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Trends and Oscillations in the Indian Summer Monsoon Rainfall over the Last Two Millennia

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Abstract

Observations show that summer rainfall over large parts of South Asia has declined over the past five to six decades. However, it remains unclear whether this trend is due to increased anthropogenic aerosol loading over South Asia or natural variability. Here we use stable oxygen isotopes in speleothems from northern India to reconstruct variations in Indian summer monsoon rainfall over the last two millennia. We find that within the long-term context of our record, the current drying trend is not outside the envelope of monsoon's oscillatory variability, albeit at the edge of this variance. Furthermore, the magnitude of multi-decadal oscillatory variability in monsoon rainfall inferred from our proxy record is comparable with model estimates of anthropogenic-forced trends of mean monsoon rainfall in the 21st century under various emission scenarios. Our results suggest that anthropogenic-forced changes in monsoon rainfall may be difficult to detect against a backdrop of large natural variability.

Among the most serious impacts of climate change to South Asia are those that relate to water availability and drought due to changes in the Indian summer monsoon (ISM) rainfall amount and spatio-temporal pattern. Although, the spatial structure of the mean seasonal rainfall over South Asia exhibits complex patterns¹, the area-weighted all India summer monsoon rainfall (AISMR) time series² serve as an adequate index of ISM intensity over continental India on interannual to decadal timescales³⁻⁴. Between AD 1870 and 1990, the AISMR appears to exhibit quasi-oscillatory multi-decadal (~60-70 years) variability with amplitudes that are generally within ± 5.0 % of historical normals (~ 850 mm)⁴⁻⁶. However, in context of the last 50-60 years, it reveals a statistically significant decreasing trend, which is also observed in various gridded precipitation datasets⁷⁻⁹. This decline stands in contrast to the near-unanimous results from climate models, which indicate increasing rainfall in a warming climate due to increasing greenhouse gas (GHG) concentrations¹⁰⁻¹¹. In some model simulations, the observed summertime drying trend has been attributed to increased anthropogenic aerosol loading over South Asia^{8-9,12} but considerable uncertainty remains in modeling the role of aerosols on monsoon rainfall¹³. A fundamental question therefore emerges is whether the weakening of ISM rainfall, particularly over northern India (NI), is anomalous within the long-term context of natural variability. Deciphering whether the recent decades-long ISM drying trend is caused by human-induced aerosol emissions or a manifestation of ISM natural variability is critical to understanding both the near and long-term behavior of the ISM.

Although, the AISMR is among the longest instrumental records of precipitation on Earth, it is still too short (~145 years) to allow a robust characterization of ISM multidecadal-centennial variability from which, one can examine the uniqueness of the observed drying trend. Existing proxy reconstructions from South Asia suggest that the ISM exhibits complex spatio-temporal variations, the finer details of which are highlighted by speleothem oxygen isotope ($\delta^{18}\text{O}$) records from central India (CI) and

northeastern India¹⁴⁻¹⁸. Building upon our previous work, here we present a new high-resolution and absolute-dated speleothem $\delta^{18}\text{O}$ -based reconstruction of ISM rainfall over the last two millennia. The new record is developed from two stalagmites (SAH-A and SAH-B) collected from Sahiya Cave in NI. The climatologically unique location of our site allows us to cast temporal changes in speleothem $\delta^{18}\text{O}$ in terms of decadal-averaged variations in ISM rainfall, and thus, the low frequency aspect of our record can be construed as a paleo-counterpart of the instrumental AISMR time series. Our reconstruction reveals a wide range of ISM behavior, consisting of multi-centennial trends forced by Northern Hemisphere temperatures, distinct multi-decadal oscillatory variability, and sporadic sub-decadal intervals of anomalously high and low rainfall. Within the long-term context of our record, the observed decline in ISM rainfall since the middle of the 20th century is not outside the envelope of natural variability, albeit at the edge of this variance. In addition, the magnitude of ISM multi-decadal oscillatory variability inferred from our record is comparable with model estimates of anthropogenic-forced trends of mean monsoon rainfall in the 21st century under various representative concentration pathway scenarios¹¹. Results from our study underscore the challenge of detecting anthropogenic-forced changes in ISM against a background of large internal variability.

Results

Regional climatology. Sahiya Cave ((30°36'N, 77°52'E, ~1190 m above sea level (masl)) is located ~200 km NNE of New Delhi in the Uttarakhand State of India (Fig. 1 and Supplementary Fig. 1). Mean annual temperature and annual precipitation in the region are ~22.0°C and 1600 mm, respectively, with ~80% of precipitation falling during the monsoon season (between June and September) (Supplementary Fig. 2). The onset of ISM rainfall in the study area is preceded by the development of the cross-equatorial low-level jet (LLJ) that serves as the primary conduit for transporting moisture across the Indian Ocean and the Arabian Sea onto the Indian sub-continent¹⁹. The LLJ generates strong cyclonic

vorticity in the planetary boundary layer (typically, north of 15°N) leading to large convective instabilities over the central portion of India, a region referred to as the Monsoon Trough (MT)²⁰. At the height of the ISM season (typically during July and August), the LLJ develops into a southeasterly flow, transporting moisture from the Bay of Bengal (BoB) to the northwest and northern India²¹. Additionally, the negative pressure anomalies in the MT region trigger the development, and propagation of synoptic-scale weather systems, collectively referred as the low-pressure systems (LPS)²⁰ from the BoB (typically north of 18°N during the months of July and August) onto the Indian landmass. The LPS are instrumental in advancing the ISM as far north as the Lesser Himalayas (i.e., north of our study area) and as far west as Pakistan during periods of strong monsoon circulation²¹. Another important source of precipitation variability in the study area stems from chaotic oscillations in monsoonal precipitation, referred to as active-break (A-B) spells²⁰⁻²². Distinct from the LPS, these oscillatory modes are manifestations of north-south fluctuations in the mean position of the MT. During the 'break' period, the axis of the MT shifts northwards of our site, resulting in dry air intrusions of westerly winds into northwest and central-north India, which tend to suppress rainfall through decrease of convective instability and depletion of parcel buoyancy²³⁻²⁴. In summary, the sub-seasonal to interannual precipitation variability at our study area, strategically located at the distal end of the BoB branch of the ISM and at the northern periphery of the MT region, is sensitive to changes in the strength and position of the LLJ and attendant changes in the MT, which in turn, influence the strength of the BoB branch of ISM.

Oxygen isotope in modern precipitation and monsoon circulation. We use the observational $\delta^{18}\text{O}_p$ data from the Global Network of Isotopes in Precipitation program of the International Atomic Energy Agency (GNIP-IAEA, hereafter GNIP) to characterize the $\delta^{18}\text{O}$ of precipitation ($\delta^{18}\text{O}_p$) at our study area. The GNIP data nearest to the cave site (Uttarkashi, 30.72N, 78.44E; 1140 masl) however, contains only three years of monthly $\delta^{18}\text{O}_p$ observations (AD 2004 - 2006; $n = 28$) (Supplementary Fig. 3).

Consequently, we use the $\delta^{18}\text{O}_p$ data from the GNIP station at New Delhi (28.58N, 77.52E; 212 masl, and ~200 km south of Sahiya) to characterize the $\delta^{18}\text{O}_p$ variability at our site (Supplementary Figs. 4-5). Our approach to utilize $\delta^{18}\text{O}_p$ data from New Delhi is justified by the gross similarities between the annual cycles of precipitation and $\delta^{18}\text{O}_p$ at Sahiya and New Delhi (see Methods), which suggest that both sites are influenced by similar precipitation dynamics and moisture sources throughout the year. Furthermore, in order to examine dynamical linkages between the $\delta^{18}\text{O}_p$ and changes in the low level (~850 hPa) monsoon circulation pattern, we use the Hybrid Single-Particle Lagrangian Integrated Trajectory Model (HYSPPLIT)²⁵ to generate backward (168 hours) moisture trajectory ensembles for those periods where GNIP data is available. Our paired analyses of GNIP $\delta^{18}\text{O}_p$ data and the low-level monsoon moisture trajectory patterns (Fig. 1 and Supplementary Fig. 6) suggest that the $\delta^{18}\text{O}_p$ variability at the study area is dominated by large-scale changes in low-level monsoon circulation and moisture transport history upstream of our site and to a lesser extent from changes in local precipitation intensity. We find that periods of strong (weak) monsoon circulation are characterized by an enhanced (reduced) flux of isotopically depleted BoB moisture and reduced (enhanced) flux of isotopically enriched Arabian Sea moisture to our site (see Methods).

Speleothem oxygen isotope interpretation. The contrasting patterns of strong and weak monsoon circulation produce distinct signatures in the $\delta^{18}\text{O}_p$, which in theory should predominantly drive changes in $\delta^{18}\text{O}$ of speleothems from the study area. This inference rests in part on the assumption that non-monsoon rainfall and snowmelt are insignificant sources of surface waters that infiltrate into the cave system. Indeed, this assumption is supported by the $\delta^{18}\text{O}$ values of the cave's dripwaters ($-6.95 \pm 0.31\text{‰}$, $n = 7$), which are indistinguishable from the amount-weighted $\delta^{18}\text{O}$ of annual precipitation (-6.97‰) and are slightly higher than the $\delta^{18}\text{O}$ of JJAS-integrated precipitation (-8.19‰) in the study region (Supplementary Fig. 3d). Furthermore, the regional precipitation-evaporation (P-E)

budget of the region is positive only during the summer monsoon season (from June to September) (Supplementary Fig. 2b), implying that effective infiltration of surface waters into the local karst occurs primarily during the ISM season (Supplementary Fig. 3). The large karstic overburden over Sahiya Cave (~30 m) would however, tend to act as a lowpass filter by mixing $\delta^{18}\text{O}_p$ signals from several seasons, thus damping high frequency $\delta^{18}\text{O}_p$ variations. Nonetheless, the low frequency (multi-year to decadal) component of the speleothem $\delta^{18}\text{O}$ record should reflect time-integrated shifts in $\delta^{18}\text{O}$ of karstic waters stemming from an asymmetric occurrence of ^{18}O enriched (depleted) precipitation events during weak (strong) monsoon seasons. Consequently, we interpret temporal changes in the speleothem $\delta^{18}\text{O}$ at our site to reflect space-time integrated upstream changes in ISM intensity. Indeed, comparisons of the speleothem $\delta^{18}\text{O}$ time series with instrumental records of precipitation at local, sub-regional, and sub-continental spatial scales (Supplementary Fig. 7) indicate that the strength of inverse $\delta^{18}\text{O}_p$ -rainfall amount relationship is higher at sub-regional and sub-continental scales. This observation suggests that upstream changes in ISM circulation and rainfall variability rather than changes in the local precipitation amount are the primary drivers of $\delta^{18}\text{O}$ variability at our site. Our observations are consistent with the results from other studies, which suggest that local $\delta^{18}\text{O}_p$ variability at some monsoonal and tropical regions reflect upstream changes in regional circulation and rainfall²⁶⁻²⁷.

Speleothem stable isotope record and age model. The Sahiya Cave $\delta^{18}\text{O}$ record (SAH) is a composite of two $\delta^{18}\text{O}$ stalagmite records (SAH-A and SAH-B) consisting of 1848 $\delta^{18}\text{O}$ measurements with an average temporal resolution of 1.5 years per sample (Supplementary Fig. 8). The chronologic frameworks of SAH-A and SAH-B $\delta^{18}\text{O}$ profiles are provided by 13 and 15 ^{230}Th dates, respectively (Supplementary Fig. 9 and Supplementary Table 1). The age models are developed by generating multiple cubic interpolations between the discrete ^{230}Th dates²⁸. The SAH $\delta^{18}\text{O}$ record (mean = -9.0‰; $1\sigma = 0.24$) covers the last two thousand years (BC 141 to AD 2006). The $\delta^{18}\text{O}$ profile is characterized by

prominent decadal to multi-decadal variability with amplitude of $\sim 0.5\text{‰}$ around the long-term mean (Fig. 2b). The observed weakening trend in ISM rainfall since the 1950s manifests as an increase of $\sim 0.5\text{‰}$ in speleothem $\delta^{18}\text{O}$ values, which is a robust feature of our record even with generous assumptions of age model and analytical uncertainties. Comparison between the composite $\delta^{18}\text{O}$ profile and an 11-year running mean of AISMR, exhibits a significant inverse correlation ($r = -0.75$ with 95% confidence interval $[-0.82; -0.64]$) that explains $\sim 50\%$ of variance in the instrument record (Fig. 2a, Supplementary Fig. 7). After correcting for standard instrumental error in $\delta^{18}\text{O}$ measurements (0.06‰), the slope of the speleothem $\delta^{18}\text{O}$ -AISMR regression yields a conversion of 1.0‰ in speleothem $\delta^{18}\text{O}$ values to $\sim 86\text{ mm}$ change in AISMR, the latter equivalent to $\sim 10\%$ of the long-term mean ($\sim 850\text{ mm}$) (Fig. 2a).

The $\delta^{18}\text{O}$ record is characterized by several episodic intervals of anomalously ($>2\sigma$) positive and negative $\delta^{18}\text{O}$ values. Particularly noteworthy are two sub-decadal length intervals of strong pluvial and drier conditions, centered at AD ~ 1100 and 1620 , respectively (Fig. 2b). The timing and duration of the latter event, which constitutes the largest positive $\delta^{18}\text{O}$ anomalies in our reconstruction coincides (within age uncertainties) with widespread droughts and famines in north India²⁹ and with the collapse of the Guge Kingdom in western Tibet³⁰. Incidentally, these two intervals of substantially enhanced and reduced ISM rainfall occur within the periods of peak thermal anomalies during the Medieval Climate Anomaly (MCA) (nominally, AD 1000-1200) and the Little Ice Age (LIA) (nominally AD ~ 1600 -1850) (Fig. 2), respectively.

Spectral analysis³¹ of the record reveals significant power in the 60-80 and 15-30 year bands (Fig. 3). The time-frequency³² and wavelet analyses³³ of the speleothem $\delta^{18}\text{O}$ record (Supplementary Fig. 10) suggest that the amplitudes of aforementioned scale-averaged oscillations are non-stationary. For example, the 60-80 year oscillatory band is dominant during the AD 300-500, 1500-1700, and 1850-

2000 intervals and nearly absent during other times. Significant concentration of variance in 60-80 year band is also common in numerous instrumental and proxy global and regional SST records³⁴. The nonstationary aspect of the ISM multidecadal oscillatory mode likely arises from the influence of specific (and time-space varying) arrays of different modes of SST variability, which themselves are coupled in complex manners³⁵. On multi-centennial and longer time scales, the $\delta^{18}\text{O}$ profile over the last 1000 years is characterized by a pronounced trend towards higher values.

Comparison with central India record. The Sahiya Cave $\delta^{18}\text{O}$ record from NI shares key similarities and notable differences with the speleothem $\delta^{18}\text{O}$ record from CI¹⁴⁻¹⁶ (19°N and 82°E) (Fig. 3b and Supplementary Fig. 11). The latter has been shown to reflect the ISM rainfall variability over the core monsoon zone of India (defined here as ~18°–27°N and 69°–88°E). Importantly, both records exhibit strong multidecadal variability and share common spectral power (Fig. 3b). Although the CI $\delta^{18}\text{O}$ record has slightly larger amplitude and is more variable than the NI $\delta^{18}\text{O}$ record (Fig. 3b), such differences may result, at least partially, from varying degree of smoothing of isotopic signal imparted by processes unique to local karst at each location. The visual comparison between the two records shows similar long-term patterns but no consistent peak-to-peak relationship on decadal-multidecadal timescales. For example, the most prominent positive anomaly in CI record at ~1370 has a muted presence in the NI record whereas the most positive $\delta^{18}\text{O}$ anomaly in the latter ~1600 AD has a clear counterpart in the CI record. While such inconsistencies may result from spatial contrast in ISM rainfall, the combined age uncertainties of both records preclude definitive assessment of phase relationships between these records on short-time scales.

Discussion

The low frequency (multi-centennial) aspect of our ISM reconstruction from NI is broadly similar to Northern temperature reconstructions³⁶ over the last two millennia (Fig. 2b). The close correspondence between these reconstructions on such time scale supports the notion that NH

temperature is an important driver of ISM variability via changes in meridional surface /tropospheric thermal gradients^{12,37} and interhemispheric thermal contrast^{6,37}. However, neither proxy nor instrumental data comparisons between ISM rainfall and NH temperatures reveal consistent relationships on decadal to multi-decadal timescales. This suggests that the thermodynamic effect on ISM rainfall maybe more important on longer timescales via changes in boundary conditions at larger spatial-scale (global or hemispherical)³⁷. At shorter timescales, the ISM is more strongly influenced by changes in local and/or remote sea surface temperature (SST) fields and other aspects of external boundary conditions⁴⁻⁶. Nonetheless, given that the magnitude of NH temperature rise since the 1950s is unprecedented in the context of the last 2000 years, the decreasing trend in the ISM rainfall seems anomalous in light of changes expected from simple thermodynamical-scaling argument³⁸. The latter postulates that atmospheric water vapor should increase in line with the Clausius Clapeyron relationship at $\sim 7\%^{\circ}\text{C}^{-1}$ rise in atmospheric temperature due to increasing concentrations of GHGs. The lack of an upward trend in ISM rainfall since the 1950s supports results from coupled climate models that include aerosol forcing^{8-9,12}. These model experiments indicate that the increased anthropogenic aerosol loading over south Asia may have substantially counterbalanced a precipitation increase over continental India that would otherwise have occurred in response to singular increase in global temperatures.

Parsing the relative contribution of aerosol forcing to the observed decline in ISM rainfall requires establishing an envelope of ISM baseline natural variability on decadal to multi-decadal timescales. We use the observed inverse speleothem $\delta^{18}\text{O}$ -AISMR relationship in deriving, admittedly raw, quantitative estimates of decadal-scale variations in ISM. Using the standard linear regression fitted by means of ordinary least square method, we use the slope of speleothem $\delta^{18}\text{O}$ -AISMR relationship ($\sim 1.0\text{‰}/86\text{ mm}$) to estimate decadal-scale ISM rainfall variability over the last 2000 years. Although more sophisticated transfer functions employing a forward modeling approach may ultimately be more

robust, any model for which the $\delta^{18}\text{O}$ of monsoon precipitation (and consequently, the $\delta^{18}\text{O}$ of speleothem) is inversely linked to area-weighted changes in ISM rainfall, will lead us, at least qualitatively, to the conclusion we reach below. Our analysis shows that the magnitude of ISM rainfall decrease since the mid-20th century is not unique in the long-term context of our record, but nonetheless near the edge of the envelope of ISM long-term variability (Fig. 3a). These observations imply that the observed drying can be due to ISM natural variability alone. However, it will be imprudent to ignore the possible influence of aerosol forcing on ISM rainfall, especially considering that previous excursions to comparably drier intervals inferred from our record, typically occurred in the context of cooler NH temperatures and in absence of significant human-induced aerosols emissions. It is also instructive to compare the maximum expression of multi-decadal peak-to-trough ISM variability in our record, for example from AD 1720 to ~1800 (~10%) (Fig. 3b) with anthropogenic-forced trends of mean monsoon rainfall projected from climate models for the 21st century. A recent evaluation of ISM rainfall for the period 1850 to 2100 from 20 climate models within the Coupled Model Intercomparison Project 5 (CMIP-5) showed¹¹ a consistent increase in mean monsoon rainfall for the period AD 2070-2100 (relative to the historic period AD 1870-1900) under different representative concentration pathways (RCP) scenarios, with the highest increase in the strongest warming scenario RCP 8.5. However, only half of these models can realistically simulate ISM¹¹. The projected increase in mean ISM rainfall for the period AD 2070-2100 (relative to AD 1870-1900) from all but one of the realistic models is within 10% for all warming scenarios, which is similar to the maximum magnitude of ISM natural variability inferred from our record. These observations highlight the fundamental problem of detecting forced trends in ISM rainfall against the background of relatively large internal variability.

Understanding the current and future changes in ISM rainfall spatio-temporal pattern and amount has enormous societal implications for more than 1.5 billion people in the region. The paleoclimate data

presented here provides a baseline of ISM natural variability on multi-decadal to centennial timescales. It is on precisely these timescales that the anthropogenic-forced changes are likely to be superimposed on monsoon's natural climate variability. Detection and attribution of climate change on ISM rainfall will require a coordinated effort through improvements in both climate models and proxy reconstructions of ISM rainfall. The paper presented here highlights the promise of this approach and provides impetus for further refinement of ISM rainfall reconstructions.

Methods

Cave characteristics and sample collection. In April 2007, we collected a pair of columnar-shaped stalagmites (SAH-A and SAH-B) from a small and poorly ventilated chamber, ~100 m from the cave's entrance. The stalagmite SAH-A was found wet (and presumably actively growing) at the time of collection. In-situ measurements indicate relative humidity (rH) in the excess of 95%, numerous active feeder drips and a narrow range of ambient temperatures ($T_{\text{cave air}} = 21.8 \pm 0.6^\circ\text{C}$, $n = 7$) throughout the cave, with the latter being similar to the regional mean annual air temperature (MAT) ($\sim 22.0^\circ\text{C}$) (Supplementary Fig. 2). Unlike higher altitude regions to its north, the study area (elevation ~1000-1200 m) does not experience snowfall. Consequently, spring snowmelt is not a source of meteoric water into the local karst above the Sahiya Cave. The regional precipitation-evaporation is positive only during the summer monsoon season (Supplementary Fig. 2b); suggesting that effective infiltration of meteoric waters into the local karst is only significant during the ISM season.

Sample Preparation. The stalagmites were cut with a thin diamond blade along the growth axes. A digitally controlled tri-axial micromill was used to obtain sub-samples for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ analyses by continuously milling in 100 μm increments along the stalagmites growth axes, yielding sub-samples weighing typically between 100-130 μg for isotopic analyses. Sub-samples for ^{230}Th dating (weighing typically between 150 to 300 mg) were obtained by drilling with a 0.9 mm carbide dental burr.

Stable isotope analysis. Sub-samples for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ analyses were measured using a Multi-Prep Dual Inlet System attached to an Iso-prime Stable Isotope Ratio Mass Spectrometer in the University of Southern California. The long-term precision of the Ultissima standard for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ measurements over the course of this study was $\sim 0.06\text{‰}$ and 0.03‰ , respectively.

Isotopic equilibrium. Two lines of evidences support that the calcite deposition in Sahiya cave occurs at or near isotopic equilibrium. First, strong replication between the SAH-A and SAH-B $\delta^{18}\text{O}$ records during the period of overlap ($r = 0.60$ with 95% confidence interval [0.17, 0.84]) (Supplementary Fig. 8) suggest that both stalagmites have recorded a primary climate signal that has not been significantly altered by disequilibrium processes associated with conditions unique to each stalagmite, such as different flow pathways through the overlying bedrock, groundwater mixing and residence time, $p\text{CO}_2$, etc. Secondly, the theoretical range of the caves' ambient temperature (20.8 to 23.9°C) under isotopic equilibrium conditions⁴⁵ calculated from the measured range in $\delta^{18}\text{O}$ of dripwaters ($-6.95 \pm 0.31\text{‰}$, $n = 7$) and the $\delta^{18}\text{O}$ of modern calcite (-8.53‰) is well within the measured range of ambient temperature in Sahiya cave ($21.8 \pm 0.6^\circ\text{C}$, $n = 7$).

^{230}Th dating and age model. ^{230}Th dating was performed at the Minnesota Isotope Laboratory on a Multi-Collector Inductively Coupled Plasma Mass Spectrometer (MC-ICPMS, Thermo-Finnigan Neptune). We use standard chemical procedures to separate uranium and thorium for ^{230}Th dating as reported in Edwards *et al.*⁴⁰. The standardization and half-lives are reported in Cheng *et al.*⁴¹. All U and Th isotopes were measured on a MasCom multiplier behind the retarding potential quadrupole in peak-jumping mode. We followed similar procedures of characterizing the multiplier as described in Shen *et al.*⁴² Initial ^{230}Th values were corrected and all ages are reported with 2σ error and are in correct stratigraphic order (Supplementary Table 1). The chronologic frameworks of stalagmite SAH-A and SAH-B $\delta^{18}\text{O}$ profiles are constrained by 13 and 15 ^{230}Th dates, respectively. We use Monte Carlo simulations ($N = 2000$) to account for the dating uncertainties (2.5% and 97.5% quantile confidence limits) within the time series²⁸ (Supplementary Fig 9 and Supplementary Table 1).

Statistical analyses. Confidence intervals of correlation coefficients were determined by using pairwise moving-block bootstrap method to preserve the serial dependence of time series and interval calibration to increase the accuracy⁴³. To identify periodic components in the spectrum of unevenly spaced SAH-A, SAH-B, and composite SAH $\delta^{18}\text{O}$ time series, we applied the Lomb–Scargle Fourier Transform in combination with Welch's Overlapped Sequence Averaging (WOSA) procedure by using the REDFIT software³¹. REDFIT produces first-order autoregressive (AR1) time series with sampling times and characteristic timescales matching those of the input climate data. A varying number of overlapping (50%) detrended segments were used to optimize bias/variance properties of spectrum estimates. The univariate spectra were bias-corrected using 1,000 Monte Carlo simulations. To assess the statistical significance of a spectral peak, REDFIT estimates the upper confidence interval of the AR1 noise for various significance levels based on a χ^2 distribution. Temporal changes in the average amplitude of signal components were estimated using a modified harmonic-filtering algorithm³², which fits sinusoidal waves to an unevenly spaced series by means of least-squares (analogous to the Lomb-Scargle Fourier Transform). We used a sliding window of length four times the signal period and ascribe to the window's midpoint the average signal amplitude; the window is then shifted in steps of 0.1 yr. This leads to a truncation of the time interval for the resulting estimate of the average amplitude of $4 \times 80/2 \text{ yr.} = 160 \text{ yr.}$, from both the lower and upper time interval bound. In addition, we also applied the conventional continuous transform wavelet analysis to annually interpolated $\delta^{18}\text{O}$ data being mindful of the caveat that here interpolation might introduce artifacts, potentially biasing the derived spectra. Singular spectrum analysis-Multitaper Method (SSA-MTM) Toolkit⁴⁴ was used to remove nonstationary linear trends from the $\delta^{18}\text{O}$ time series by examining the variability of the largest eigenvalue from an autocovariance matrix. SSA decomposes timeseries into several modes of variability (monotonic and oscillatory) and the remaining low-amplitude noise. Time series trends and associated break points (timing of trends change) are quantified using a parametric, nonlinear regression technique⁴². This model is a continuous function consisting of two linear parts that are joined at the break point. The break model is fitted to data using a weighted least-squares method with a brute-force search for the break point. Statistical uncertainties of the trend and break point are evaluated using 2000 block bootstrap simulations, which preserve the distribution and serial dependence of the data over the length of a block.

Statistical analysis of observational $\delta^{18}\text{O}_p$ data. We extensively utilized observational monthly $\delta^{18}\text{O}_p$ data from two sites— Uttarkashi³⁹ and New Delhi (both dataset are accessible from www.naweb.iaea.org) to characterize the $\delta^{18}\text{O}_p$ variability and understand the dynamical controls of $\delta^{18}\text{O}_p$ –precipitation ($\delta^{18}\text{O}_p$ –P) relationship at our study area. The $\delta^{18}\text{O}_p$ data from Uttarkashi (~50 km from Sahiya Cave) contains only three years of monthly $\delta^{18}\text{O}_p$ observations (AD 2004 - 2006; $n = 28$),

whereas the GNIP station at New Delhi (~200 km south of Sahiya Cave), contains 341 monthly $\delta^{18}\text{O}_p$ observations (available with gaps from AD 1961 to 2008). Despite the large difference in number of observations, the annual cycles of the $\delta^{18}\text{O}_p$ and rainfall amount from both sites are remarkably similar (Supplementary Fig. 2a-b). This close correspondence is not surprising considering that both sites are under the influence of similar precipitation dynamics due to their proximity. Additionally, we find strong correlation between the observational $\delta^{18}\text{O}_p$ data from New Delhi and simulated $\delta^{18}\text{O}_p$ data (extracted from the grid point nearest to Sahiya Cave) from Experimental Climate Prediction Center's Global Spectral Model (ECPC-GSM) fitted with isotope tracers (IsoGSM)⁴⁷ (Supplementary Fig. 3). In light of these observations, we argue that the GNIP $\delta^{18}\text{O}_p$ database from New Delhi data from can be reliably used to understand the nature of $\delta^{18}\text{O}_p$ -P relationship at our study area.

The New Delhi GNIP database is characterized by a large range in monthly $\delta^{18}\text{O}_p$ and precipitation amount (Supplementary Fig. 4). The distribution of JJAS-averaged $\delta^{18}\text{O}_p$ and precipitation amount illustrates a negative $\delta^{18}\text{O}_p$ -P relationships ($r = -0.37$, with 95% confidence interval $[-0.62; -0.04]$, $n = 35$). The inverse $\delta^{18}\text{O}_p$ -P relationship for the individual JJAS months (Supplementary Fig. 5) is most pronounced at the onset (June) and end (September) of the ISM season. In contrast, the inverse $\delta^{18}\text{O}_p$ -P relationship during the height of the monsoon season (JA), is relatively weaker despite July and August being typically the wettest months. One plausible explanation for this observation is that lower humidity, higher evaporation, and unsaturated soil conditions in June and September lead to faster re-evaporation of precipitation below cloud base and from the ground, thus explaining the observed inverse $\delta^{18}\text{O}_p$ -P relationship through the classic *amount effect*. However, a critical examination of the GNIP dataset, together with composite trajectory analyses of low level monsoon winds during strong and weak monsoon seasons, suggest that sub-seasonal changes in monsoon circulation and associated changes in the moisture source and rainout history of air parcels, are more dominant contributors to the observed inverse $\delta^{18}\text{O}_p$ -P relationship at our study area. This can be explained as follows. The monthly observations of $\delta^{18}\text{O}_p$ and precipitation amount in the GNIP data exhibit a large range with some values representing extreme ends of both, $\delta^{18}\text{O}_p$ and rainfall amount ranges. We define the extreme values as those that fall in the lower and the upper quartiles of observed range of $\delta^{18}\text{O}_p$ and precipitation amount (Supplementary Fig. 5). Based on this scheme, we classify these extreme $\delta^{18}\text{O}_p$ -P pairs (and their relative abundance in the GNIP database) into four groups: (a) negative-wet (NW, relatively common); (b) negative-dry (ND, rare); (c) positive-wet (PW, rare) and (d) positive-dry (PD, relatively common). We find that the presence or absence of the inverse $\delta^{18}\text{O}_p$ -P relationship is strongly sensitive to the abundance of the NW and PD pairs in the dataset; and if these data pairs are removed from the regression analysis, the inverse $\delta^{18}\text{O}_p$ -P relationship during June weakens and vanishes for the remaining summer monsoon season. These observations suggest that the primary source of inverse $\delta^{18}\text{O}_p$ -P relationship in the GNIP database arises from occurrence of anomalously strong (^{18}O depleted) and weak (^{18}O enriched) precipitation events.

HYSPLIT trajectory analyses. We utilized the HYSPLIT²⁵ model provided by the National Oceanographic and Atmospheric Administration Air Resources Laboratory (NOAA -ARL) together with the National Center for Environmental Prediction (NCEP) and National Center for Atmospheric Research Reanalysis (NCAR) version 1 data⁴⁶ to generate quasi-isentropic backward (168 hours) trajectory ensembles of strong and weak monsoon seasons (Fig. 1 and Supplementary Fig. 6). We are cognizant that HYSPLIT-derived trajectories can have large computational errors that are inherited from uncertainties in the reanalysis products. Therefore, we increase our confidence in trajectory analysis by computing hundreds of trajectories over an extended period to identify major moisture source areas and broad circulation features that produce precipitation in our study area. For each month of the monsoon season (JJAS), we produce the 'wet' and 'dry' trajectory ensembles

(Supplementary Fig. 6). The ‘wet’ and ‘dry’ months are defined here as those that fall in the upper and lower quartile of the observed range in precipitation amount in the GNIP database. For each wet and dry composite, we computed ~600 isentropic back-trajectories (168 hours)(sampled four times daily at UTC 00, 06, 12 and 18) initialized at 1500 meters above ground level corresponding to ~850 hPa). Results from a smaller sub-set of randomly selected trajectories initialized at 1.0 or 1.5 km above ground level (corresponding to ~900 and 850 hPa levels, respectively) and computed for either 120 or 168 hours duration, were nearly identical.

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Author contributions

A.S. H.C. and G.K. conceived the project and A.S. wrote the first draft of the paper. H.C. R.L.E. and G.K. contributed to ^{230}Th dating. S.F.M.B. and M.M. performed the age modeling and statistical analysis. All authors discussed the results, edited and commented on the manuscript.

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Figure 1: Cave location and ensembles of backward (~168 hours) air-parcel trajectories originating from the study area. The trajectory ensembles depicting air-parcel/moisture transport routes, highlighting the maximum contrast between strong and weak monsoon circulation. The green and black lines in each panel are multi-year monthly composites of daily trajectories originating from the study area (red dot) associated with periods of anomalously high and low rainfall, respectively during June (a), July (b), August (c), and September (d). Anomalously low rainfall (relative to the 20th century mean) during June, July, August, and September at our study area is characterized by near-cessation of ¹⁸O-depleted moisture flux from the Bay of Bengal, resulting in relatively higher flux of ¹⁸O-enriched moisture from the Arabian Sea and dry air intrusions of westerly winds that tend to suppress rainfall over central and northern India through decrease of convective instability. On other hand, anomalously high rainfall, particularly during July and August, is associated with significantly enhanced transport of ¹⁸O-depleted moisture from the Bay of Bengal (see Methods and Supplementary Fig. 6).

Figure 2: Comparison of speleothem $\delta^{18}\text{O}$ record with instrumental and proxy data. (a) The comparison between the modern portion of the speleothem $\delta^{18}\text{O}$ anomaly (up on the y-axis denotes stronger ISM intensity) and AISMR (JJAS) precipitation anomaly (11 yr. running mean) expressed as % departure from the 20th century mean (~850 mm) (shaded). (b) The NI speleothem $\delta^{18}\text{O}$ record (blue) along with the regressed AISMR anomaly (%) (dashed lines) and NH temperature anomalies reconstruction³⁶ (red). Bold lines are estimated trends⁴³ in the time series over the last 1000 years. White circles mark the break points (i.e., timing of trends change).

Figure 3: Time series analysis of Indian speleothem $\delta^{18}\text{O}$ records. (a) The residual $\delta^{18}\text{O}$ anomaly (gray) and 11 year running mean (black) of the NI speleothem $\delta^{18}\text{O}$ record along with the regressed AISMR anomaly (%). (b) The residual speleothem $\delta^{18}\text{O}$ anomaly time series of NI (black) and CI (red)

obtained after removing the long-term nonstationary trends by subtracting the first reconstructed component indicated by singular spectrum analysis⁴⁴ from the raw data. (c) Power spectrum of the composite NI (left) and (d) CI (right) SSA-detrended $\delta^{18}\text{O}$ time series obtained by using REDFIT³¹ software. A varying number of Welch overlaps were used to optimize bias/variance properties. Spectral band significant above the 90% level are labeled with their period.