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Key Points:

- Longest record of summer precipitation from speleothem from India
- Abrupt changes in summer monsoon precipitation
- Less precipitation during Younger Dryas and Heinrich events

Supporting Information:

- Supporting Information S1
- Table S1

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Abrupt changes in Indian summer monsoon strength during 33,800 to 5500 years B.P.

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Abstract Speleothem proxy records from northeastern (NE) India reflect seasonal changes in Indian summer monsoon strength as well as moisture source and transport paths. We have analyzed a new speleothem record from Mawmluh Cave, Meghalaya, India, in order to better understand these processes. The data show a strong wet phase 33,500–32,500 years B.P. followed by a weak/dry phase from 26,000 to 23,500 years B.P. and a very weak phase from 17,000 to 15,000 years B.P. The record suggests abrupt increase in strength during the Bølling-Allerød and early Holocene periods and pronounced weakening during the Heinrich and Younger Dryas cold events. We infer that these changes in monsoon strength are driven by changes in temperature gradients which drive changes in winds and moisture transport into northeast India.

1. Introduction

The sudden arrival of the rainy season after a period of hot and dry summer marks the onset of the summer monsoon in India [Chao, 2000]. About 80% of India's total annual rainfall occurs during summer or southwest (SW) monsoon season (June–September) with the exception of a few areas in the northwest and extreme south [Parthasarathy, 1960]. Historically, Indian summer monsoon (ISM) failures led to the demise and/or displacement of numerous cultures and civilizations in South Asia [Staubwasser *et al.*, 2003; Buckley *et al.*, 2010; Dixit *et al.*, 2014]. Extreme future changes in Indian summer monsoon (ISM) rainfall may have adverse impacts as well [Intergovernmental Panel on Climate Change (IPCC), 2013].

Climate models predict increases in extreme and heavy precipitation events in a warm world [Trenberth *et al.*, 2003; Allan and Soden, 2008; IPCC, 2013]. These studies indicate a direct relation between rainfall extremes and temperature, with intense rain events increasing during warm intervals and decreasing during cold intervals. It is also anticipated that as temperatures rise, the hydrological cycle will experience a shift in precipitation form from snow to rain and early snowpack melting [Pavelsky *et al.*, 2012]. Understanding changes in the intensity and frequency of these precipitation extremes is important to the welfare of human societies and the ecosystems on which they depend.

Model and observational studies indicate that the Asian summer monsoon weakened and the westerlies strengthened significantly during the Younger Dryas (YD) and the Last Glacial Maximum (LGM) [Porter and Zhisheng, 1995; Sinha *et al.*, 2005]. The East Asian summer monsoon precipitation also decreased significantly during the YD [Wang *et al.*, 2001; Yuan *et al.*, 2004]. Similarly, ISM strength decreased during Holocene North Atlantic cold intervals; the so-called Bond events [Gupta *et al.*, 2003]. These studies demonstrate a link between North Atlantic temperature changes and Asian monsoon strength. Model experiments show that the westerlies transmitted these North Atlantic anomalies to the Asian region [Sun *et al.*, 2011].

Late Pleistocene speleothem records from the Hulu and Dongge Caves of China indicate centennial- to millennial-scale changes in East and South Asian monsoon precipitation [Wang *et al.*, 2001; Yuan *et al.*, 2004]. However, such high-resolution, long-ranging ISM precipitation records are scarce from the Indian landmass especially from the Indian Himalaya [Sinha *et al.*, 2005]. Oxygen isotope records in speleothems (cave carbonates) across Asia are commonly interpreted as a proxy for the strength of the Asian monsoons [Wang *et al.*, 2001; Yuan *et al.*, 2004; Sinha *et al.*, 2005]. Recent studies, however, suggest that the cave records from

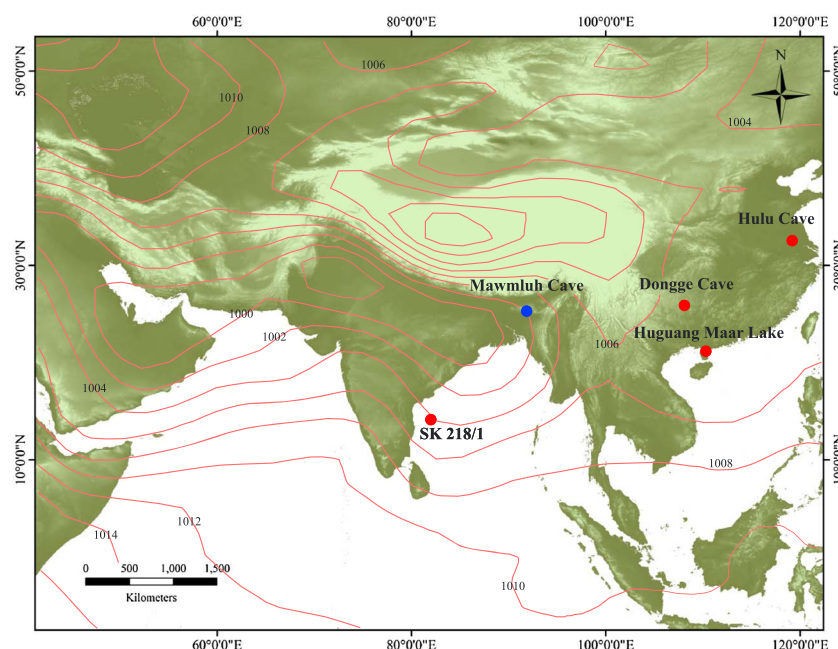


Figure 1. Location of the Mawmluh Cave, Meghalaya, NE India, superimposed with July Surface Pressure lines based on International Research Institute for Climate and Society, Earth Institute, Columbia University (source: http://iridl.ldeo.columbia.edu/maproom/Global/Climatologies/Vector_Winds.html). Also shown are location of other caves and lake from China and a marine core from the Bay of Bengal whose records have been compared with that from Mawmluh Cave.

Asia also reflect changes in moisture source, changes in fractionation at the source, and changes in transport pathways as well as the isotopic composition of Bay of Bengal (BoB) surface water [LeGrande and Schmidt, 2009; Breitenbach et al., 2010; Dayem et al., 2010; Pausata et al., 2011; Berkelhammer et al., 2012].

The stable oxygen isotope ($\delta^{18}\text{O}$) record of speleothems from Mawmluh Cave, NE India, provides information about changes in ISM strength on decadal to centennial timescales during the past few millennia. Mawmluh Cave (25°15'44"N, 91°52'54"E altitude 1290 m) is in a region of NE India near Cherrapunji which is swept by strong SW monsoon winds (Figure 1 and supporting information Figures S1–S3). Cherrapunji is among the wettest locations on Earth with an annual average precipitation of >11 m, 70% of which is received during the summer monsoon months from June to September, during the active SW monsoon season [Murata et al., 2007].

The weighted mean $\delta^{18}\text{O}$ versus precipitation and $\delta^{18}\text{O}$ versus temperature relationships for Shillong GNIP site near the cave indicate that precipitation amount accounts for ~30% of the variance in modern $\delta^{18}\text{O}$ of precipitation while temperature shows no relation to the $\delta^{18}\text{O}$ of precipitation (supporting information Figure S4) (source: http://www.univie.ac.at/cartography/project/wiser/gui/gnip_all_index.php). These modern relationships suggest that changes along the transport path and in the isotopic composition of source surface water account for the majority of the modern variance in the local rainfall $\delta^{18}\text{O}$ [Breitenbach et al., 2010; Liu et al., 2014] (Figure S4). Therefore, we relate changes in speleothem $\delta^{18}\text{O}$ to changes in isotopic composition of source waters and processes (evaporation and precipitation) occurring along the transport path; all of which, in turn, reflect changes in ISM strength [Yuan et al., 2004; Pausata et al., 2011].

To assess long-term changes in ISM strength and to understand whether LGM, YD, and Heinrich cold events were accompanied by decreased summer monsoon strength in the Indian subcontinent, we studied cave carbonate $\delta^{18}\text{O}$ from the Mawmluh Cave, near Cherrapunji, Meghalaya (NE India) spanning in age from 33,800 to 5500 years B.P. (Figure 1). Our main objective is to evaluate the climatological significance of the abrupt changes recorded in this $\delta^{18}\text{O}$ time series.

2. Methods

Samples were taken at 0.5 mm interval along the central growth axis of the speleothem sample with average sampling intervals varying from 15 to 77 years. The $\delta^{18}\text{O}$ profile is based on 367 isotopic measurements

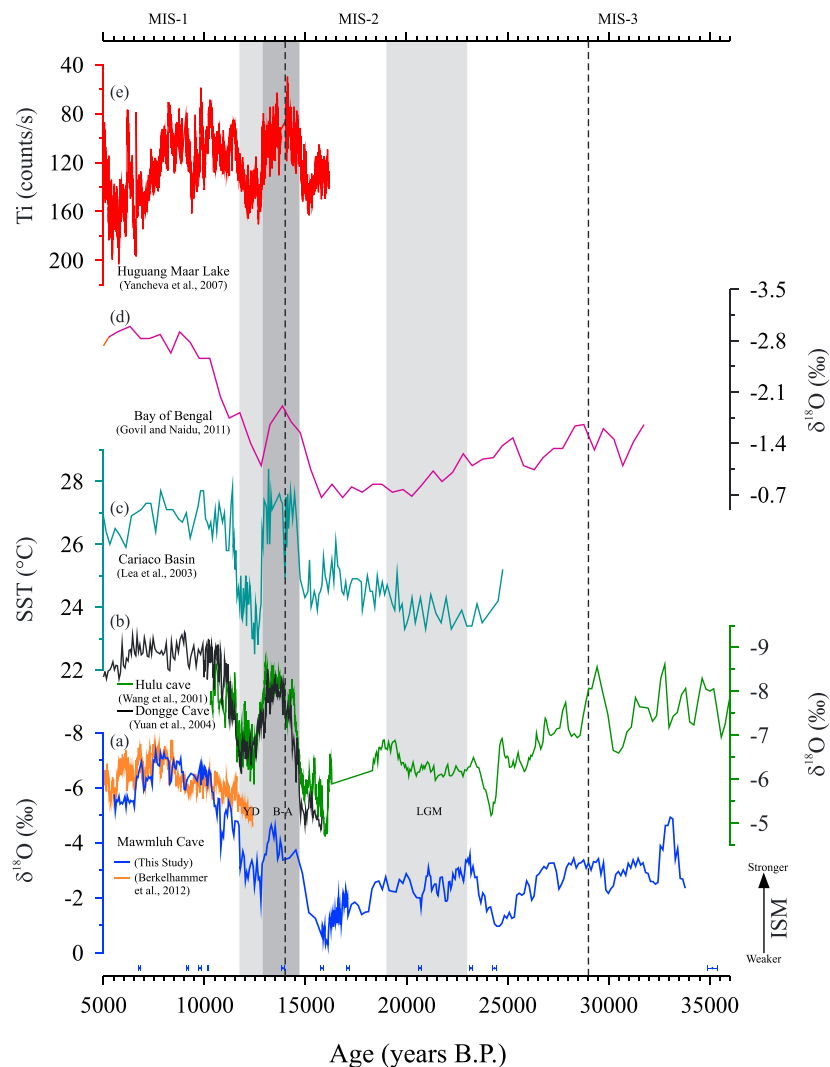


Figure 2. Indian summer monsoon (ISM) proxy record from the Mawmluh Cave, Meghalaya, India, compared with cave and lake records from China, and marine records from the Bay of Bengal and Cariaco Basin. (a) Mawmluh Cave $\delta^{18}\text{O}$ record for the interval 33,800 to 5500 years B.P. (dated intervals are marked by diamonds with error bars in blue color); also superimposed is the published $\delta^{18}\text{O}$ record from the Mawmluh Cave [Berkelhammer *et al.*, 2012]. (b) Dongge Cave $\delta^{18}\text{O}$ record (black color) [Yuan *et al.*, 2004] combined with Hulu Cave $\delta^{18}\text{O}$ record (green color) [Wang *et al.*, 2001]. (c) SST values from Cariaco Basin core PL07-39PC (cyan color) [Lea *et al.*, 2003]. (d) Bay of Bengal planktic foraminifer $\delta^{18}\text{O}$ record as proxy for precipitation change [Govil and Naidu, 2011]. (e) Huguang Maar Lake record of Ti counts/s [Yancheva *et al.*, 2007]. Broken vertical black lines mark the boundaries between marine isotopic stages (MIS). Light grey bars mark the Younger Dryas (YD) period and Last Glacial Maximum (LGM), whereas the dark grey bar indicates the Bølling-Allerød (B-A) period.

carried out at Brown University, USA (supporting information Table S1). Replicate analysis of Carrara marble, treated in the same as the speleothem samples, indicates reproducibility of ± 0.07 (1σ , $N = 26$). Hendy test [Hendy, 1971] of one layer suggests no significant correlation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ (supporting information Figure S5). Comparison of $\delta^{18}\text{O}$ values with the earlier Mawmluh Cave record [Berkelhammer *et al.*, 2012] from 12,000 to 3650 years eliminates the chances of kinetic fractionation and validates our data (Figure 2).

The chronology of Mawmluh Stalagmite-1 (MWS-1) was constrained by 11 absolute U-Th series dates determined by multiple collector-inductively coupled plasma-mass spectrometry at the University of Minnesota using the technique as described by Cheng *et al.* [2013] (Figure 4 and Table 1). All ages fall close to the regression line, suggesting that the growth of the stalagmite occurred without any major break during 33,800 to 5500 years B.P. (Figures 4 and 5). The relationship between age and the distance of the

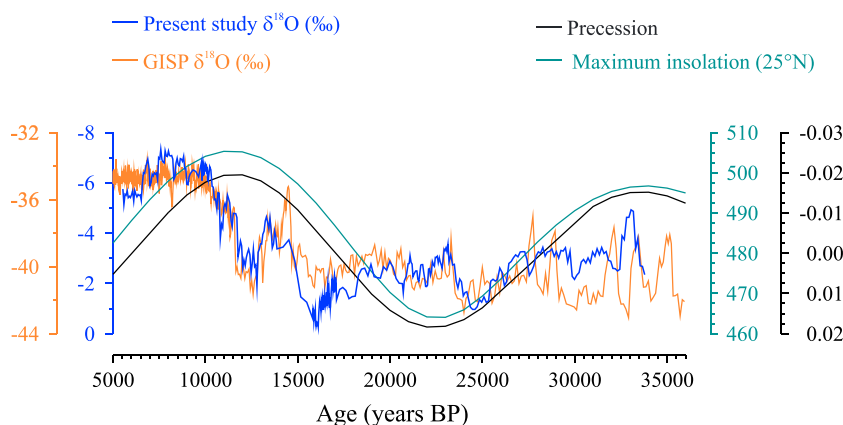


Figure 3. Mawmluh Cave $\delta^{18}\text{O}$ data (blue color) compared with Greenland Ice Sheet Project 2 (GISP2) core $\delta^{18}\text{O}$ data (orange color) [Stuiver and Grootes, 2000]. Also superimposed are maximum solar insolation at 25°N (cyan color) [Huybers, 2006] and orbital precession (black color) [Berger and Loutre, 1991].

proxy measurement along the growth axis of the speleothem is established using the StalAge Model [Scholz and Hoffmann, 2011]. The StalAge program is run in the open source statistical software R [R Development Core Team, 2010]. The StalAge program at first step identifies the major outliers. In next step, the age data are screened for minor outliers and age inversions, and the uncertainty of potential outliers is increased using an iterative procedure. Finally, the age model and corresponding 95% confidence limits are calculated by a Monte Carlo simulation fitting ensembles of straight lines to subsets of the age data (Figure 5). Calculated uncertainties are in the range of those calculated by combined application of Bayesian chronological ordering and a spline, providing the stratigraphic information in order to reduce age model uncertainty.

Sample ages were interpolated linearly between adjacent age measurements. The typical age uncertainty is 269 years from 33,800 to 24,000 years B.P. and less than 100 years from 24,000 to 5500 year B.P. (Table 1). Spectral analysis of the oxygen isotope time series shows statistically most significant (strongest) periodicity (95% confidence level) centered at 1700 year using PAST Red Fit method [Hammer *et al.*, 2001] (supporting information Figure S6) which resembles 1667 year suborbital cycle reported in Hulu Cave record from China [Clemens, 2005]. The other significant cycles lie at 1118, 802, and 602 years indicating millennial-scale changes in ISM strength/precipitation (Figure S6).

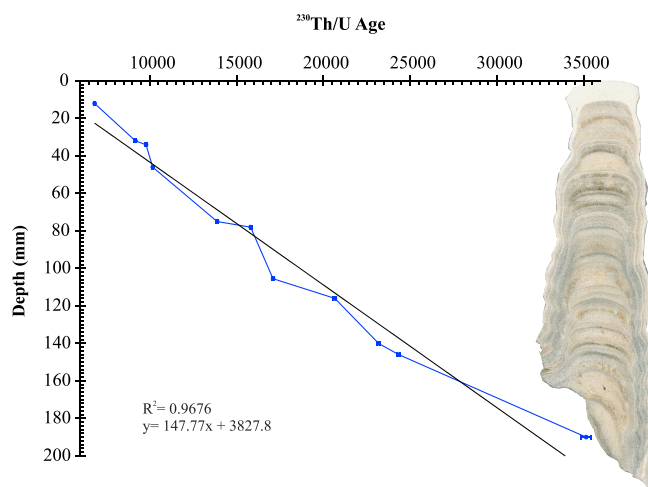


Figure 4. Age-depth relationship in MWS-1 speleothem sample (MWS-1). The U-Th ages are plotted with 2 sigma (2σ) error of dating. Also shown scanned image of working face of sample MWS-1.

3. Results and Discussion

This $\delta^{18}\text{O}$ data set from the Mawmluh Cave and the earlier published record from the same cave [Berkelhammer *et al.*, 2012] show fairly a good match; however, small divergence in the two records may have arisen due to different karst processes controlling different drips which fed the two cave samples (Figure 2). A Hendy test [Hendy, 1971] of one layer suggests no significant correlation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ (Figure S5). Comparison of $\delta^{18}\text{O}$ values with the earlier Mawmluh Cave record [Berkelhammer *et al.*, 2012] from 12,000 to 3650 years

Table 1. ^{230}Th Dating Results^a

Sample Number	Depth (mm)	^{238}U (ppb)	^{232}Th (ppt)	$^{230}\text{Th}/^{232}\text{Th}$ (atomic $\times 10^{-6}$)	$\delta^{234}\text{U}$ (Measured)	$^{230}\text{Th}/^{238}\text{U}$ (Activity)	^{230}Th Age (year) (Uncorrected)	^{230}Th Age (year) (Corrected)	$\delta^{234}\text{U}_{\text{Initial}}$ (Corrected)	^{230}Th Age (year B.P.) (Corrected)
MWS-1-T	12 $\pm$ 1	2,122 $\pm$ 7	2,758 $\pm$ 56	625 $\pm$ 13	-198.8 $\pm$ 1.8	0.0493 $\pm$ 0.0002	6937 $\pm$ 38	6,889 $\pm$ 51	-203 $\pm$ 2	6,826 $\pm$ 51
MWS-1-3	32 $\pm$ 1	1,967 $\pm$ 2	2,167 $\pm$ 44	989 $\pm$ 21	-187.8 $\pm$ 1.0	0.0661 $\pm$ 0.0003	9280 $\pm$ 47	9,241 $\pm$ 54	-193 $\pm$ 1	9,178 $\pm$ 54
MWS-1-4	34 $\pm$ 1	2,003 $\pm$ 2	3,621 $\pm$ 73	647 $\pm$ 13	-181.7 $\pm$ 1.1	0.0709 $\pm$ 0.0003	9918 $\pm$ 44	9,853 $\pm$ 63	-187 $\pm$ 1	9,790 $\pm$ 63
MWS-1-5	46 $\pm$ 1	2,146 $\pm$ 2	248 $\pm$ 7	10,501 $\pm$ 281	-175.5 $\pm$ 1.1	0.0737 $\pm$ 0.0002	10,241 $\pm$ 32	10,237 $\pm$ 33	-181 $\pm$ 1	10,174 $\pm$ 33
MWS-1-7	75 $\pm$ 1	1,574 $\pm$ 2	635 $\pm$ 14	4,109 $\pm$ 94	-159.9 $\pm$ 1.2	0.1006 $\pm$ 0.0004	13,960 $\pm$ 56	13,946 $\pm$ 57	-166 $\pm$ 1	13,883 $\pm$ 57
MWS-1-8	78 $\pm$ 1	2,945 $\pm$ 4	3,477 $\pm$ 70	1,577 $\pm$ 32	-164.9 $\pm$ 1.3	0.1129 $\pm$ 0.0003	15,915 $\pm$ 52	15,874 $\pm$ 60	-172 $\pm$ 1	15,811 $\pm$ 60
MWS-1-9	105 $\pm$ 1	3,738 $\pm$ 7	367 $\pm$ 8	20,161 $\pm$ 459	-171.4 $\pm$ 1.5	0.1201 $\pm$ 0.0003	17,169 $\pm$ 54	17,165 $\pm$ 54	-180 $\pm$ 2	17,102 $\pm$ 54
MWS-1-10	116 $\pm$ 1	1,748 $\pm$ 2	916 $\pm$ 21	4,448 $\pm$ 101	-179.5 $\pm$ 1.3	0.1413 $\pm$ 0.0005	20,749 $\pm$ 85	20,730 $\pm$ 86	-190 $\pm$ 1	20,667 $\pm$ 86
MWS-1-11	140 $\pm$ 1	4,796 $\pm$ 9	237 $\pm$ 9	51,005 $\pm$ 1870	-197.8 $\pm$ 1.4	0.1528 $\pm$ 0.0004	23,254 $\pm$ 77	23,252 $\pm$ 77	-211 $\pm$ 1	23,189 $\pm$ 77
MWS-1-12	146 $\pm$ 1	3,311 $\pm$ 5	601 $\pm$ 15	14,414 $\pm$ 361	-202.8 $\pm$ 1.4	0.1586 $\pm$ 0.0004	24,422 $\pm$ 84	24,416 $\pm$ 84	-217 $\pm$ 1	24,353 $\pm$ 84
MWS-1-B	190 $\pm$ 1	5,887 $\pm$ 26	8,638 $\pm$ 177	2,323 $\pm$ 49	-239.9 $\pm$ 1.7	0.2068 $\pm$ 0.0012	35,257 $\pm$ 266	35,199 $\pm$ 269	-265 $\pm$ 2	35,136 $\pm$ 269

^aThe error is 2σ . The method is based on Cheng *et al.* [2013]. Year B.P. = years before the present (A.D. 1950). Bold characters highlight the ^{230}Th age (year; corrected) and ^{230}Th age (year B.P.; corrected) used for constraining the chronology in the study.

using a different speleothem sample argues against kinetic fractionation and validates our data (Figure 2).

Speleothem $\delta^{18}\text{O}$ values are influenced by orbital variations that bring changes in local precipitation processes including transport pathways, moisture source, and seasonal precipitation balances [Wang *et al.*, 2001; Cheng *et al.*, 2009; LeGrande and Schmidt, 2009; Pausata *et al.*, 2011]. Breitenbach *et al.* [2010] proposed that the BoB surface waters serve as the primary source of moisture to Mawmluh Cave and therefore are the dominant influence on Mawmluh Cave $\delta^{18}\text{O}$. Such a linkage can be explained by the fact that strengthened monsoons increase river output to the BoB, therefore lowering the surface seawater $\delta^{18}\text{O}$.

A comparison of MWS-1 cave $\delta^{18}\text{O}$ with that of Hulu/Sanbao [Wang *et al.*, 2001] and Dongge Caves [Yuan *et al.*, 2004] from China suggests reasonable similarity which might be due to similar isotopic composition of source of air parcels in two regions. The eastern China caves receive moisture both from the Indian Ocean and from the Pacific Ocean, whereas moisture sources for the Mawmluh Cave is the Indian Ocean including the Bay of Bengal Arabian Sea [Breitenbach *et al.*, 2010; Liu *et al.*, 2014]. As such, MWS-1 speleothem $\delta^{18}\text{O}$ mainly reflects the Indian monsoon signal.

Mawmluh Cave $\delta^{18}\text{O}$ are plotted with maximum insolation at 25°N [Huybers, 2006], orbital precession [Berger and Loutre, 1991] and the Greenland Ice Sheet Project 2 (GISP 2) record to better assess the role of orbital insolation and high-latitude glaciation in driving changes in ISM strength (Figure 3). Mawmluh Cave $\delta^{18}\text{O}$ is exceptionally well aligned with GISP $\delta^{18}\text{O}$ from 25,000 years ago to the present. The most enriched Holocene values (-7‰ at 8 ka) lag maximum local solar insolation by 3500 years, consistent with the lag (relative to July 21 insolation) measured for the late Pleistocene Hulu/Sanbao record [Wang *et al.*, 2001, 2008]. These relationships indicate links to both solar insolation and North Atlantic climate signals transmitted via a westerlies teleconnection and sea ice. Increased North Atlantic sea ice extent causes cool temperatures throughout major parts of the Northern Hemisphere and the northern Indian Ocean. These cool anomalies, transported in the westerlies, delay the arrival of the Indian monsoon as well as reduce precipitation over the Indian subcontinent [Gupta *et al.*, 2003; Pausata *et al.*, 2011]. Such changes are likely to be recorded in heavier $\delta^{18}\text{O}$ values of cave carbonates.

Our record also shows good correlation with Northern Hemisphere tropical and subtropical climate proxy records from the BoB and Cariaco Basin, indicating both the tropics and subtropics were swept by similar climatic regimes during the studied interval (Figure 2). Gupta *et al.* [2003]

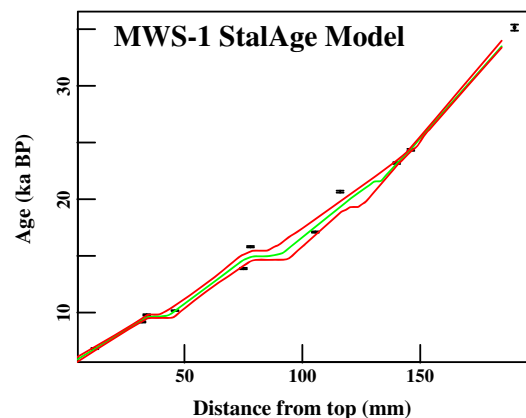


Figure 5. ^{230}Th age-depth relationship and the age model of MWS-1 stalagmite. The age model and corresponding 95% confidence limits are calculated by a Monte Carlo simulation fitting ensembles of straight line (green) to subsets of the age data. The age model is established by using ^{230}Th dates (Table 1) and StalAge [Scholz and Hoffmann, 2011]. The vertical error bars depict ^{230}Th dating errors (2σ). The MWS-1 sample total length is 19.2 cm growing for 28,310 years marking no hiatus.

linked such contemporaneous changes in the tropics and Northern Hemisphere to solar variability.

The Mawmluh Cave $\delta^{18}\text{O}$ record indicates a strengthened ISM circulation from 33,500 to 32,500 years B.P. and weakened circulation in the later part of marine oxygen isotope stage (MIS) 3 and early part of MIS 2 with a weak ISM phase during 26,000 to 23,500 years B.P. coinciding with a cold phase in the North Atlantic (Figure 3). During the latter interval, monsoon runoff to the Bay of Bengal weakened owing to a reduction in ISM rainfall [Govil and Naidu, 2011]. During the LGM the ISM circulation strengthened similar to that during the late MIS 3 followed by its weakest phase lasting from 17,000 to 15,000 years B.P. (Heinrich event H1). The Mawmluh Cave $\delta^{18}\text{O}$ record shows an abrupt increase in

ISM strength during the Bølling-Allerød from 15,000 to 12,900 years B.P. and early Holocene (~10,000 to 6500 years B.P.) warm intervals suggesting increased ISM strength, which corresponds to increased river discharge to the BoB [Govil and Naidu, 2011]. The Bølling-Allerød increase is consistent with the ISM maximum in northern India as documented in the speleothem $\delta^{18}\text{O}$ record from the Timta Cave [Sinha et al., 2005]. ISM strength decreased significantly during the Heinrich events H1–H3 and YD cold event (Figure 2).

The strengthened Holocene ISM (10,000–7000 years B.P.) is followed by a sudden decrease at 6500 years B.P. indicating a weak phase of ISM [Gupta et al., 2003]. Similarities among Mawmluh Cave $\delta^{18}\text{O}$, BoB planktic foraminifer $\delta^{18}\text{O}$, Chinese cave records and solar insolation suggest that changes in $\delta^{18}\text{O}$ of precipitation along the transport pathway and source effects were perhaps linked to solar variability [Wang et al., 2001; Gupta et al., 2003]. We agree with Yuan et al. [2004] that rainfall isotopic composition integrated from source to cave site increases with $\delta^{18}\text{O}$ depletion and vice versa. The same mechanism worked on centennial to millennial timescales [Pausata et al., 2011]. Mawmluh Cave $\delta^{18}\text{O}$ terminations align well with the GISP 2 ice core record, implicating a westerlies mechanism for transmitting the millennial-scale ice volume and sea ice signal from the North Atlantic to moisture $\delta^{18}\text{O}$ composition in Indian and Asian caves during cold intervals.

The largest amplitude excursions recorded in Mawmluh Cave $\delta^{18}\text{O}$ are associated with cold events including the YD (~13,000 years B.P.), Heinrich #1 (~17,000 years B.P.), and Heinrich #2 (~25,000 years B.P.). Variance within the warm Holocene interval (~10,000 years B.P. to the present) is significantly reduced compared to that in the colder MIS3 interval (35,000 to 23,000 years B.P.). If the underlying physics are similar, these data suggest that increased warmth will be accompanied by reduced amplitude monsoon variability.

4. Conclusions

Our results demonstrate that abrupt weak/strong phases in Indian summer monsoon (ISM) strength correspond to cold/warm intervals driven by weakening/strengthening of thermal contrasts between land and sea since MIS3. The novelty of our work lies in the fact that it is the southernmost longest record that is uniquely an Indian signal, offering the opportunity to assess isotopic contrasts relative to the Chinese records and the millennial-scale differences with other climate records. Our study suggests that $\delta^{18}\text{O}$ depletion along the transport pathway and isotopic composition of Bay of Bengal surface water mainly controlled $\delta^{18}\text{O}$ of Mawmluh Cave carbonate. We have also demonstrated explicitly for the first time the presence of YD cold event of Greenland with sharp boundaries in a cave carbonate record from the Indian landmass.

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References

- Allan, R. P., and B. J. Soden (2008), Atmospheric warming and the amplification of precipitation extremes, *Science*, **321**, 1481–1484, doi:10.1126/science.1160787.
- Berger, A., and M.-F. Loutre (1991), Insolation values for the climate of the last 10 million years, *Quat. Sci. Rev.*, **10**, 297–317, doi:10.1016/0277-3791(91)90033-Q.
- Berkelhammer, M., A. Sinha, L. Stott, H. Cheng, F. S. R. Pausata, and K. Yoshimura (2012), An abrupt shift in the Indian monsoon 4000 years ago, in *Climates, Landscapes, and Civilizations*, *Am. Geophys. Monogr.*, vol. 198, edited by L. Giosan et al., pp. 75–87, AGU, Washington, D. C., doi:10.1029/2012GM001207.
- Breitenbach, S. F. M., J. F. Adkins, H. Meyer, N. Marwan, K. K. Kumar, and G. H. Haug (2010), Strong influence of water vapor source dynamics on stable isotopes in precipitation observed in Southern Meghalaya, NE India, *Earth Planet. Sci. Lett.*, **292**, 212–220, doi:10.1016/j.epsl.2010.01.038.
- Buckley, B., et al. (2010), Climate as a contributing factor in the demise of Angkor, Cambodia, *Proc. Natl. Acad. Sci. U.S.A.*, **107**, 6748–6752, doi:10.1073/pnas.0910827107.
- Chao, W. C. (2000), Multiple quasi equilibria of the ITCZ and the origin of monsoon onset, *J. Atmos. Sci.*, **57**, 641–651.
- Cheng, H., R. L. Edwards, W. S. Broecker, G. H. Denton, X. Kong, Y. Wang, R. Zhang, and X. Wang (2009), Ice age terminations, *Science*, **326**, 248–252, doi:10.1126/science.1177840.
- Cheng, H., et al. (2013), Improvements in ^{230}Th dating, ^{230}Th and ^{234}U half-life values, and U–Th isotopic measurements by multi-collector inductively coupled plasma mass spectrometry, *Earth Planet. Sci. Lett.*, **371**–372, 82–91, doi:10.1016/j.epsl.2013.04.006.
- Clemens, S. C. (2005), Millennial-band climate spectrum resolved and linked to centennial-scale solar cycles, *Quat. Sci. Rev.*, **24**, 521–531, doi:10.1016/j.quascirev.2004.10.015.
- Dayem, K. E., P. Molnar, D. S. Battisti, and G. H. Roe (2010), Lessons learned from oxygen isotopes in modern precipitation applied to interpretation of speleothem records of paleoclimate from eastern Asia, *Earth Planet. Sci. Lett.*, **295**(1–2), 219–230.
- Dixit, Y., D. A. Hodell, and C. A. Petrie (2014), Abrupt weakening of the summer monsoon in northwest India ~4100 yr ago, *Geology*, **42**, 339–342, doi:10.1130/G35236.1.
- Govil, P., and P. D. Naidu (2011), Variations of Indian monsoon precipitation during the last 32 kyr reflected in the surface hydrography of the Western Bay of Bengal, *Quat. Sci. Rev.*, **30**, 3871–3879.
- Gupta, A. K., D. M. Anderson, and J. T. Overpeck (2003), Abrupt changes in the Asian southwest monsoon during the Holocene and their links to the North Atlantic Ocean, *Nature*, **421**, 354–357, doi:10.1038/nature01340.
- Hammer, Ø., D. A. T. Harper, and P. D. Ryan (2001), PAST: Paleontological statistics software package for education and data analysis, *Palaeontol. Electron.*, **4**, 1–9.
- Hendy, C. H. (1971), The isotopic geochemistry of speleothems—I. The calculation of the effects of different modes of formation on the isotopic composition of speleothems and their applicability as palaeoclimatic indicators, *Geochim. Cosmochim. Acta*, **35**, 801–824, doi:10.1016/0016-7037(71)90127-X.
- Huybers, P. (2006), Early Pleistocene glacial cycles and the integrated summer insolation forcing, *Science*, **313**, 508–511, doi:10.1126/science.1125249.
- Intergovernmental Panel on Climate Change (IPCC) (2013), Climate change 2013: The physical science basis, in *Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by T. F. Stocker et al., pp. 35–118, Cambridge Univ. Press, Cambridge, U. K.
- Lea, D. W., D. K. Pak, L. C. Peterson, and K. A. Hughen (2003), Synchronicity of tropical and high-latitude Atlantic temperatures over the last glacial termination, *Science*, **301**, 1361–1364, doi:10.1126/science.1088470.
- LeGrande, A. N., and G. A. Schmidt (2009), Sources of Holocene variability of oxygen isotopes in paleoclimate archives, *Clim. Past*, **5**, 441–455.
- Liu, Z., et al. (2014), Chinese cave records and the East Asia Summer Monsoon, *Quat. Sci. Rev.*, **83**, 115–128, doi:10.1016/j.quascirev.2013.10.021.
- Murata, F., T. Hayashi, J. Matsumoto, and H. Asada (2007), Rainfall on the Meghalaya plateau in northeastern India: One of the rainiest places in the world, *Nat. Hazards*, **42**(2), 391–399.
- Parthasarathy, K. (1960), Some aspects of rainfall in India during the Southwest monsoon season, in *Proceedings of the Symposium on Monsoon of the World*, pp. 185–194, India Meteorological Department, New Delhi.
- Pausata, F. S. R., D. S. Battisti, K. H. Nisancioglu, and C. M. Bitz (2011), Chinese stalagmite $\delta^{18}\text{O}$ controlled by changes in the Indian monsoon during a simulated Heinrich event, *Nat. Geosci.*, **4**(7), 474–480.
- Pavelsky, T. M., S. Sobolowski, S. B. Kapnick, and J. B. Barnes (2012), Changes in orographic precipitation patterns caused by a shift from snow to rain, *Geophys. Res. Lett.*, **39**, L18706, doi:10.1029/2012GL052741.
- Porter, S. C., and A. Zisheng (1995), Correlation between climate events in the North Atlantic and China during the last glaciation, *Nature*, **375**, 305–308, doi:10.1038/375305a0.
- R Development Core Team (2010), *R: A Language and Environment for Statistical Computing*, R Foundation for Statistical Computing, Vienna, Austria.
- Scholz, D., and D. L. Hoffmann (2011), StalAge—An algorithm designed for construction of speleothem age models, *Quat. Geochronol.*, **6**, 369–382.
- Sinha, A., K. G. Cannariato, L. D. Stott, H.-C. Li, C.-F. You, H. Cheng, R. L. Edwards, and I. B. Singh (2005), Variability of Southwest Indian summer monsoon precipitation during the Bølling-Ållerød, *Geology*, **33**, 813–816, doi:10.1130/G21498.1.
- Staubwasser, M., F. Sirocko, P. M. Grootes, and M. Segl (2003), Climate change at the 4.2 ka BP termination of the Indus valley civilization and Holocene south Asian monsoon variability, *Geophys. Res. Lett.*, **30**, 1425–1428, doi:10.1029/2002GL016822.
- Stuiver, M., and P. M. Grootes (2000), GISP2 oxygen isotope ratios, *Quat. Res.*, **53**, 277–284, doi:10.1006/qres.2000.2127.
- Sun, Y., S. C. Clemens, C. Morrill, X. Lin, X. Wang, and Z. An (2011), Influence of Atlantic meridional overturning circulation on the East Asian winter monsoon, *Nat. Geosci.*, **5**, 46–49, doi:10.1038/NGEO1326.
- Trenberth, K. E., A. Dai, R. M. Rasmussen, and D. B. Parsons (2003), The changing character of precipitation, *Bull. Am. Meteorol. Soc.*, **84**, 1205–1217, doi:10.1175/BAMS-84-9-1205.
- Wang, Y. J., H. Cheng, R. L. Edwards, Z. S. An, J. Y. Wu, C.-C. Shen, and J. A. Dorale (2001), A high-resolution absolute-dated Pleistocene monsoon record from Hulu Cave, China, *Science*, **294**, 2345–2348, doi:10.1126/science.1064618.
- Wang, Y. J., H. Cheng, R. L. Edwards, X. Kong, X. Shao, S. J. W. Chen, X. Jiang, X. Wang, and Z. An (2008), Millennial- and orbital-scale changes in the East Asian monsoon over the past 224,000 years, *Nature*, **451**, 1090–1093, doi:10.1038/nature06692.
- Yancheva, G., et al. (2007), Influence of the intertropical convergence zone on the East Asian monsoon, *Nature*, **445**, 74–77, doi:10.1038/nature05431.
- Yuan, D., et al. (2004), Timing, duration, and transitions of the last interglacial Asian monsoon, *Science*, **304**, 575–578, doi:10.1126/science.1091220.