different future emission scenarios. It is also predicted that by the mid-twenty first century, the northern parts of the mid- and high-latitudes and Antarctica would receive more winter precipitation, while at low latitudes, there are both regional increases and decreases in rain. With increasing precipitation, the inter-annual variability is also expected to increase[2].

Potential impact of global change on water resources include enhanced evaporation due to warming, geographical changes in precipitation intensity, duration and frequency, together affecting the average runoff, soil moisture, and the frequency and severity of droughts and floods. Future projections using climate models point to an increase in the monsoon rainfall in most parts of India with increasing greenhouse gases and sulphate aerosols[4].

In a preliminary study[5] of the effect of global warming on the water resources of India, a distributed hydrological model was used on 12 major Indian river basins, with simulated daily weather data from Hadley Centre Regional Model-2 to study the present water availability in space and time without incorporating man-made changes such as dams, canals, etc. (a 20-year control run). Assuming no change in future land use pattern, the same model was used to predict future (20 years) changes that might follow global warming. It was seen that the severity of extreme events such as droughts and floods in different parts of the country increased. Also the available runoff decreased significantly. One major problem with such models is lack of proper calibration, precluded by *scarcity of reliable river discharge and soil moisture data*. To quote Kulkarni[6], 'in the absence of actual hydrological data such as observations of river flow data at a number of points along a river and its tributaries, the long period basic meteorological data like rainfall, temperature, humidity, etc., can be used for estimating water potential of a region or a basin and its variation in space and time by using suitable technique'.

Many north Indian rivers such as the Ganga, Yamuna have shown a sharp decline in the summer discharge in

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the recent past, possibly due to shrinking of the Himalayan glaciers that feed them. During the past 40 years an estimated reduction of 40% in the Himalayan ice volume has occurred[5]. These observations lead us to an important question: is this a consequence of global warming, or is it only a part of the low frequency climate variability inherent to the system? To answer this would require reliably dated, high-resolution proxy records of palaeo-precipitation.

Instrumental records of monsoon

Monsoon is the most important climatic phenomenon occurring in the Indian subcontinent and the adjoining regions. It forms the backbone of Indian economy; a high degree of correlation exists between monsoon rainfall and agricultural production[7]. Droughts and floods cause havoc in terms of economic backlash. It is therefore important to understand the inherent natural variability in monsoon and the factors controlling it, before attempting long-term prediction. 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 have been functioning for a long time. For longer time scales we must take recourse to various palaeoclimatic proxies, such as sediments deposited in the Indian Ocean, ice deposited in Tibetan/Himalayan glaciers, speleothems, etc.

The all-India summer monsoon time series[6], obtained by the arithmetic average of the district area weighted rainfall of 306 stations spread over Indian plains, has a mean value of 852.6 mm, with a coefficient of variation of 9.9%. Years of excess rainfall during 1871–2000 were 20, out of which 17 were La Niña years: 1874, 1878, 1884, 1892–4, 1910,1916–7, 1933, 1942, 1947, 1956, 1959, 1961, 1970, 1975, 1983, 1988 and 1994; and deficit rainfall occurred in 22 years, of which 12 were associated with El Niño: 1873, 1877, 1899, 1901, 1904–5, 1911, 1918, 1920, 1928, 1941, 1951, 1965–6, 1968, 1972, 1974, 1979, 1982, 1985–7. Though there is no secular trend in the all-India rainfall, regional data show increasing (west coast, north Andhra Pradesh and northwest India) and decreasing (east Madhya Pradesh, northeast India and parts of Gujarat) trends[4].

The frequency of monsoonal cyclonic disturbances in the northern Indian Ocean has shown a significant decreasing trend during the twentieth century, with about 50% reduction during the whole period. The frequency of monsoon cyclones has also registered a decreasing trend, more pronounced in the recent decades. A decreasing trend in the drought-affected areas has also been reported[4].

Proxy records of monsoon

Most of the previous studies (e.g., ref. 8) on sea sediments concentrated on millennial scale climate changes using relative dating methods or radiocarbon dates on bulk sedimentary matter that might prove to be relatively inaccurate. In the last decade, with the advent of AMS (Accelerator Mass Spectrometry), highly accurate chronologies have become available because, 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, where sedimentation rates are high) are available then we can quantify palaeomonsoon 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 reliability up to about 40,000 years back in time (the limit of radiocarbon dating method).

Several compilations of proxy climate records from south Asia, including palaeomonsoon, are available[4,9–14]. The northern Indian Ocean is well suited for monsoonal studies as it experiences intense biogeochemical changes associated with monsoons[14,15]. It can be divided into three distinct regimes as far as palaeomonsoon reconstruction is concerned: (i) The western Arabian Sea, off the Somalian coast: it experiences intense upwelling during south-west monsoon resulting in increased organic/inorganic productivity[14] and negligible freshwater runoff due to meagre 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 4ºC, the monsoon signal is more easily detectable here. A large number of previous studies[15–18] concentrated here, reconstructing the monsoon winds over the past several millennia, using the variation in the abundances of either lithic grains transported by wind from the adjoining desert or 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 the Gulf of Aden and Somalia), this species proliferates here and has become the marker for palaeomonsoon reconstruction. (ii) The eastern Arabian Sea off the Western Indian coast: it experiences moderate upwelling along the coastal regions of western India and copious fresh water runoff due to intense precipitation (1000–4000 mm/yr) on the adjoining land (between Mumbai and Cochin)[19,20]. (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 of planktonic foraminifera such as *G. sacculifer* and *G. ruber*, is the monsoon signal[21] to look for in cases (ii) and (iii) above.

Other long records of palaeomonsoon are derived from unique deposits such as varved sediments (annual laminations, thickness of which is assumed to be directly proportional to the monsoon strength) from the northern Arabian Sea (off Karachi)[22], speleothems from the Arabian (stronger the monsoon, more to the north is the movement of the Inter Tropical Convergence Zone) and Indian (more monsoon rain results in less $^{18}O/^{16}O$ in calcite) peninsulas[23–28], ice cores from the Tibet and Himalaya[29–31] (stable oxygen isotopic composition of ice and trapped dust concentration inversely proportional to the monsoon strength) and lake sediments from Rajasthan[32] (carbon isotopic composition of lake organic matter related to lake level and hence to monsoon strength). In what follows, we avoid discussion of some short span records (a few centuries) such as tree rings and corals[33] (transfer functions to monsoon not clear), diagenetic alteration-prone continental sedimentary records of dubious chronology, and bulk ^{14}C dated records, except for the purpose of comparison.

Long-term palaeorecords of the monsoon indicate that in general, the south Asian monsoon is stronger during warm climate and weaker during cooler climate, while the winter monsoon behaves opposite[8,13]. The monsoon intensity appears to be basically controlled by the sun–earth geometry and consequent changes in the incoming solar radiation (insolation). Tropical latitudes are more affected by changes in the precession of the earth's perihelion, while high latitudes are affected more by the change in the tilt of the earth's axis of rotation. This has been verified using palaeoclimate models[34–36]. A crude understanding is that ice on the Tibetan plateau during the last glacial maximum decreased the land–ocean temperature contrast in summer and increased it during winter. In addition, the monsoons are influenced by high latitude changes (such as North Atlantic Deep Water formation, ice rafting from the Greenland Ice Sheet, etc.), and surface feedbacks such as vegetation (through changes in albedo, surface friction and evapotranspiration). More recent and precisely dated records show that during the last glacial maximum, the southwest (i.e. summer) monsoon was weaker while the northeast (i.e. winter) monsoon strengthened[13]. From about 10,000 years (beginning of the Holocene), the monsoon intensified as shown by stalagmites from Oman[23] and *G. bulloides* abundance[17] in the western Arabian Sea. As the summer insolation at the top of the atmosphere started decreasing steadily (by ~8% during the last 10,000 years), the monsoon *wind*s also appear to have steadily weakened[17,23]. However, *rainfall* indicators show that from 10,000 years to 2000 years ago there was a steady enhancement of the monsoon rainfall[19,20]. Figure 1 compares these two records. Thus as more well-dated high-resolution palaeomonsoon records become available from the Indian Ocean, 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 on different time scales? Does increasing wind intensity always favour increased rainfall over land? This is especially important because, while winds are necessary to advect the marine moisture over land, often it is observed that excess winds cause farther advection, away from areas badly needing precipitation. Until this question is settled, we must take recourse to continental and ocean records of precipitation rather than wind to learn about the palaeohydrology of India.

The first *quantitative* reconstruction of monsoon rainfall with *decadal to annual resolution* has been obtained from Indian speleothems[24–28]. This has been possible by using what is known as the *amount effect* in the precipitation.

In the deeper parts of a cave, where air circulation is poor and high humidity prevails, carbonate precipitates slowly, maintaining isotopic equilibrium between different ionic species. In such a case, the stable isotope ratio of oxygen ($^{18}O/^{16}O$) of the ions in the dripping water, which is influenced by the ambient environment of the cave, is preserved in the growing speleothem lamina and can be used to reconstruct the past environment[37–39].

$\delta^{18}O_c$ (oxygen isotopic composition of speleothem) is related to temperature, due to (a) temperature-dependent fractionation[40] ($\delta^{18}O_c/dT \approx -0.21‰/°C$ at 25°C) and (b) the dependence of $\delta^{18}O_w$ (oxygen isotope composition of precipitation) on the condensation temperature which is 0.5 to 0.9‰/°C between 40°N and 60°N[41]. For mid latitude and semiarid climatic zones $\delta^{18}O_w$ decreases with increasing rain amount[42–45] and the temperature dependence is very weak. In tropical locations any obvious temperature correlation is not observed for the modern rainfall[42,44–46]. The $\delta^{18}O_w$ is rather dependent on the amount of rainfall[46]: more rainfall is associated with less of $\delta^{18}O$ content in the precipitation; this is termed as[42] the 'amount effect'. Hence, in tropical caves the $\delta^{18}O$ of freshly deposited calcite layers

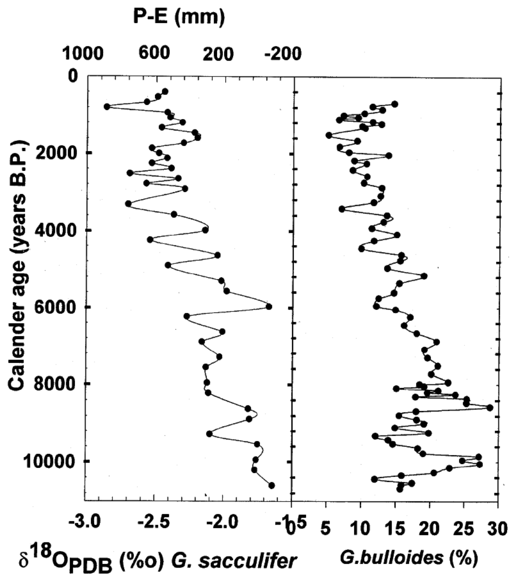

Figure 1. Comparison of palaeorecords of monsoon *rainfall* (excess of precipitation over evaporation, mm, see top scale), derived from the stable oxygen isotopic composition of *G. sacculifer* in a sediment core 3268G5 from the eastern Arabian Sea[20], with that of monsoon *wind*, derived from the abundance of *G. bulloides* in core 723A from the western Arabian Sea[17].

on a growing speleothem decreases with increasing precipitation and hence $\delta^{18}O$ of the speleothem is a good proxy for the past variations in rainfall.

Amount dependence of $\delta^{18}O$ of rainfall

Tropical regions are characterized by converging air masses that are forced to move vertically rather than horizontally. They are cooled due to adiabatic expansion, while surface temperature gradient remains negligible. Amount effect is believed to be[41,42,45] due to the preferential loss of $\delta^{18}O$ from the cloud as rainout continues, and a gradual saturation of air mass below the cloud base, an effect that diminishes evaporation during precipitation and hence shifts towards lower $\delta^{18}O$ values.

For island stations in the equatorial belt annual temperature fluctuation remains within a narrow range; therefore, the amount of precipitation is largely dependent on the air circulation patterns. For such stations a linear relationship[46] between the mean monthly $\delta^{18}O_m$ of precipitation and the mean monthly rainfall is observed:

$$\delta^{18}O_m = (-0.015 \pm 0.002)*P_m - (0.47 \pm 0.42).$$

P_m is the mean monthly rainfall, with linear correlation coefficient $r = 0.87$ for 14 island stations (each has at least 40 monthly observations). Average rate of depletion is found to be $-1.5 \pm 0.2‰$ for a 100 mm 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[28] to be $-2.2 \pm 0.8‰$ (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 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. Thus $\delta^{18}O$ of speleothem can be converted into the amount of monsoon rainfall. Similarly, in the stable hydrogen isotopic composition of monsoon rain (samples at Orissa), there is an amount effect (Figure 2): $\delta D_d = -31.5–0.55\ P_d$, where P_d is the daily rainfall in mm ($r = 0.81$, significant at 0.05 level), and for the monthly average, $\delta D_m = -3.6–0.17\ P_m$ ($r = 0.93$). Figure 3 shows that both δD_d and $\delta^{18}O_d$ are well related: $\delta D_d = 7.43\ \delta^{18}O_d + 2.57$ ($r = 0.95$). Using these data also, one can expect an amount effect in $\delta^{18}O_m$ of $(-0.17 \times 100/7.43=)\sim -2.3‰$ per 100 mm increase in rainfall.

The present day annual rainfall in India shows large geographical variations[11] and therefore, ideally one should have a calibration of amount of the annual rainfall and its $\delta^{18}O$ for each cave site, for a more accurate rainfall reconstruction. It is possible to calculate the spatial variations of the amount effect at least in the Indo-Gangetic belt under some simplifying assumptions: Suppose that a vapour enters India at the Head Bay, with a certain initial isotopic composition (equally applicable to stable hydrogen and oxygen isotopes) and travels progressively inland and rains out. If M_i, and δ_i are respectively the mass and isotopic composition of the vapour mass before it enters a given location (i denotes initial) and M_f, δ_f, the values after it has rained out in the location (subscript f denotes final), the amount of rain being $a = M_i–M_f$, and isotopic mass bal-

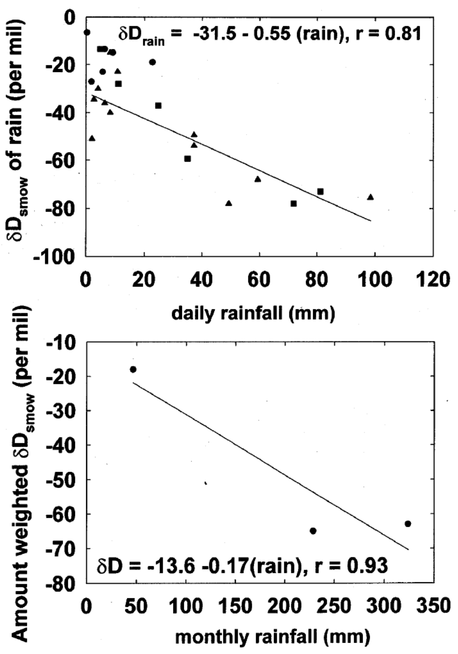

Figure 2. Amount effect in the hydrogen isotopic composition of monsoon rainfall quantified in daily (top) and monthly (bottom) rainfall. In the top panel, circles denote July, squares August and triangles September rain events. The linear correlation coefficient, r is also shown.

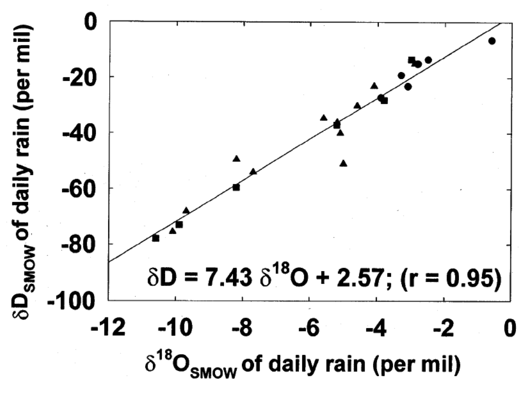

Figure 3. Correlation between stable hydrogen and oxygen isotope ratios in monsoon rainfall. Circles denote July, squares August and triangles September rain events.

ance gives the isotopic composition of the rain at the location:

$$\delta_r = (M_i\, \delta_i - M_f\, \delta_f)/a.$$

If the fraction of vapour mass that remains after the rainout at the location is $f\,(= M_f/M_i)$, then

$$\delta_r = M_i(\delta_i - f\,\delta_f)/a.$$

We know from the Rayleigh distillation equation that $(1 + 0.001\delta_f)/(1 + 0.001\delta_i) = f^{\alpha-1}$, where α is the isotopic fractionation factor between the rain and vapour (magnitude greater than unity, as the heavier isotopes prefer the liquid rather than the solid state during condensation). Using $f = 1-(a/M_i)$:

$$\delta_r = \delta_r^o - [(\alpha-1)(\delta_r^o + 1000)/2]a/M_i.$$

The above equation gives the amount effect as $-[(\alpha-1)(\delta_r^o + 1000)/(2M_i)]$, always negative, because M_i, $1000 + \delta_r^o$ and $(\alpha-1)$ are always positive. The intercept signifies the isotopic composition of the first condensate. It is clearly seen that higher the (time averaged) mass of vapour that reaches the location, lower will be the amount effect at the location (it is assumed that when the vapour source and the storm track do not change significantly year to year, M_i has a small variance around a mean). Thus closer to the coast, the amount effect is likely to be much lower than that in the interior. Thus speleothems in the interior may be more sensitive recorders of past monsoon variations. On the other hand, as the above calculation ignores the re-evaporated component in the rain, it is likely to over-estimate the sensitivity in the interior.

For example, the amount effect in the summer monsoon rainfall of New Delhi, where both $\delta^{18}O$ and the monthly rainfall data[47] for the period 1960 to 1995 AD, is represented by:

$$\delta^{18}O_m = (-0.013 \pm 0.003)*P_m - (2.82 \pm 0.60)$$

with a correlation coefficient of 0.4, significant at 0.001 level ($n = 125$, Student's t value 4.75). The correlation improves if we disregard the unusually high rainfall (> 350 mm) events during this period:

$$\delta^{18}O_m = (-0.027 \pm 0.004)*P_m - (1.24 \pm 0.71)$$

with a correlation coefficient of 0.5, significant at 0.001 level ($n = 107$, $t = 6.08$).

The noise in the data (unexplained variance of 75–80%) is possibly due to the two different sources of moisture for the rainfall at New Delhi, the Arabian Sea and the Bay of Bengal).

Intercomparison on proxy monsoon records

Two other (comparable in resolution with speleothems[28]; i.e. ~ 1 year to ~ 10 year) palaeomonsoon records from southern Asia, spanning the last 3400 years are available: (i) a high resolution (~ 7 year) record has been reconstructed[22] using the thickness variations in the varved sediments collected from off Karachi, Pakistan (northeastern Arabian Sea): precipitation and hence the river runoff have been assumed to control varve thickness. The precipitation at the sampling site occurs both during the summer and winter monsoons. These alternatively dark and light coloured sediment sequences form an annual couplet. The precipitation may have fluctuated 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. (ii) Another high resolution (~decadal) record is based on the stable oxygen isotope variations from stalagmites in southern Oman[23], though not calibrated with the amount of local rainfall.

The variations in the varve thickness are presented[22] in the middle panel of Figure 4. The top panel shows the stable oxygen isotope data of a stalactite from Gupteswar cave[28], Orissa; it has different time resolutions in different periods. Varve thickness is a 7-point moving average of annual thickness data[22] and hence our data values were smoothed by an appropriate running mean during different time periods. The bottom panel depicts stable oxygen isotope ratios of the Oman stalagmites[23].

We see that Gupteswar (GUP) and Oman speleothem records are well matched. The monsoon was stronger around 3000 BP as indicated by more depleted $\delta^{18}O$ values and also by the increased growth rate (higher sampling den-

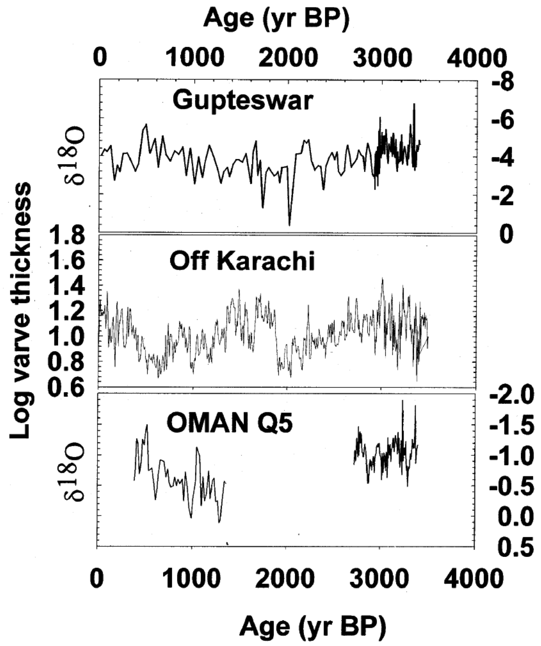

Figure 4. Comparison of high resolution palaeomonsoon records (top panel) from Gupteswar stalactite[28] (Orissa), (middle panel) varved sediments[22] (northern Arabian Sea) and Oman stalagmites[23] (bottom panel).

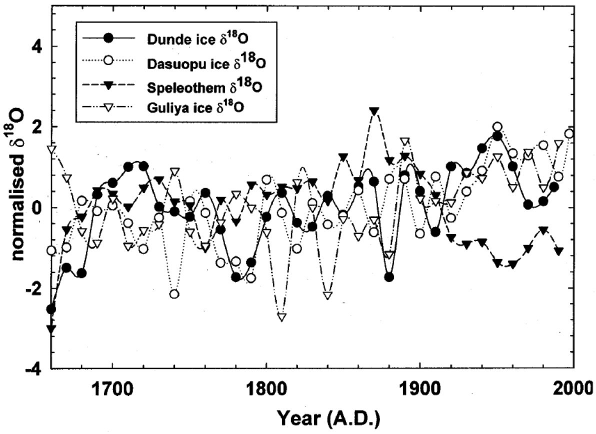

Figure 5. Comparison of normalized decadal oxygen isotope records from three ice cores[29–31] (Dunde, Dasuopu and Guliya from Tibet); and a stalagmite[27] from northern Karnataka.

sity). The increasing trend between 1200 yr BP and 400 yr BP is seen in both the records. Also during the extremely low rainfall epochs of 1700 and 2000 yr BP shown by GUP, the Oman stalagmites did not grow, probably due to the complete lack of rain. It must be noted that Oman is more like a desert relative to eastern peninsular India and therefore growth of the stalagmite itself is more sensitive to rainfall fluctuations. The good concordance of these two records lends credence to our assumption that at least for the past 3400 years, the radiocarbon chronology and our assumption regarding constant dilution by dead carbon[28] have not been invalid.

The decreasing trend of rainfall from 3400 to 1900 yr BP is reflected both in the varve and GUP records. But the two records differ significantly during the last 1500 years. Pollen records from different lakes in Rajasthan[48–50] show higher monsoon rainfall around 600 yr BP, similar to GUP data. As the speleothem records separated by a larger distance agree very well, it is likely that the response of the varve thickness to the monsoon is nonlinear.

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 palaeomonsoon indicator[29–31] (Figure 5). Subtracting the respective means and dividing by the respective standard deviations we first normalized the decadal $\delta^{18}O$ series from the Guliya, Dasuopu, and Dunde ice caps in Tibet. These are compared with another palaeomonsoon record with annual resolution, from a stalagmite from northern Karnataka[27]. This time series had 6 to 12 measurements per decade (note that though there were annual layers, all layers could not be sampled for stable isotope measurements, because of very narrow layers). Decadal averages were calculated and normalized as the other series before comparison. The good agreement between the stalagmite record (filled triangle) and Dunde ice record (filled circle) up to the beginning of the industrial revolution (~ 1850 AD), clearly breaks down afterwards. In the last century the glaciers show a warming trend and the monsoon shows an increasing trend, i.e. a decreasing trend in the $\delta^{18}O$ values (recall that although there is no observed trend in the all-India monsoon rainfall, the west coast shows an increasing trend in the instrumental records).

Conclusions

We have compared the high-resolution, precisely dated palaeorecords of the South Asian summer monsoon during the Holocene. Our analysis reveals contrasting monsoon signals from the eastern and the western Arabian Sea: if the interpretations of the primary data are correct, then it appears that over the Holocene, the monsoon winds weakened, causing reduced upwelling, in phase with the 8% reduction in insolation, while precipitation over India increased. This problem needs to be resolved by using multiple proxies rather than a single proxy such as *G. bulloides* abundance. Amount effect is demonstrated in the monsoon rainfall stable isotope ratios of both oxygen and hydrogen; which offers a unique method to quantify palaeomonsoon variations with decadal to sub-decadal time resolution using the stable oxygen isotope variations of speleothems. A simple model to calculate the geographical variation in the amount effect has been outlined, which implies a lower sensitivity at the coast than inland. Quantitative monsoon reconstruction now available for the past 3400 years is in good concordance with stalagmite oxygen isotope record of Oman, and also with varve records in mid-Holocene. However the varve response to changing monsoon appears to be nonlinear. High-resolution oxygen isotopic record from a speleothem in Karnataka indicates the recent trend in the increasing monsoon rainfall in the west coast, as surface temperature increased in Tibet, shown by oxygen isotope variations in Tibetan ice.

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