Addressing past monsoon variability from speleothems

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Numerous observations indicate that speleothems can record signatures of the past climate variability. Systematics of stable isotopes and trace elements in speleothems and current methods of dating are discussed. $\delta^{18}O$ in Indian speleothems is presently being used as a monsoon proxy. Several records from Indian karst locations are available, many of these cover recent several millennia and some extend back to ~280 ka. Salient features of monsoon variability reconstructed so far from Indian speleothems is briefly discussed. Some of the $\delta^{18}O$ records show presence of non-persistent periodic changes suggesting controls of subtle variations in solar output and internal changes in the climate system.

Keywords: Indian monsoon, palaeoclimate, periodic analysis, speleothem.

Introduction

INDIAN economy depends primarily on the crop yield, which in turn is decided by the variability in monsoon rain. Severe drought or flood events when persistent for a long period of time, may lead to country wide devastation. Since the Indian Summer Monsoon (ISM) contributes to ~80% of the annual rainfall in India, the economy of the country critically depends upon the performance of the monsoon[1]. Understanding the processes affecting rainfall occurrences requires continuous instrument-based observations of meteorological parameters, and in parallel development of mathematical models validated by the observations. The annual variations are well recorded by meteorological stations, but are limited to past 100 to 150 years. Due to their limited time coverage, the 'instrumental' or 'systematic' records do not capture all modes of climate variability. Climatic anomalies such as the occurrence of glacial phases, transitions of glacial and interglacial periods and variability of monsoon as reflected in the frequency and intensity of droughts need to be examined with the datasets much longer than the instrumental records. Using palaeoclimate records, performance of climate models beyond instrumental period can be tested. Successful simulations of the past would ensure that climate system is very well understood and it can be used to predict the future.

So far model simulations have successfully produced large scale global patterns of changes in climate at the LGM and mid-Holocene, however, simulated magnitude of regional changes is often not as large as the observed magnitudes[2]. Model simulated Indian monsoon changes when compared with the limited available palaeoclimate data[3] show a good spatial agreement during the Medieval Warm Period (900–1100 AD) and disagreement during the Little Ice Ages (1515–1715 AD). High resolution palaeoclimate records are therefore essential in this regard.

The growth rate of trees, speleothems and chemistry of foraminiferal tests are influenced by variables such as precipitation and temperature. In terrestrial proxies, focus is mainly on results based on tree-rings and speleothems (cave deposits such as stalactites and stalagmites), because these are well dated and used for high resolution monsoon reconstructions[4–11].

In the past two decades, the use of speleothems as a palaeoclimate proxy has developed to be a frontier research topic. The growing interest and advancements in this field are due to: (i) potential palaeoclimate records may show annual to decadal resolution[12]; (ii) contrary to radiocarbon dating technique, using U/Th dating method the reconstructions can be extended to late Pleistocene period[13,14], and (iii) they have wide geographic extent, hence are excellent tool to study global climate teleconnections[15–20]. Use of micromills and automated measurements on high precision high-throughput mass spectrometers on microgram of samples, produce high resolution continuous time series[21,22].

Speleothems as palaeoclimate archive

The word speleothem is derived from Greek words 'Spelaion' meaning cave and 'thema' meaning deposit[23,24]. The speleothems used for palaeoclimate studies are made of calcite or aragonite or a mixture of the two. A schematic representation of speleothem formation is shown in Figure 1. Partial pressure of carbon dioxide (P_{CO_2}) in equilibrium with rain is $\sim 10^{-3.5}$ bars making it a very weak acid. However, as rainfall percolates further into the soil, it dissolves a higher amount of additional CO_2

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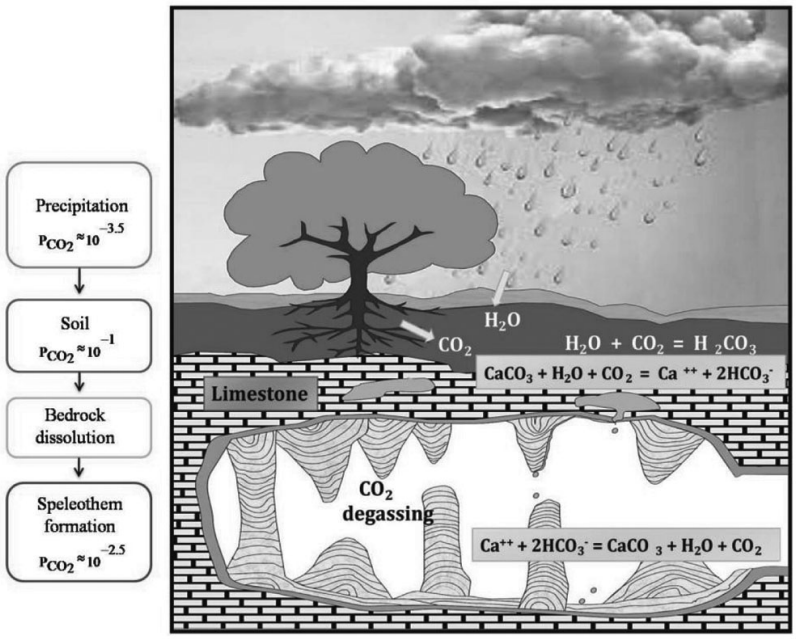

Figure 1. Conceptual model explaining the formation of speleothems in a karstic cave. Typical values for partial pressure of CO_2 in rainfall, soil and cave air are taken from Sasowsky and Mylroie[26].

released from microbial decay and plant root respiration[25]. Usually, P_{CO_2} in soil is 10 to 100 times more than the atmospheric level, a typical value may be 10^{-1} bar. This 'corrosive' solution reacts with the host carbonate rocks in epikarst and forms HCO_3^-, CO_3^{-2} ions[26]. As the saturated water descends through the crevices and fractures into the karst, it comes in contact with cave air having a lower P_{CO_2} (~$10^{-2.5}$ bars). This leads to degassing of CO_2 from the solution and precipitation of $CaCO_3$ in the form of calcite or aragonite. Of the various morphotypes, stalactites and stalagmites are favoured for palaeoclimate studies as the layers may be unperturbed and have less detrital content. Stalagmites are mostly preferred over stalactites, as they are cylindrical with near flat layers along the growth direction. This has advantages when extracting reasonable amount of sub-samples for isotope and age analysis along each layer as opposed to conical growth layers of stalactites, which may also be contaminated and interrupted in the centre.

$\delta^{18}O$, $\delta^{13}C$ and trace elements in speleothems

Oxygen isotopes

Potential of oxygen isotopes of speleothems for palaeoclimate studies was reported early in the 1960s. Ratio of the stable isotopes ^{18}O and ^{16}O in a sample, relative to the standard material, is measured on a mass spectrometer. Since range of observed values is very small, these values are expressed in per mil (‰, parts per thousand)

$$\delta^{18}O = \left[\frac{\left(\frac{^{18}O}{^{16}O}\right)_{Sample}}{\left(\frac{^{18}O}{^{16}O}\right)_{Standard}} - 1 \right] \times 10^3‰.$$

Oxygen isotope ratio of calcite $(^{18}O/^{16}O)_c$ which is precipitated from the calcite-water system is more than the ratio of water $(^{18}O/^{16}O)_w$. The fractionation factor α_{cw} relates these changes and is also temperature dependent, known from definitive laboratory experiments[27]

$$\alpha_{cw} = (^{18}O/^{16}O)_c/(^{18}O/^{16}O)_w,$$

$$\Delta_{cw} = (2.78 \times 10^6/T^2) - 2.89,$$

where T is the ambient temperature in K and $\Delta_{cw} = 10^3 \ln \alpha_{cw}$.

The temperature inside the cave remains constant throughout the year in poorly ventilated caves because of thermal inertia[28,29]. Changes in the annual temperature affect the $\delta^{18}O$ of the precipitating calcite, where heavier $\delta^{18}O_c$ values imply lower temperatures. The degree of enrichment is given by the equation

$$d(\Delta_{cw})/dT = -0.21/°C^{-1} \text{ at } 25°C.$$

Caves located in the tropics however, show less temperature variability. The $\delta^{18}O_c$ of speleothems in such caves, are strongly dependent inversely on the amount of

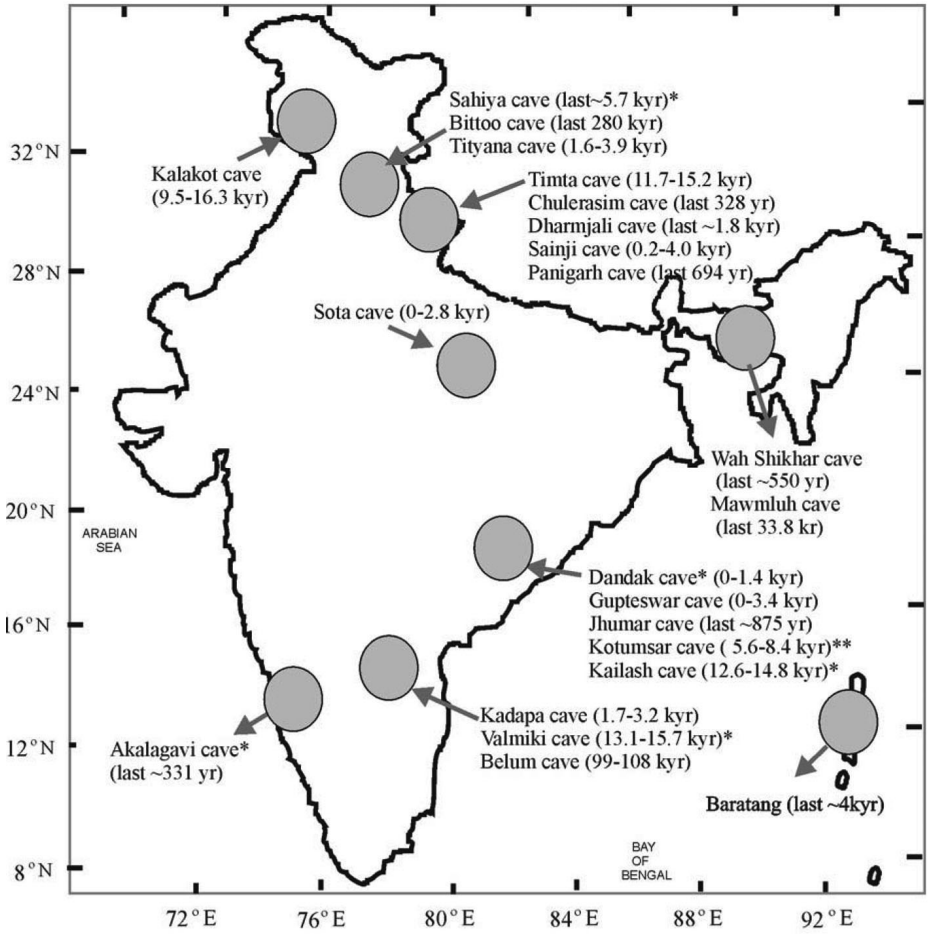

Figure 2. Indian caves studies so far to address past monsoon. Their exact locations and further details can be obtained in the references cited in the text. The time duration for which data is available is shown in the numbers in brackets. *Caves where periodic analysis on the reconstructed data is carried out. **Results of periodic analysis presented in Figure 4.

rainfall[30]. However, isotope-enabled climate modelling suggests that large scale changes in atmospheric circulation system may also contribute to changes in speleothem $\delta^{18}O$ and hence variability in speleothem $\delta^{18}O$ may not necessarily correspond to changes in amount of rainfall[31].

This necessitates verification of $\delta^{18}O$ and amount of rain for modern precipitation samples, ideally near the cave location, for correct interpretation of the speleothem $\delta^{18}O$. Some studies have been carried out to identify isotopic signatures in modern precipitation samples[30,32–34]. The ratio of oxygen isotopes in speleothems can be traced back to the processes controlling the hydrological cycle. Evolution of ratio of oxygen isotopes in speleothems depend upon the phase changes on its course from ocean water → vapour → precipitation → soil water → epikarst solution → karst drip water. Average meteoric precipitation for Earth has a value of –4.5‰ (ref. 35), whereas local rainwater may have large range of values, e.g. from –3 to –10‰ for the Indian sites[30,32,36,37]. Evaporation on the surface and mixing in the epikarst zone may modify these signatures and make these further enriched. Drip water $\delta^{18}O$ values are reported from Dandak cave and Timta cave (Figure 2), these are –1.8‰ (ref. 38) and –8.9 to –11.0‰ (ref. 36) respectively. The $\delta^{18}O$ value of the soil water is determined by the $\delta^{18}O$ value of the precipitation infiltrating the soil pores. Processes such as evaporation and transpiration play a crucial role in controlling the $\delta^{18}O$ in soil pores. In arid regions, evaporation leads to enrichment in ^{18}O, whereas in humid climates the role of evaporation is minimal[19,39]. In order for speleothems to track hydrological changes, it is important that the isotopic equilibrium is maintained between the drip water, degassing CO_2 and precipitating $CaCO_3$ (ref. 40). When there is a rapid loss of carbon dioxide or water then speleothem growth is non-equilibrium type and hence empirical relations obtained for isotopic ratios cannot be applied. Such non-equilibrium type deposits or periods of deposition can be tested if $\delta^{18}O$ and $\delta^{13}C$ are significantly correlated along the flank part of a single growth layer. This test is called Hendy's test and is practically feasible if laminations are clearly seen so that sub-samples are collected genuinely from the same[41].

Carbon isotopes

Similar to the $\delta^{18}O$ expression, stable carbon isotope ratio is expressed as $\delta^{13}C = (((^{13}C/^{12}C_{sample}/^{13}C/^{12}C_{standard}) - 1) \times 10^3$, in per mil. The dissolution reaction which occurs in the limestone part is reversed in the cave environment, where Ca^{++} and HCO_3^- ions recombine to precipitate $CaCO_3$; this is shown by the precipitation reaction in Figure 1. Carbon in speleothem-$CaCO_3$ is derived from two sources: (i) soil CO_2 which is much lighter (–35‰ to –12‰) than the atmospheric carbon dioxide, with a $\delta^{13}C$ value of about –7‰ (ref. 42) and (ii) bedrock carbon with $\delta^{13}C$ value close to 0‰. Evolution of the carbon isotopes is a complex process and depends upon several factors[40]. Due to a systematic kinetic type isotopic fractionation involved in the biology, C3 and C4 type of plants produce biogenic carbon (degradation of which results in soil CO_2) with $\delta^{13}C$ values around –35‰ to –21‰ (ref. 43) and –10‰ to –16‰ (ref. 44) respectively. Hence, in principle $\delta^{13}C$ of the soil CO_2 should depend upon the type of vegetation cover. Since climate driven changes can affect the type of vegetation; therefore, it may affect $\delta^{13}C$ of speleothems[35]. However, change in vegetation type, in response to climate variation is usually a slow process, observed on centennial timescales or more.

The $\delta^{13}C$ of calcite is also influenced by cave setting vis-á-vis, closed type dissolution and open type dissolution. For example, if water seeping through the bedrock and soil zone remains in contact with soil CO_2 environment enroute, it will be open type dissolution and in such a case $\delta^{13}C$ of soil CO_2 will predominantly influence the carbon isotopic composition of the dissolved ionic species and subsequently the speleothem $\delta^{13}C$. However, if at the initial stage itself, once the bedrock carbonate is dissolved, the seepage water is kept isolated from further interacting with the soil CO_2, the bedrock $\delta^{13}C$ will also have significant effect on the speleothem $\delta^{13}C$. This being an example of closed type dissolution may happen when bedrock fissures are very narrow and well isolated from the soil CO_2 environment.

Considering the above mechanisms, on longer time scales (centennial or more) unless variation in climate causes significant changes in vegetation cover or the type of dissolution, there may not be a noticeable variation in the speleothem $\delta^{13}C$.

On shorter time scales (comparable to drip rate) a Rayleigh type of carbonate precipitation[45] may be envisaged that may contribute to variability in $\delta^{13}C$ of carbonate precipitation. Since isotopically lighter ionic species of the drip water are more mobile, preferentially they will precipitate first as carbonate. If drip water stays for short duration on the tip of the freshly growing speleothem, then more of lighter ionic species will go in the solid phase ($\delta^{13}C$ is depleted). However, in case it stays for long duration, heavier ionic species also go in the solid carbonate phase ($\delta^{13}C$ will have relatively enriched values). A good correlation between $\delta^{18}O$ and $\delta^{13}C$ time series (along the growth axis, not on the flank part as discussed above for Hendy's test), in a speleothem (that grew under high humidity condition in a deeper locations of caves) may be due to such a Rayleigh type of carbonate precipitation process. Here, depleted levels in $\delta^{18}O$ can be attributed to enhanced ISM and in $\delta^{13}C$ due to increased dripping rate.

In case of a small size cave where reasonable air circulation is often realized, evaporation may affect $\delta^{18}O$ (making more positive) and rapid degassing of CO_2 from dripping water may also affect $\delta^{13}C$ (more positive value). Although such effects may be suppressed during wet seasons (due to high humidity), they could dominate during dry seasons (contribute to enrichment) and hence a good and significant correlation in $\delta^{18}O$ and $\delta^{13}C$ time series may be observed because of the role of air circulation in such caves.

A kinetic type of isotope effect resulting from hydration and hydroxylation of dissolved carbon dioxide is also likely to show a positive correlation between $\delta^{18}O$ and $\delta^{13}C$ time series[46]. Overall $\delta^{13}C$ will depend upon vegetation cover (C3/C4), dissolution (open/closed), drip rate, cave air circulation and kinetic processes.

During speleothem formation, mineralogy may be either calcite or aragonite or a mixture of both[47]. Based on occurrences of various sedimentary carbonate deposits, it is observed that aragonite precipitation is favoured if supply of carbonate ion (CO_3^-) is high, else calcite is formed[48]. Also, more of ^{18}O and ^{13}C are incorporated in aragonite compared to calcite: laboratory experiments have shown that oxygen in aragonite is enriched by ~0.6 % and carbon by ~1.8% (refs 49, 50). Mineralogy can be confirmed from X-ray diffraction (XRD) method. Before making palaeoclimate inferences, a weighted mean approach can be applied to the stable isotope data of a speleothem having mixed mineralogy to have a pure calcite (or aragonite) equivalent stable isotope data[47].

Trace elements

$CaCO_3$ while precipitating incorporates trace elements (Tr) from the solution. In the solution, various elements form divalent cations and substitute for Ca ion in the $CaCO_3$ crystal lattice[51,52]. Atoms of elements such as Mg, Sr, Ba with similar ionic radii, substitute for Ca under different physical conditions.

$$CaCO_3 + Tr^{2+} \leftrightarrow TrCO_3 + Ca^{2+}.$$

A simple equation is used to define the distribution coefficient to relate solution and mineral compositions

$$(TrCO_3/CaCO_3)_{speleothem} = K_{Tr}(Tr/Ca)_{solution},$$

where Tr is the trace element in ionic form and K_{Tr} is the distribution coefficient (value less than 1). Cations such as Mg^{2+}, Sr^{2+}, Ba^{2+} with similar ionic radii substitute for Ca^{2+} under different physical conditions[53]. Since ionic radius of Mg (0.72 Å) is less than that of Ca (1.00 Å), it is likely that it substitutes Ca in the growing speleothem lattice. Since temperature increase raises the rate of diffusion of Mg, in principle, Mg substitution (or Mg/Ca ratio) should be sensitive to temperature changes. Higher values of Mg/Ca in the speleothem profile should correspond to increase in temperature. Gascoyne[54] measured Tr/Ca ratio in modern seepage water and calcite deposited in different caves which had different ambient temperatures. It was found that the Mg/Ca ratio in speleothems deposited at different temperatures correlated with ambient temperature of a cave. Between temperatures ranging from 7 to 24°C, $K_{Mg} = +0.0017/°C$. As Sr and Ba have larger ionic radii (1.16 Å and 1.36 Å respectively) relative to Ca, non-lattice substitution may be possible during carbonate precipitation. This is observed to have no temperature dependency, but is influenced by the source conditions[54,55] such as residence time of water in the epikarst and hence indirectly on rainfall[56]. In the same study, the K_{Sr} was found not to have any clearly observed temperature dependency, although it showed sensitivity to the seepage water conditions, therefore Sr/Ca can indicate source conditions. Behaviour of Ba/Ca is yet to be understood[57], suggesting that the growth and incorporation of Tr in such materials may be complex. Other than substitution, trace elements come through the aqueous medium where they are adsorbed on the surface of detrital particles and hence get incorporated in speleothems.

Carbonate may also be precipitated before dripping on the speleothem tip, along the seepage pathway if partial degassing is possible. This may happen if there are gas filled conduits in the soil, leading to carbonate precipitation in the pathways, known as prior calcite precipitation (PCP). For example, during drier climatic conditions conduits may be partially dry, the resultant degassing leads to supersaturation of water for $CaCO_3$ and carbonate precipitation[51,58,59]. As K_{Tr} values for the carbonate precipitated are less than one, there is larger reduction of Ca in solution than that of the trace element and hence an increase in the ratio of trace element to Ca in solution[60]. Co-variability among trace element ratios Mg/Ca, Sr/Ca and Ba/Ca may be due to changes in degree PCP[51]. PCP could also favour aragonite formation in speleothems[59,61].

Methods for chronology

If speleothem has clearly observable layers, these may be annual in nature and in such cases stable isotope record is assigned chronology by counting individual layers (e.g. Yadava *et al.*[62]). If they are not annual or distinctly seen, radiometric dating methods (e.g. ^{14}C and by U-series) are used for age estimation. Radiocarbon dating can be used if speleothems are younger than the dating limit of the method which is 40–50 ka. Since carbon in speleothems is usually from both soil environment (zero in age) and from bedrock (geological origin, hence free of ^{14}C), knowledge of their proportion in the dripping water is required. This is estimated from the apparent age of the actively growing calcite surface (e.g. instead of zero age it may show a finite age) and by subtracting it from the other estimated ages, true ages are calculated for different locations of the speleothem (e.g. Yadava and Ramesh[30]). However, this is based on assuming the same proportion of dilution by bedrock carbon throughout the growth history of the speleothem. This may be hardly true in real case since internal hydrological pathways are expected to keep changing with time. U–Th decay systematics is another dating method used on carbonates with accretionary growth and deposited as a closed system such as corals and speleothems. ^{238}U–^{230}Th disequilibrium (U–Th dating) which is based on ^{238}U–^{234}U–^{230}Th decay chain has proven to be a reliable technique for dating speleothems[63]. In an ideal situation, carbonate bedrock in which ^{238}U–^{234}U–^{230}Th is in secular equilibrium, when dissolved by acidic percolating rain water, contributes U–Th to the seepage water. However, due to the distinct difference in their chemical behaviour, U remains in dissolved form and Th is attached to the detritus particles. Later when carbonates are precipitated as speleothems, only U is incorporated in the calcite and Th still remains attached to detritus particles which go along with the water pathways and discarded away from the growing speleothem location. With time the U decays into ^{230}Th, their subsequent rise in numbers towards attaining secular equilibrium, is used to estimate age of the speleothem[13]. This is an excellent method currently employed for precise age estimation of speleothems from hundreds to over 500,000 years[13,64]. However, in real situation most of the speleothems also trap detritus particles in the growing carbonate matrix and hence initial ^{230}Th is also contributed, posing challenge to account for it by analytical methods and also while applying corrections in the final age estimate. Sometimes speleothems having lot of detritus content (called 'dirty' speleothems) need to be attempted by radiocarbon dating method[8,65].

Laboratory techniques

The stalagmite samples are first cut along the growth direction to see the patterns of laminations and decide the strategy for sub-sampling using micro drill machine. Computer aided drill machine (such as Micromill from New Wave Research) is preferred for large sampling throughput with precise control on the sampling tracks and repeatability of sample amount (around 200–500 μg). Carbonate samples are then converted to CO_2 by allowing it to react with 100% H_3PO_4 and the liberated CO_2 is

directed to the stable isotope ratio mass spectrometer subsequently for isotope ratio measurements.

Two methods are currently adopted for dating by radiocarbon method. The first and the old method called conventional technique, by liquid scintillation spectrometry (LSC) requires sample carbon of speleothem to be converted into benzene. Residual beta decay of benzene is measured by standard scintillation spectrometry approach[4]. The second method is by using accelerator mass spectrometry (AMS) method which requires carbon to be converted into graphite; the ratio of $^{14}C/^{13}C$ or $^{14}C/^{12}C$ is measured in the accelerator mass spectrometer[66,67]. Radiocarbon dating by conventional method requires more than 10–20 g of powdered sample, as against the Accelerator Mass Spectrometry based method where about 5–9 mg powder is sufficient for age estimation.

For dating by U–Th method, dissolution of U and Th by ultra-pure acids and then subsequent separation of U–Th by ion-exchange column chemistry, under ultra-clean environment is required[13]. The U–Th solutions are injected into the Multi-collector–Inductively coupled Mass Spectrometer (MC-ICPMS) for the required isotope ratio measurements.

For measuring concentrations of trace elements, standard methods such as ICPMS[68,69] are followed.

Speleothem work in the Indian context

While moisture sources during ISM, for the major part of India is Arabian Sea and Bay of Bengal, for sites at higher latitudes westerly winds passing through the Mediterranean Sea also contribute to it significantly. Use of speleothem as a proxy to reconstruct the Indian monsoon, is quite well established. Speleothem formations are found where there are large outcrops of limestone bedrocks. Locations of the speleothem-based studies, in different parts of India carried out so far are shown in Figure 2 and summary of the key results are presented below.

North India

A few high-resolution climate reconstructions from the central part of the Indian Himalaya are available. Imprints of Older Dryas, Allerød period and Younger Dryas, during 14.3 to 12.2 ka, were observed in a stalagmite from Kalakot cave[70]. High-resolution (~1.8 year) $\delta^{18}O$ record from Sahiya cave over the last 5.7 ka showed significant shifts in monsoon rainfall that possibly affected human civilization[71,72]. Over larger time scale (last 280 ka) $\delta^{18}O$ data was reported from Bittoo cave. It shows strong coherence between $\delta^{18}O$ records from North Indian and Chinese speleothem influenced by the East Asian monsoon. Both are found to have responded to the precession induced insolation changes[73]. Radiocarbon dated stalagmite from Tityana cave, from 1.6 to 3.9 ka showed trends in $\delta^{18}O$ that may be due to variability in the relative contributions of the two moisture sources, viz. Indian summer monsoon and western disturbances[65]. Multidecadal climate variability similar to past radiocarbon changes, between 11.7 and 15.2 ka, was reconstructed by a stalagmite from Timta cave[36]. Little Ice Age signature was seen as shift in values of stalagmite $\delta^{18}O$, over the last 328 years in Chulerasim cave[74], 1.8 ka in Dharamjali cave[75] and 694 years in Panigarh cave[76]. Stronger westerlies may have contributed to high precipitation as was seen in a 4 ka old stalagmite from Sainji cave[77]. Stalagmites from Dharamajali cave were also used to identify major earthquakes events that occurred possibly at ~4.3 ka, 2.8 ka, 2.5 ka, 1.5 ka, 1.3 ka and 0.7 ka (ref. 78).

Northeast India

Multicentennial length episodes of states of ISM were observed over the last ~550 years in a stalagmite from Wah Shikhar cave[79]. A study from Mawmluh cave, Meghalaya reconstructed ISM variability from 33.8 to 5.5 ka (ref. 80). Abrupt increase in rainfall during Bølling–Allerød and early Holocene, and significant weakening during Younger Dryas and Heinrich cold event was observed. Short duration climate anomaly observed as shift in speleothem $\delta^{18}O$ from Mawmluh cave, indicating prolonged decrease in Indian monsoon at 4.2 ka (ref. 81) is used to rename 'late-Holocene' as 'Meghalayan age'[82]. A high resolution data (3.78 to 4.44 ka) from the same cave showed that ISM had abruptly decreased with onset of reduced ISM[83] at ~4.0 ka. Based on analysis of glycerol dialkyl tetraethers in a stalagmite from Mawmluh cave, estimated temperature during LGM was found to be lower by ~4°C (ref. 84).

Central India

Several caves are known in Bastar district, Jagdalpur. This area falls in the core monsoon region of India; speleothems formed here have responded to changes in the amount of monsoon precipitation[30]. Monsoon variability during 14th and 15th century was captured in a 900 years (600–1500 AD) old stalagmite from Dandak cave[85]. Major famines like Durga Devi famine that lasted from 1396 to 1409 AD during 'Little Ice Age' coincides with the enriched oxygen isotope values. Higher precipitation during 'Medieval warming' during 900–1350 AD was also recorded in the stalagmite. Variability of monsoon during Little Ice Age and Medieval warming was correlated with change in the solar activity[85]. In the same cave, buried charcoal layers (radiocarbon ages 4 to 6 ka) were reported that seems to be evidence of human presence[86].

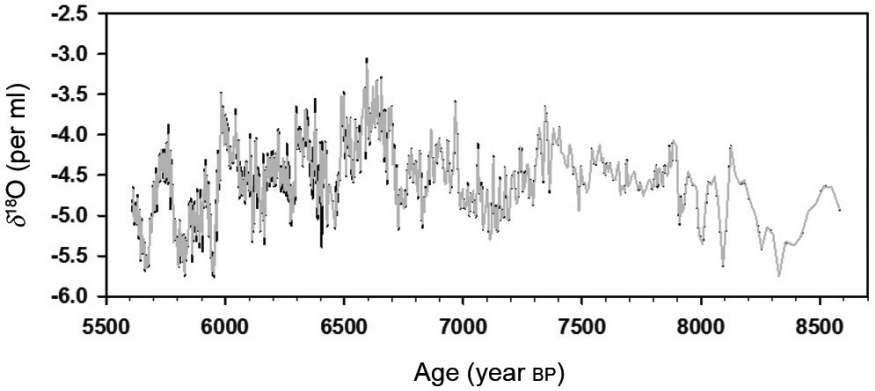

Figure 3. Raw data of stable isotope composition for the Kotumsar stalagmite (in black). For wavelet analysis unevenly spaced data is interpolated to make it evenly spaced at yearly scale (grey, in front) which follows the fine details of the raw data. δ^{18}O data is taken from Band *et al.*[10].

Based on a 3400-year-old Gupteshar stalactite, it was inferred that high rainfall persisted from 3400 to 2900 year with declining monsoon intensity during 2900–1200 year. Since then, increase in precipitation was recorded till present[5,30]. A calcite stalactite from Sota cave (covering last 2.8 ka, age by conventional radiocarbon dating method) showed varying δ^{18}O scenario[5]. High resolution data up to 875 yr from Jhumar cave shows multicentennial long states of ISM[79]. A high resolution oxygen isotope time series, during 8.4–5.6 ka based on a stalagmite is reported from Kotumsar cave[10]. It registered abrupt climate changes that occurred at 8.2 and 5.9 ka, concurrent with similar changes in North Atlantic. The recent study on Kailash cave stalagmite[20] highlights ISM variability during the last deglacial period from 14.8 to 12.6 ka, covering Bølling–Allerød period; it is attributed to the past changes in the ocean–atmospheric circulation driven by solar forcing.

Peninsular India and Andaman Islands

Some studies are reported from peninsular India. A 1 ka record from Valmiki cave during 15.7–14.7 ka showed large fluctuations at decadal and multi-decadal level in ISM[9]. Using another stalagmite[87] from the same cave, the record was extended further from 15.6 ka to 13.1 ka. Variability in Valmiki δ^{18}O may have been due to past changes in ISM or the flooding events[87] as well. A stepwise shift in δ^{18}O in a speleothem from Belum cave was reported during 99–108 ka period, suggesting abrupt reduction in monsoon rain around that period[88]. Solar forcing and strong ocean–atmospheric circulation were suggested as possible controlling factors of the ISM dynamics[88]. Rapid climate change at ~2.6 ka was observed in δ^{18}O record of a stalagmite from Kadappa cave[89]. A ~331-year-old stalagmite from the Akalagavi cave of Northern Karnataka, India, revealed distinct annual layers. Variability in the monsoon rain was observed during CE 1650–1997, with observation of highest precipitation at CE 1666 and the lowest around CE (ref. 38) and presence of 22 solar year solar cycle in the δ^{18}O record[90]. A radiocarbon dated stalagmite from Baratang cave, located in the Andaman Islands where signatures of amount effect is recently established[37] showed weaker ISM during the Roman Warm Period (2100–1800 cal BP) and strong ISM during the Medieval warming[8].

Additional details on Indian speleothem-monsoon work can also be obtained in Kaushal *et al.*[91]. Since limestone outcrop in India is well documented for mineral exploration purpose, preliminary information on karst formations is available. However, details on existence of caves are rarely presented. It requires several field based studies from researchers familiar with the nuances of the discipline, to locate new accessible caves suitable for palaeoclimate reconstructions.

Periods observed in the Indian speleothem δ^{18}O records

Monsoon is a part of the complex climate system that includes the atmosphere, oceans and cryosphere. Its natural variability on multi-decadal to multi-centennial scales, driven by external and internal forcing need to be known; it can be explored by periodic analysis of the reconstructed data, and by looking for a possibility to attribute it to either solar variability or behaviour of the Earth's own climate system.

A few results on the periodic analysis of Indian speleothem-based reconstructed data is available. Power clustering during 1870, associated with the 21-year solar cycle, in the wavelet spectrum was earlier observed in the reconstructed data covering the last 330 years from Akalagavi cave[90]. Temporal variability in the occurrence of the Gleissberg solar cycle was also observed in the Dandak stalagmite δ^{18}O record[92]. Strong influence of solar forcing on ISM activity, during 13.1–15.6 ka, was

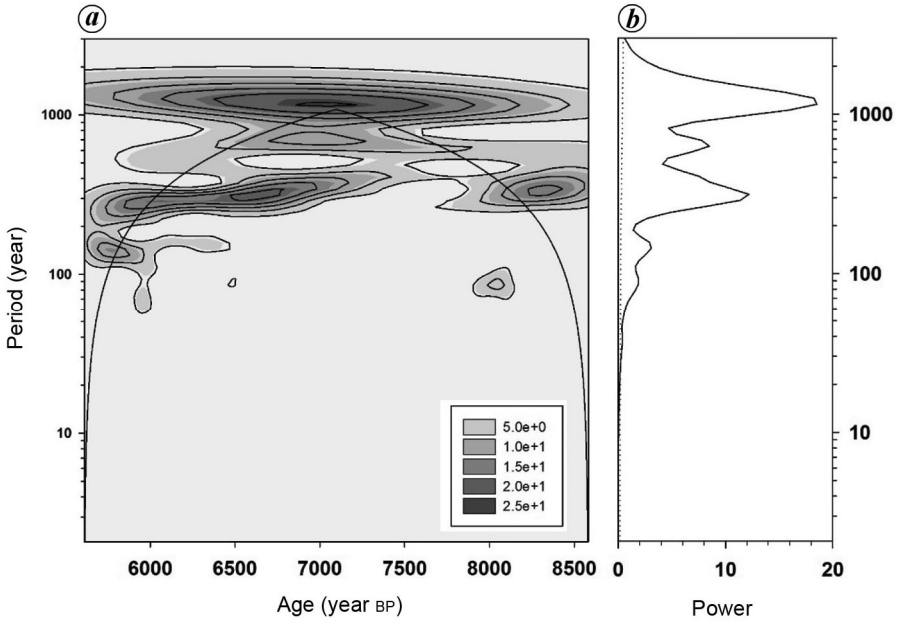

Figure 4. ***a***, Wavelet map of the interpolated δ^{18}O time series. Fourier period (in year) is shown on the *Y*-axis. Line contours show boundaries that are significant at 95% level. Grey level shown in the box is the range of power obtained in the map. Edge effects are important beyond the 'cone of influence' shown by thick curve. ***b***, Global wavelet spectrum for the same data. The dashed line shows the 95% significance level. Wavelet analysis was carried out using codes given by Torrence and Compo[94].

observed in Valmiki cave record[9,87]. Sahiya cave record had revealed significant power for the 60–80 and 15–30-year period[71]. Periodicities close to ~284, ~147, ~66, ~116, ~66 and ~25 year are observed in wavelet output of a stalagmite from Kailash cave[20].

As an example, spectral analysis of the recently available reconstructed δ^{18}O data (Figure 3) from Kotumsar cave[10] is shown in Figure 4; global wavelet spectrum for δ^{18}O shows peaks centered at ~1000, 600, 300 and 150 years appearing significantly (at 95% level). Occurrence of these periods in the δ^{18}O time series is not persistent; for instance, power for 300 year peak is clustered near 6500–7000 year. The two peaks corresponding to 600 year and 1000 year also appear strongly around 7000 years. Past atmospheric radiocarbon activity over the last 8000 years, derived from tree ring has already shown variabilities with periods that include: 940, 570, 500, 420, 360, 230 years, which are attributed to time varying solar output and internal dynamics of climate system[93]. Periods observed in radiocarbon activity variations are not seen distinctly in the global wavelet map of δ^{18}O (Figure 4) due to insufficient resolution of the periodogram. For example, in the δ^{18}O map, width of the base of the peak at ~300 year is between 200 and 500 year, which means periods corresponding to radiocarbon activity variations (viz. 500, 420, 360 and 230 years) may be part of it. Two points can be noted from the periods reported so far in Indian speleothem records: (i) most of them are not observed commonly and (ii) these are not persistent, which seems to be partly due to differences in data resolution. This also suggests that regional responses of the forcing mechanisms are complex in nature.

Perspectives for future research

There is high possibility of locating new caves near sites already studied so far; hence, attempts are required to locate new unexplored caves. A single stalagmite-based climate reconstruction may not be a genuine regional representation; proxy signal commonly observed in the analysis of more than one stalagmite from a cave will be a better depiction of the regional change. Despite large uncertainties in the actual estimated radiometric ages, age models are often developed where highly resolved climate changes are discussed. Therefore, rigorous statistical analysis of the available data, their validity and limitations need to be adopted. δ^{18}O variability near cave locations may be due to either amount effect or changes in vapour sources. A systematic study of isotopic variability in the modern drip water shall be monitored for different seasons to establish if the proxy variability can be faithfully used to infer past climate changes. Some of the rare speleothems, having significant detritus content, are not suitable for U–Th dating. Radiocarbon-based chronology shall be exploited in such cases to utilize and retrieve past environmental signals. Fluids trapped in speleothem matrix need to be studied to infer past isotopic composition of meteoric water. Also, organic carbon trapped in some of the speleoethems can be exploited to address palaeo vegetation.

Conclusion

Application of $\delta^{18}O$, $\delta^{13}C$ and trace elements in speleothems is discussed. $\delta^{18}O$ is being used primarily to address past monsoon variability. However, $\delta^{13}C$ and trace elements can indicate monsoon driven past environment changes. Several records addressing past monsoon from Indian caves are available. Most of the records are largely discontinuous in time coverage; high resolution data is available for many thousands of years, with the oldest record going back to ~280 ka. In order to get a continuous record of past monsoon variability, and hence to fill time gaps in the available record so far, more reconstructions are required in the future. Periodicities observed in the reconstructed data show non-persistent nature of monsoon variability; due to the complex nature of the climate system, so far, a clear attribution to the causative factor is not feasible.

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