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Strong influence of water vapor source dynamics on stable isotopes in precipitation observed in Southern Meghalaya, NE India

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A B S T R A C T

To calibrate $\delta^{18}O$ time-series from speleothems in the eastern Indian summer monsoon (ISM) region of India, and to understand the moisture regime over the northern Bay of Bengal (BoB) we analyze the $\delta^{18}O$ and δD of rainwater, collected in 2007 and 2008 near Cherrapunji, India. δD values range from +18.5‰ to −144.4‰, while $\delta^{18}O$ varies between +0.8‰ and −18.8‰. The Local Meteoric Water Line (LMWL) is found to be indistinguishable from the Global Meteoric Water Line (GMWL). Late ISM (September–October) rainfall exhibits lowest $\delta^{18}O$ and δD values, with little relationship to the local precipitation amount. There is a trend to lighter isotope values over the course of the ISM, but it does not correlate with the patterns of temperature and rainfall amount. $\delta^{18}O$ and δD time-series have to be interpreted with caution in terms of the 'amount effect' in this subtropical region. We find that the temporal trend in $\delta^{18}O$ reflects increasing transport distance during the ISM, isotopic changes in the northern BoB surface waters during late ISM, and vapor re-equilibration with rain droplets. Using an isotope box model for surface ocean waters, we quantify the potential influence of river runoff on the isotopic composition of the seasonal freshwater plume in the northern BoB. Temporal variations in this source can contribute up to 25% of the observed changes in stable isotopes of precipitation in NE India. To delineate other moisture sources, we use backward trajectory computations and find a strong correlation between source region and isotopic composition. Palaeoclimatic stable isotope time-series from northeast Indian speleothems likely reflect changes in moisture source and transport pathway, as well as the isotopic composition of the BoB surface water, all of which in turn reflect ISM strength. Stalagmite records from the region can therefore be interpreted as integrated measures of the ISM strength.

1. Introduction

The climate of Northeastern India and Bangladesh is dominated by the Indian summer monsoon (ISM), characterized by seasonally strong rainfall and floods. The variability of the ISM, its onset, duration, and failure has clear impacts on regional societies and agriculture (Butler, 1908; Mukherjee et al., 2007). To better understand, and potentially forecast, monsoonal dynamics researchers have attempted to reconstruct past climate conditions of the region, by linking terrestrial proxy records (e.g. stalagmites) to hydrological and meteorological conditions (Fleitmann et al., 2003, 2007; Sinha et al., 2007; Staubwasser and Weiss, 2006; Wang et al., 2005, 2008; Yancheva et al., 2007; Zhang et al., 2008).

Delineating the relationship between vapor source and transport pathways to rainfall $\delta^{18}O$ on synoptic timescales may help to understand the factors that lead to changes in palaeoclimate archives. Calibrations of this type require that we understand the modern regional hydrological conditions. Whereas Europe and North America are covered with a dense network of meteorological and hydrological stations, publicly available data for NE India are limited. In addition, the correlation between the well known all India rainfall index (AIRI) and rainfall in the Bay of Bengal (BoB) in the NE Indian summer is 0.06 (Hoyos and Webster, 2007). The AIRI cannot be taken as representative for our study area. Early regional isotope studies (Bhattacharya et al., 1985; Krishnamurthy and Bhattacharya, 1991) give only short-term results so that uninterrupted long-term records are lacking. Overall the stable isotope variability of Indian monsoonal precipitation

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has been characterized for only a few stations (Araguaś-Araguaś et al., 1998; Bhattacharya et al., 2003; Datta et al., 1991; Sengupta and Sarkar, 2006).

The isotopic signature of precipitation provides valuable information about vapor source and atmospheric circulation pattern that give rise to the rainfall, and can be used to reconstruct past climate conditions. δ^{18}O and δD in precipitation show a distinct empirical relationship, described by the Global Meteoric Water Line (GMWL, $\delta D = 8 \times \delta^{18}O + 10$). This relation was first established for fresh surface waters and then for precipitation itself (Craig, 1961; Rozanski et al., 1993). Comparing precipitation isotope data with the GMWL helps to understand the precipitation pathways of a given region. Several effects (such as moisture re-cycling, and source water evaporation) can lead to deviations of rainwater isotopes of a given location from the GMWL, and the δ^{18}O–δD relationship is then better described by a Local Meteoric Water Line (LMWL).

For many tropical regions, where no correlation is observed between rainfall δ^{18}O and temperature, a statistical relationship was found between δ^{18}O and rainfall amount. This empirical 'amount effect' of a negative correlation between δ^{18}O and rainfall amount was first noted by Dansgaard (1964). He, and later Rozanski et al. (1993), recognized several mechanisms for this relationship in regions where rainfall δ^{18}O does not correlate with temperature. First, with increasing convection intensity and cloud height, vapor cooling in the cloud will lead to depleted δ-values of the condensate. Second, smaller drops from light rainfall can reach isotopic equilibrium with the vapor below the cloud. Third, evaporation of droplets falling through low humidity air below the cloud increases the δ-values in the remaining droplet. Thus, dry season rainfall is likely enriched in ^{18}O and ^{2}H, and can fall below the GMWL. Fourth, isotopic depletion of water vapor and subsequent rainfall with time is more likely in heavy showers. Dansgaard (1964) concluded that low δ-values in heavy tropical rainfall are the result of deep convectional cooling and reduced influence of enrichment processes. However, Risi et al. (2008) show that the amount effect is governed by processes related to fall and re-evaporation of droplets rather than processes occurring during air ascent in the cloud. This finding is similar to results of a numerical model designed by Lee and Fung (2007). They conclude that the amount effect depends on drop size and degree of equilibration with the sub-cloud atmospheric layer.

In this contribution we show, that none of the processes ascribed for the amount effect is found in our dataset from NE India and we propose several other controlling mechanisms for the observation that rainfall trends to lighter values during the extent of the ISM.

2. Geographical setting

The sampling site (915 m above sea level (m.a.s.l.), 25°13′09″ N, 91°39′46″ E, Fig. 1) is located ~400 m below, and just south of, the meteorological station at Cherrapunji (WMO station # 42515 at 25°18' N, 91°42' E, 1313 m.a.s.l.) on the south-facing ridge of the Meghalaya Plateau. Meghalaya receives virtually all of its moisture from the BoB during the monsoon season. The air masses coming from the south are split into a NE and a NW arm in front of the Himalaya. The orographic barriers of the Meghalaya Plateau and the Himalaya (Bookhagen et al., 2005), together with lifting of unstable upstream air (Murata et al., 2008), lead to excessive windward rainfall.

Located close to the Tropic of Cancer, this area experiences a strongly seasonal climate (Fig. 2). The meteorological year can be divided into three seasons: pre-monsoon (January to May), ISM (June to mid-October), and post-monsoon (mid-October to December) (similar to Mukherjee et al., 2007). Pre-monsoon rainfall derives mostly from local convection. During the ISM, the study area receives >80% of its annual rainfall (Fig. 2). The ISM is governed by deep convection and cyclonic activity over the Indian Ocean and the BoB, transporting large, vapor saturated air masses over Meghalaya and onto the southeastern Tibetan Plateau (Tian et al., 2001a,b; Zhou and Yu, 2005, Yoon and Chen, 2005). The orographic rise leads to adiabatic cooling of the northward moving air masses and subsequently to heavy rainfall. During the post-monsoon, relatively drier conditions prevail, with occasional rainfall associated with atmospheric disturbances over the BoB.

A strong trend in the isotopic composition in rainfall, with lowest δ^{18}O and δD values during late ISM, has been reported for N India and Nepal (Gajurel et al., 2006). The Global Network for Isotopes in Precipitation (GNIP) mean oxygen isotopic composition of precipitation for February falls between −2‰ and +2‰ VSMOW (Vienna Standard Mean Ocean Water) and for September between −14‰ and −10‰ VSMOW (data from the International Atomic Energy Agency GNIP; IAEA 2006). Previous work attributes the precipitation isotope signal in our study area to seasonal changes of the vapor source and to the amount effect (Aggarwal et al., 2004; Araguaś-Araguaś et al., 1998; Dansgaard, 1964; Mukherjee et al., 2007; Zhang et al., 2006). However, similar to the pattern at our site, a temporal offset between

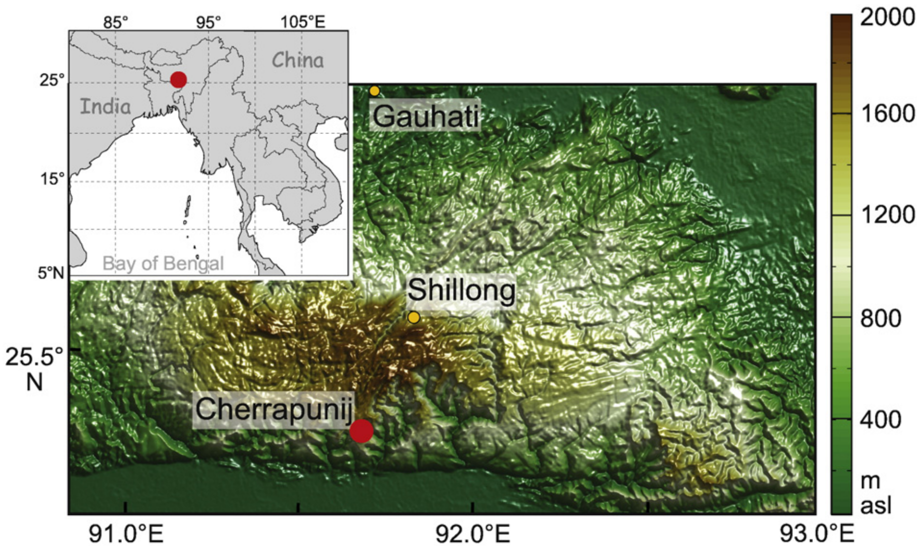

Fig. 1. Digital elevation map of the Meghalaya Plateau (SRTM elevation data courtesy of the U.S. Geological Survey, http://seamless.usgs.gov/). The southern border shows the sharp contrast between the lowlands of Bangladesh and the highlands of the Plateau.

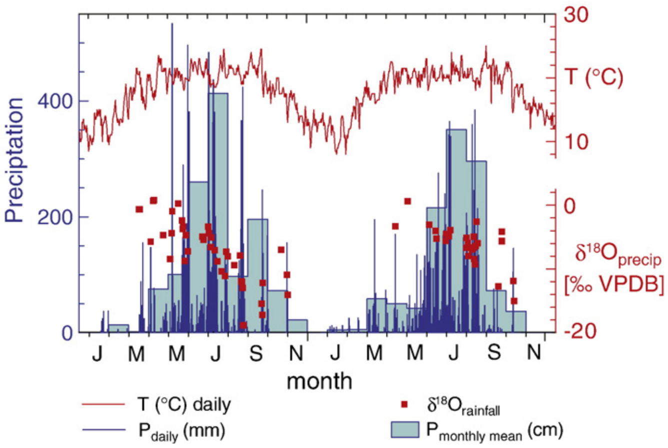

Fig. 2. Climate data from the Cherrapunji meteorological station (WMO station #42515) over 2007 and 2008. Daily temperature (T) is given together with monthly averaged and daily precipitation (P) for the years 2007 and 2008. Annual rainfall exceeds 10,700 mm. Individual $\delta^{18}O$ (‰ VSMOW) values of rainwater are presented as red squares. The strong seasonality of precipitation during the monsoon is not mirrored by the isotopic data. The Pune Institute of Tropical Meteorology provides daily precipitation and temperature data.

the highest rainfall and the most negative $\delta^{18}O$ is also found at Colcatta, New Delhi, Mumbai, Shillong, Bangkok, in SW China, and Myanmar (Datta et al., 1991; Araguaś-Araguaś et al., 1998; Rozanski et al., 1993; Bhattacharya, et al., 2003; Sengupta and Sarkar, 2006; Zhang et al., 2006, Kurita et al., 2009). This agreement between stations around the BoB supports the idea that our dataset is representative for the region and sampling period.

The BoB mixed layer is very shallow (5–15 m, Vinayachandran et al., 2002), and runoff from several of the largest river systems in the world have a profound impact on its salinity and isotopic composition (Delaygue et al., 2001). Salinity and $\delta^{18}O$ variability exhibit a linear relationship that depends on evaporation, runoff, the $\delta^{18}O$ of rainfall, and oceanic mixing (Delaygue et al., 2001). However, the influence of the $\delta^{18}O$ of runoff on the BoB surface waters has yet to be quantified. Additionally, the flood plains of Bangladesh and Bengal might serve as a depleted vapor source during late ISM.

3. Methods

3.1. Sampling and analysis

68 rainwater samples have been collected over the period from March 2007 to October 2008 (Table 1). One to twelve samples per month were collected, each representing snap-shots of individual rainfall events. Due to logistical restrictions, we conducted event-sampling only, without volume and/or time integration. Some rain fell at the sampling site when no rainfall was recorded at the meteorological station Cherrapunji. Between November 2007 and mid-January 2008, no rainfall occurred at the study site. Little rainfall occured from January to mid-March 2008 (47.9 mm in January, 54.0 mm in February, 48.1 mm in mid-March). In the second half of March, 536 mm of rainfall were recorded at Cherrapunji. Due to technical reasons we were not able to collect rainwater between January and March 2008. The rainwater samples presented here were collected mostly during the pre-monsoon and the ISM, with very few samples representing the post-monsoon period. Notwithstanding the relatively short sampling period and the event sampling method, we can use the collected data to carefully test our hypothesis.

60 ml samples were stored in sealed airtight Nalgene® polypropylene bottles. Aliquots were analyzed for their stable isotope composition (δD and $\delta^{18}O$, referenced to VSMOW) at the Alfred

Table 1
Isotopic data for the water samples collected in 2007–2008.

Sample	Sampling date	Season[a]	Lab. No.	$\delta^{18}O$ (VSMOW)	δD (VSMOW)	d excess
1	02.–03.04.2007	P	5756	−0.66	7.9	13.1
2	03.04.2007	P	5758	−0.65	5.5	10.7
3	20.04.2007	P	5759	−5.71	−28.8	16.9
4	24.04.2007	P	5760	0.72	18.0	12.2
5	25.04.2007	P	5764	0.76	18.5	12.4
6	09.05.2007	P	5765	−4.66	−20.7	16.6
7	20.05.2007	P	5766	−8.39	−54.8	12.3
8	22.05.2007	P	5768	−4.34	−15.7	19.0
9	23.05.2007	P	5769	−0.93	5.5	12.9
10	01.06.2007	M	5771	0.28	10.1	7.9
11	07.06.2007	M	5772	−2.35	−5.4	13.4
12	08.06.2007	M	5773	−3.73	−14.5	15.4
13	09.06.2007	M	5774	−3.33	−10.7	15.9
14	11.06.2007	M	5776	−8.71	−61.3	8.4
15	13.06.2007	M	5777	−4.64	−22.6	14.5
16	16.06.2007	M	5778	−7.23	−44.1	13.7
17	08.07.2007	M	5802	−4.94	−26.9	12.7
18	10.07.2007	M	5804	−5.43	−30.7	12.7
19	17.07.2007	M	5805	−3.35	−14.3	12.5
20	18.07.2007	M	5806	−4.37	−21.7	13.3
21	21.07.2007	M	5879	−6.56	−40.2	12.3
22	22.07.2007	M	5880	−5.06	−28.6	11.8
23	27.07.2007	M	5881	−7.07	−43.6	12.9
24	31.07.2007	M	5883	−8.78	−55.5	14.8
25	07.08.2007	M	5884	−10.35	−70.8	12.0
26	12.08.2007	M	5886	−11.07	−76.8	11.8
27	14.08.2007	M	5887	−7.31	−47.9	10.5
28	16.08.2007	M	5888	−8.01	−51.6	12.5
29	26.08.2007	M	5889	−9.39	−61.1	14.1
30	04.09.2007	M	5891	−7.87	−49.3	13.6
31	06.09.2007	M	5892	−11.92	−86.4	9.0
32	07.09.2007	M	5893	−11.98	−86.8	9.1
33	08.09.2007	M	5894	−12.97	−94.2	9.5
34	09.09.2007	M	5896	−18.82	−144.4	6.2
35	07.10.2007	M	5897	−15.43	−115.9	7.5
36	08.10.2007	M	5898	−17.19	−123.7	13.8
37	09.10.2007	M	5971	−12.19	−83.3	14.2
38	06.11.2007	PM	5972	−7.00	−46.7	9.3
39	15.11.2007	PM	5973	−10.91	−78.5	8.8
40	16.11.2007	PM	5975	−14.06	−99.5	13.0
41	29.04.2008	P	14,814	−3.31	−9.7	16.7
42	18.05.2008	P	14,815	0.62	14.5	9.5
43	20.06.2008	M	14,816	−3.07	−12.9	11.6
44	30.06.2008	M	14,817	−4.01	−22.5	9.6
45	01.07.2008	M	14,819	−5.21	−32.8	8.9
46	15.07.2008	M	14,820	−5.57	−33.9	10.7
47	16.07.2008	M	14,821	−4.53	−29.5	6.8
48	20.07.2008	M	14,822	−4.85	−27.4	11.4
49	22.07.2008	M	14,824	−3.92	−18.5	12.9
50	16.08.2008	M	14,825	−5.10	−32.4	8.4
51	16.08.2008	M	14,826	−6.66	−42.1	11.1
52	17.08.2008	M	14,827	−6.71	−48.7	5.0
53	18.08.2008	M	14,829	−5.90	−37.7	9.5
54	19.08.2008	M	14,830	−7.96	−54.0	9.6
55	25.08.2008	M	14,831	−6.92	−43.0	12.3
56	27.08.2008	M	14,837	−4.94	−27.1	12.4
57	27.08.2008	M	14,838	−6.61	−40.6	12.3
58	28.08.2008	M	14,839	−8.45	−55.4	12.2
59	29.08.2008	M	14840	−9.27	−64.1	10.0
60	30.08.2008	M	14842	−6.40	−38.5	12.8
61	31.08.2008	M	14843	−2.58	−16.6	4.0
62	02.09.2008	M	14,844	−5.95	−35.6	12.0
63	04.10.2008	M	14,845	−12.71	−89.2	12.4
64	09.10.2008	M	14,847	−5.64	−27.4	17.7
65	09.10.2008	M	14,848	−5.57	−29.4	15.1
66	09.10.2008	M	14,849	−4.15	−23.0	10.2
67	27.10.2008	PM	14,850	−11.90	−80.9	14.3
68	28.10.2008	PM	14,852	−15.00	−104.6	15.4
Average				−6.73	−42.0	11.9
STDEV				4.25	34.4	2.9

[a] P = pre-monsoon, M = monsoon, PM = post-monsoon.

Wegener Institute for Marine and Polar Research (AWI) Potsdam, Germany. A Finnigan MAT Delta-S mass spectrometer equipped with two equilibration units was used for the online determination of hydrogen and oxygen isotopic composition. The external errors for standard measurements of hydrogen and oxygen are better than 0.8‰ and 0.1‰, respectively (Meyer et al., 2000).

3.2. Trajectory computations

To deduce the probable source regions of the air masses from which our water samples derived, we generated backward trajectories based on the Hybrid Single-Particle Lagrangian Integrated Trajectories (HYSPLIT) ARL trajectory tool database of the National Oceanic and Atmospheric Administration (NOAA, http://www.arl.noaa.gov/ready/hysplit4.html). Trajectories for time periods of 98 h were computed for 500, 1500, and 2500 m.a.s.l., because rainfall is expected to originate from these altitudes. However, backward trajectory analysis indicates only the synoptic situation, and is only an approximation of the general origin of an air mass. Minor local moisture sources cannot be excluded. The trajectories were calculated for every rainwater-sampling day. Additionally, we calculated weekly backward trajectories for 2007 and 2008. This procedure helped to identify changes in the source regions for rainfall over our study site.

4. Results

4.1. Stable isotope variations in precipitation

Rainfall δD values range from +18.5‰ to −144.4‰ while the $\delta^{18}O$ values range from +0.8‰ to −18.8‰ (Table 1, Fig. 2). Highest δD and $\delta^{18}O$ values are found in April and May and lowest values are found in September and October. The isotope values show a clear gradual decrease between April and October that is not reflected in either temperature or rainfall amount. Generally, δD and $\delta^{18}O$ values are higher during pre-monsoon months compared to months of monsoonal precipitation. Lowest δD and $\delta^{18}O$ values are observed during the late monsoon period (September–October).

4.2. Deuterium excess d

The deuterium excess ($d = \delta D - 8 \times \delta^{18}O$) was defined by Dansgaard (1964). It is used as a measure for non-equilibrium conditions during source water evaporation, and it depends on relative humidity (rH) and sea surface temperature (SST) in the source region. Lower d excess in precipitation should be observed during times of high rH over the primary source region. Although d excess does not show a defined trend at our site, highest annual values are found in April and May

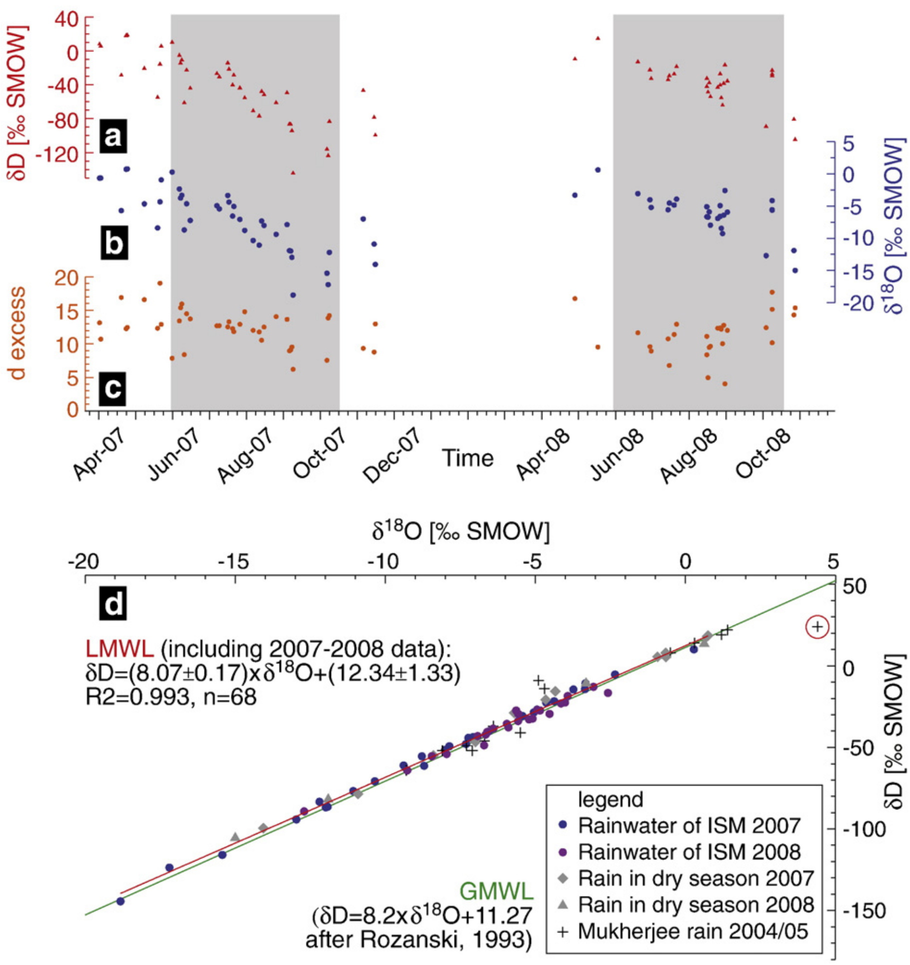

Fig. 3. Plots of the seasonal variation in a) δD, b) $\delta^{18}O$, and c) d excess, based on rainwater isotopes collected 2007–2008 near Cherrapunji. $\delta^{18}O$ and δD are depleted during the ISM (shaded) and its aftermath. d excess shows no significant change over the observation period. d) $\delta^{18}O$–δD diagram of water samples collected for this study and in West Bengal in 2004/05 (black crosses) (Mukherjee et al., 2007). Most samples closely follow the GMWL with one point (red circle) that are prone to evaporation effects. The red linear fit indicates the LMWL, based on the samples presented in this study. The GMWL with $\delta D = 8.2x\ \delta^{18}O + 11.27$ (Rozanski et al., 1993) is plotted for reference.

(~19‰), and lowest in August–September (~4‰). SST trends in the BoB can contribute only a small amount of the observed d excess variability (NOAA, 2005). Tian et al. (2001a) relate lower d excess values on the Tibetan Plateau during ISM to enhanced moisture supply from the marine BoB region.

4.3. Local meteoric water line

Prior to this work, modern rainwater isotope data has been scarce for Meghalaya (Bhattacharya et al., 1985). Our $\delta^{18}O$ and δD from local rainwater samples collected in 2007 and 2008 show a linear relationship (Fig. 3d). For the first time, a tentative (due to the limited number of samples) LMWL for Meghalaya can be described (Fig. 3d):

$$\delta D = (8.07 \pm 0.17) \times \delta^{18}O + (12.35 \pm 1.33) \quad (1)$$

with an r^2 of 0.993 ($n = 68$) and the 95% confidence intervals of the regression parameters based on Sachs (1984). This LMWL compares very well to the GMWL $\delta D = 8.2 \times \delta^{18}O + 11.27$ (Rozanski et al., 1993), but has a slope and an intercept higher than the LMWL for West Bengal ($\delta D = 7.2 \times \delta^{18}O + 7.7$, $n = 14$) (Mukherjee et al., 2007). Most of the difference between the data provided by Mukherjee and co-workers and the data presented here disappears if their most enriched sample (red circle in Fig. 3d) is excluded from the calculation of their LMWL. This sample is very heavy in both, δD (+24‰) and $\delta^{18}O$ (+4.4‰) and is therefore likely to have been exposed to evaporation. A much better fit of their LMWL ($\delta D = 8.07 \times \delta^{18}O + 12.60$) to both the GMWL and our LMWL is found without it. If only ISM samples for 2007 and 2008 are considered, the LMWL ($\delta D = (8.13 \pm 0.20) \times \delta^{18}O + (12.55 \pm 1.62)$, $n = 50$, $r^2 = 0.993$) has a higher slope compared to the dry season. The slope decreases during the pre-monsoon ($\delta D = (7.57 \pm 0.62) \times \delta^{18}O + (13.14 \pm 2.45)$, $n = 10$, $r^2 = 0.989$) and the post-monsoon ($\delta D = (7.26 \pm 0.83) \times \delta^{18}O + (3.39 \pm 10.04)$, $n = 5$, $r^2 = 0.993$) periods. However, large error estimates for the pre- and post-monsoon LMWL (due to the small numbers of samples) mean that seasonal changes in the LMWL cannot be taken as significant within this dataset. In the context of stalagmite records, the pre- and post-monsoon rainfall adds only a small amount of water to the sub-surface levels, and probably do not contribute to the cave dripwater isotope signature (see also discussion below).

5. Discussion

$\delta^{18}O$ changes at New Delhi, Mumbai, Bangkok, Yangoon (Myanmar), and southern China have been attributed to the amount effect (Clark and Fritz, 1997; Datta et al. 1991, Rozanski et al., 1993; Zhang et al., 2006, Kurita et al., 2009). However, in our isotope time-series from the Cherrapunji region, there is no conclusive evidence for the amount effect. Our dataset agrees well with available data from the BoB realm (see above and Kurita et al., 2009), and hence is unlikely to result from inadequate sampling.

Rainwater $\delta^{18}O$ and δD are distributed closely along the GMWL, suggesting that no significant evaporative enrichment takes place during rainfall (Fig. 3). The slope of the water line depends on rH during secondary sub-cloud evaporation, such that evaporation of droplets after condensation from the cloud lead to a slope <8 (Clark and Fritz, 1997). Our slope value of 8.07 ± 0.17 is not statistically different to the GMWL, and indicates little secondary evaporation in falling rain and high rH in the source area (Datta et al., 1991). These data are consistent with an Indian Ocean moisture source. Further, the Cherrapunji $\delta^{18}O_{rainfall}$ samples show no negative correlation to rainfall amount (Fig. 4). Our time-series of $\delta^{18}O$ and δD in rainwater show a clear trend to depleted δ-values during the late ISM, while

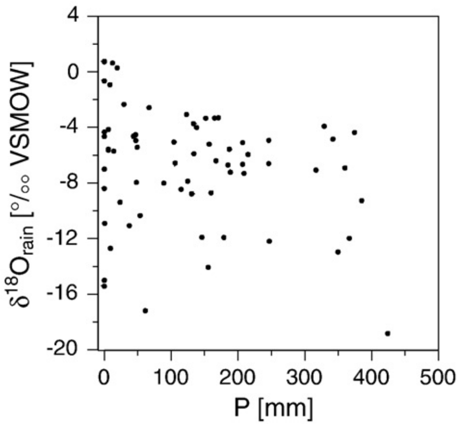

Fig. 4. Comparison between $\delta^{18}O$ in rainwater and the amount of rainfall for the time of isotope sampling. There is no empirical amount effect in our data set.

rainfall peaks earlier in the season, similar to the trend seen at Shillong (IAEA 2006).

These observations lead us to search for factors other than the amount effect to explain the isotopic composition of rainfall over Meghalaya, at least for the considered time span. The study site receives rainfall mainly during the ISM, and hence stalagmite records likely reflect ISM intensity and source effects. Below, we discuss the different factors that influence the stable isotopes in rainfall over Meghalaya, and thus potentially in palaeoclimate proxy records. Factors that can influence the rainfall $\delta^{18}O$ at our study site include: a) a change in the moisture source from the continent to the Indian Ocean, b) changes in the distance an air mass travels over the BoB and Indian Ocean, c) temporal changes in the moisture source due to freshwater runoff, and d) isotopic disequilibrium between seawater and vapor during stormy weather.

5.1. Moisture source variability

One important reason for the shift towards depleted rainwater isotope values during the ISM is a change in moisture source. The history of the air mass trajectories for all 68 of our samples is evaluated using backward trajectories for the 98 h before reaching the sampling site (Fig. 5). To control whether our samples are representative for the time of the year, we calculated trajectories for every week during 2007 and 2008 (not shown). By comparing the different seasons we find a change in moisture source just before the onset of the ISM. Two major source regions are found, one over the continent (central Asia and India) during the winter, and one over the Indian Ocean during the summer.

The dry winter is characterized by a robust air mass source in central Asia and NW India. Air transport is presumably forced by the enhancement of the Tibetan High, a southward Intertropical convergence zone (ITCZ) migration, and an associated retreat of the ISM. During fieldwork in the winter we observed strong re-evaporation of rain at our site, sometimes affecting 100% of the droplets. Our attempts to sample this presumably enriched rainfall were hampered by its presence as a 'fog' of non-condensing moisture. Therefore, we assume that only a minimal amount of water from this source can reach the sub-soils and cave water systems. ISM air masses, on the other hand, originated from the Indian Ocean and were directed to the northern BoB, before reaching Cherrapunji, in agreement with results from Zhou and Yu (2005). The ISM is related to the northward migration of the ITCZ and its associated atmospheric circulation. These two sources have distinct $\delta^{18}O$ signatures. The winter air is likely characterized by enriched $\delta^{18}O$ from recycled "continental" moisture, while the maritime air from the ocean is strongly depleted (Araguaś-Araguaś et al., 1998). However, this difference does not explain the trend to

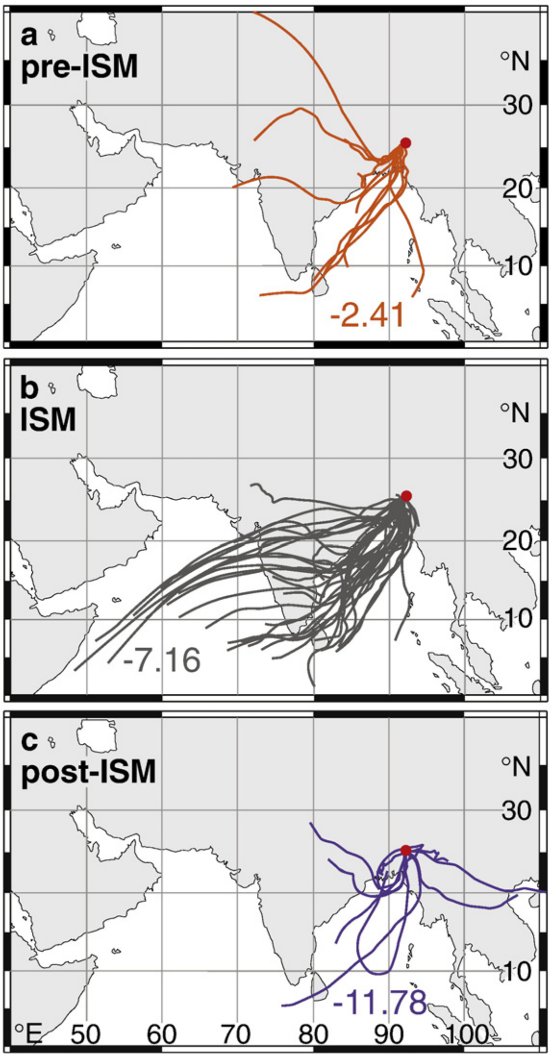

Fig. 5. Air mass trajectories are calculated for 98 h before reaching our site for every rainwater sampling day. Trajectories are plotted for 2007 and 2008 for a) the pre-ISM, b) the ISM, and c) the post-ISM seasons and clearly reflect the seasonally changing source regions. Pre-monsoon air masses reach Cherrapunji mainly from the BoB and India, while ISM air travels from the BoB and as far west as the Arabian Sea. Post-monsoon air masses arrive mainly from the northern BoB and Bangladesh and India lowlands. Rain-bearing air originates from the SW and S, with very few exceptions. The numbers given are the average $\delta^{18}O$ values for the rainwater collected during the respective season. The most negative values are found during the post-ISM, when a larger contribution from the BoB freshwater plume is expected.

most depleted values during late ISM, across a peak in rainfall amount, in our (and other stations) isotope data.

5.2. Change in air mass travel distance

We observe an increase in the distance between the moisture source and our site during summer (Fig. 5). Pre-monsoon and early-ISM rain stems largely from the BoB, while later in the ISM season the transport pathway moves to the open Indian Ocean and the Arabian Sea. Such an increase in distance allows for enhanced Rayleigh distillation during the transport to Cherrapunji and hence for a trend to isotopic depletion in rainfall over the ISM. The trajectories for rain-bearing post-monsoon air reflect transport from the northern BoB, Bangladesh, and from Asia, with clearly very negative $\delta^{18}O$ values. If distance from the source to our site was the only control on the fraction of vapor remaining after Rayleigh fractionation we might expect these post-monsoon samples to trend back to heavier isotopic values. In 2007 this was the case, but in 2008 the data is more scattered. It is our belief that a second source water variation becomes important at this time that allows the late season rainfall to remain depleted.

5.3. The influence of the BoB freshwater plume on rainfall $\delta^{18}O$

We propose an additional source water influence on the isotopic composition of rainfall at our site during the late ISM. Massive fluvial runoff from the large river systems of the Ganges (Sengupta and Sarkar, 2006; Singh et al., 2007) and the Yarlungzangbo-Brahmaputra-Padma (Sarma, 2005) follows the intense rainfall and snowmelt on the Tibetan Plateau during the summer monsoon. This runoff is depleted in $\delta^{18}O$ and δD due to several factors; the high altitude of precipitation over the Himalaya, the greater rainout fraction during the monsoon, and the influence of the amount effect over land (Liu et al., 2008; Tian et al., 2001b). Meltwater from ice and snow might add to the δD and $\delta^{18}O$ depletion of this runoff. Overall a freshwater plume develops in the northern BoB during the ISM that is accompanied by isotopic dilution of the surface water $\delta^{18}O$ pool. The depleted $\delta^{18}O$ of runoff affects the BoB mixed layer $\delta^{18}O$ and may thus deplete the $\delta^{18}O$ of the initial vapor for subsequent rainfall. This 'plume effect' on source vapor will be most effective during late ISM when runoff is at its maximum (Sengupta and Sarkar, 2006).

To quantify the influence of the 'plume effect' on the isotopic composition of the BoB mixed layer we developed a salinity and isotope mass balance box model.

$$S_{ml} = S_{South}\frac{F_{South}}{F_{South} + F_{Riv}}$$

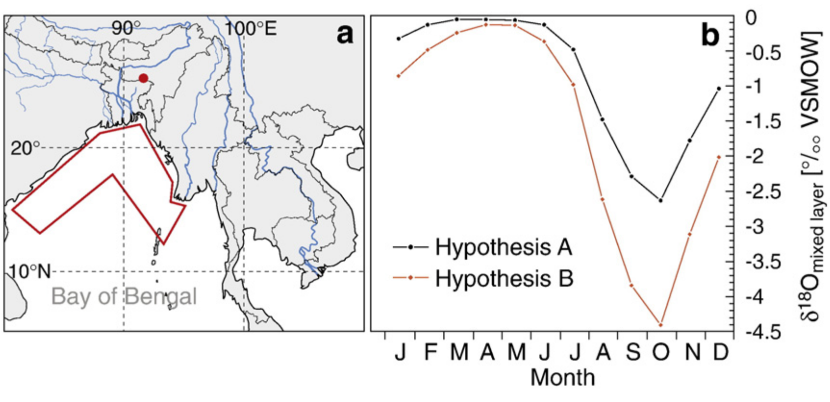

Fig. 6. a) Map of the BoB showing the area (630,050 km^2) of the box used for our isotope mass balance model. b) The modeled seasonal change of $\delta^{18}O$ in the mixed layer for the box in 6a. A hypothetical ISM runoff $\delta^{18}O$ minimum of −20‰ has to be invoked (red line, hypothesis B) which would deplete the mixed layer by ~4.5‰. An October runoff value of −12‰, consistent with the published data (Table 2), yields a surface depletion of −2.5‰. However, no data is available for monthly averaged $\delta^{18}O$ of runoff for the various rivers around the BoB to verify either scenario.

Table 2
Selected published $\delta^{18}O$ and δD values for eastern Indian rivers and the Bay of Bengal surface water. ISM values are depleted, relative to dry season values.

River	Location	Coordinates	Date	$\delta^{18}O$ (‰)	δD (‰)	Reference
Ganga	Rishikesh	30°7′N, 78°19′E	Mar-82	−9.2	−61	Ramesh and Sarin (1992)
			Sep-82	−11.5	−70	
	Downstream Patna	25°35′N, 85°14′E	Mar-82	−3.8	−39	Ramesh and Sarin (1992)
			Sep-82	−10.6	−73	Ramesh and Sarin (1992)
			Aug-01	−6.57	−47.56	Lambs et al. (2005)
	Padma/Bangladesh	24°12′N, 85°22′E	Apr-81	−5.3	–	Rozanski et al. (2001)
			May-81	−4.7	–	
			Jul-81	−6.0	–	
			Aug-81	−7.5	–	
			Jun-83	−6.6	–	
			Aug-83	−7.0	–	
			Oct-83	−7.7	–	
Manas	Goalpara	26°10′N, 90°37′E	Apr-82	−5.3	–	Ramesh and Sarin (1992)
Brahmaputra	Gauhati	26°09′N, 91°45′E	Apr-82	–	−39	Ramesh and Sarin (1992)
			Dec-82	–	−66	Ramesh and Sarin (1992)
			Aug-01	−11.97	−85.35	Lambs et al. (2005)
	Jamuna, Nagabari	25°45′N, 89°41′E	Apr-81	−7.2	–	Rozanski et al. (2001)
			May-81	−7.9	–	
			Jul-81	−9.1	–	
			Aug-81	−10.0	–	
Brahmaputra delta			Mean August	−8	–	Kudrass et al. (2001)
BoB surface		North of 15°N	Jan-Feb 94	−2	–	Delaygue et al. (2001)
		0-15°N		0-0.5	–	

$$\partial^{18}O_{ml} = \frac{\partial^{18}O_{South}F_{South} + \partial^{18}O_{Riv}F_{Riv}}{F_{South} + F_{Riv}}$$

S_{ml} and $\delta^{18}O_{ml}$ are the salinity and oxygen isotopic composition of the mixed layer region that we assume is the source water for vapor, which eventually precipitates at our site. The subscripts 'Riv' and 'South' refer to the runoff waters that enter the northern BoB and the open ocean waters south of our control volume for the mixed layer BoB, and 'F' refers to the respective flux. We treat the mixed layer region in the northern BoB as a single box that has boundaries shown in Fig. 6a. The salinity mass balance equation is used to calculate the F_{South} term. For a given relationship between $\delta^{18}O_{South}$ and S_{South}, the $\delta^{18}O$ mass balance equation is used to calculate the $\delta^{18}O_{ml}$. Results of this calculation for two different end-member $\delta^{18}O_{Riv}$ are shown in Fig. 6b. In hypothesis A, we use realistic modern values (in accordance with published river $\delta^{18}O$ values, Table 2, and similar to data from Indus river (Karim & Veizer, 2002)). A $\delta^{18}O_{Riv}$ value of −12‰ yields a seasonal difference of ~2.5‰. The resulting BoB surface water $\delta^{18}O$ change corresponds to approximately 20% of the observed range in $\delta^{18}O_{rainfall}$.

In hypothesis B, we assume extreme $\delta^{18}O$ values that could be expected during enhanced ISM rainfall in Tibet. In a warmer world scenario (like the Holocene Hypsithermal, Wanner et al. (2008), or under modern global warming), with possibly enhanced ISM rainfall in the Himalaya and on the Tibetan Plateau (high altitude precipitation), higher runoff and a larger extend of the BoB freshwater plume during late summer are expected. The plume effect would then be more pronounced ($\delta^{18}O_{surface\ water}$ is expected to be more depleted in such scenario). A maximum difference of ~4.5‰ in source water $\delta^{18}O$ between April and October derives from a hypothetical river end-member of −20‰ during the height of runoff in hypothesis B (Table 3). Such negative values for Tibetan river $\delta^{18}O$ have been reported recently by Hren et al. (2009).

The model is a simplified version of the box-model presented in Delaygue et al. (2001). These authors assumed a balance between E-P, runoff, and mixing across the thermocline. In effect we use the S_{South} and $\delta^{18}O_{South}$ end-members to account for the seasonal change in regional BoB E-P variations. Variations in this slope between $\delta^{18}O$ and salinity change the absolute value of the predicted $\delta^{18}O_{ml}$, but they do not change the seasonal amplitudes by an appreciable amount.

The dominant influence on source water $\delta^{18}O$ in the BoB is the river runoff signal and not the mixing with tropical waters to the south. The freshwater plume has its greatest extent during late ISM and early post-monsoon. Trajectory analysis shows, that during this time the distance of the storm tracks diminishes (Fig. 5c). Therefore, we assume that the influence of the 'plume effect' on rainwater $\delta^{18}O$ is largest during late ISM and allows the trend in lighter $\delta^{18}O$ of rainfall during the ISM to continue even as precipitation amount decreases. Potentially, the freshwater-flooded lowlands of Bangladesh and Bengal can serve as an additional (isotopically depleted) moisture source during late ISM. However, such depletion of the source region can only account for less than 25% of the amplitude measured at our site in 2007 and 2008. Our model might well underestimate the total discharge into the BoB, as well as the potential $\delta^{18}O$ depletion in river runoff, because both parameters are not well constrained by observations. The reliability of the input values for our hypothesis is difficult to determine, but the trend to lower δ-values is believed to be a robust feature of the system. Karim and Veizer (2002) presented $\delta^{18}O$ data from the Indus river with most depleted values during late

Table 3
Runoff and box model data. Two hypothetical runoff $\delta^{18}O$ values are used for box model calculations (hypotheses A and B). All given $\delta^{18}O$ values are in ‰ VSMOW. See main text for discussion. [a]Calculated monthly mean values from World Ocean Atlas data (NOAA, 2005); [b]Given after Sengupta and Sarkar (2006); [c]Estimated monthly values for the Bay of Bengal south of the box in Fig. 6a; and [d]Calculated using the $\delta^{18}O$–salinity relationship $\delta^{18}O = 0.16 \times S - 5.31$ given by LeGrande and Schmidt (2006) for the Indian Ocean.

Month	S_{ml} [psu] [a]	F_{Riv} [km^3/month][b]	S_{South} [psu] [c]	$\delta^{18}O^{d}_{South}$	Hypothesis A		Hypothesis B	
					$\delta^{18}O_{Riv}$	$\delta^{18}O_{ml}$	$\delta^{18}O_{Riv}$	$\delta^{18}O_{ml}$
1	31.34	100	33.20	0.00	−5	−0.33	−13	−0.86
2	32.16	40	33.50	0.05	−4	−0.13	−12	−0.49
3	32.70	20	33.30	0.02	−3	−0.05	−11	−0.25
4	32.68	80	33.00	−0.03	−3	−0.06	−11	−0.13
5	32.69	170	33.00	−0.03	−4	−0.07	−12	−0.14
6	32.57	350	33.50	0.05	−6	−0.13	−14	−0.37
7	31.87	530	33.70	0.08	−9	−0.48	−17	−0.98
8	30.23	590	33.80	0.10	−11	−1.48	−19	−2.61
9	29.31	460	33.50	0.05	−12	−2.29	−20	−3.84
10	29.76	300	33.40	0.03	−12	−2.63	−20	−4.40
11	30.47	170	33.60	0.07	−11	−1.78	−19	−3.11
12	30.87	140	33.60	0.07	−9	−1.04	−17	−2.02

ISM, which supports our hypothesis. However, future monitoring river and surface water isotopes in the BoB realm would be helpful.

The seasonal character of anticorrelation between fluvial runoff and BoB surface water $\delta^{18}O$ finds a long-term analogue in the Indian Ocean (Duplessy, 1982) and was used to reconstruct glacial-interglacial differences. We note, that the complex relationship between BoB surface water salinity and $\delta^{18}O$ is certainly influenced by hydrologic changes on longer (millennial) timescales, as found for the western Pacific (Oppo et al., 2007). Runoff into the BoB and the relative contribution of different headwaters changed significantly during the Holocene. Hence, the $\delta^{18}O_{surface\ water}$ might have varied considerably with changing inherited $\delta^{18}O_{runoff}$. Such isotopic variability is possible even if the total runoff into the BoB, and hence the salinity in the BoB, is assumed to have remained unchanged.

5.4. Surface water–vapor disequilibrium

Assuming progressively better storm organization and increased wind speed during the ISM (Narvekar and Prasanna Kumar, 2006), a further depletion in the $\delta^{18}O$ of rainfall might be expected. The observed high $\delta^{18}O$ values during the dry season may reflect relatively quiescent conditions over the BoB, which would allow for water vapor to isotopically equilibrate with the BoB surface water, similar to findings by Lawrence et al. (2004). During times of intense storm activity, these authors show that in the tropics, the $\delta^{18}O$ of water vapor is not in equilibrium with seawater and can be depleted by up to 15‰. This effect can be attributed to a switch of the substrate for vapor exchange from the ocean surface to falling water droplets. During stormy conditions, water vapor will be exposed to falling rain droplets, which are isotopically depleted and the vapor will more completely equilibrate with this secondary source as the water content of the air increases. By entraining this vapor back into the clouds, subsequent rainfall is then even further depleted. Very negative rainfall $\delta^{18}O$ is correlated with heavier rainfall and is more fully expressed when the total amount of water in the air is at a maximum during heavy storm intensity (Lawrence et al., 2004). However, this effect is expected to be most pronounced during maximum ISM and hence does not help to explain the most depleted $\delta^{18}O$ values during late-ISM, when storminess diminishes.

Taken together, the mechanisms listed above can explain, the large amplitude, the trend, and the delay of the $\delta^{18}O$ minima relative to maximum rainfall in our rainwater $\delta^{18}O$ data. A somewhat stronger depletion of $\delta^{18}O$ in rainwater can be expected for Meghalaya compared to the lowlands of West Bengal (Mukherjee et al., 2007), due to the altitude effect. Given the ~900 m of forced orographic uplift from the Bengal lowlands to our sampling site, the altitude effect could account for a ~ -1.35‰ to -4.5‰ depletion in the $\delta^{18}O$ of precipitation. Unfortunately, without data from multiple sites along the side of the plateau, evaluation of this effect has to wait for future monitoring.

6. Conclusions

A local meteoric water line and seasonal variations in the isotopic composition of precipitation for Meghalaya provide insight into the hydrological regime of northern Bangladesh and NE India. The effect of evaporation on the stable isotopes of ISM precipitation is not found in our dataset. Strong evaporation often affects dry season rainfall, with re-evaporation of up to 100% of the rain/drizzle (as mentioned above). However, our few dry-season rainwater samples do not show secondary evaporation, as they all are found close to the GMWL. Apparently, these samples fell through high-humidity air below the cloud.

In addition, there is no correlation between $\delta^{18}O$ and the amount of rainfall in our data. This implies that the amount effect has no significant influence on the isotopic signature of precipitation over Meghalaya/NE India, at least on event time scales.

The temporal changes in the stable isotopes of precipitation are pronounced and depend on the seasonal dynamics of moisture sources and transport pathways. We propose a) a shift from a northwestern continental moisture source (maintained during winter) to a southern marine source (during ISM), that leads to two distinct isotopic signatures in rainfall, b) that the distance of the marine moisture source increases during ISM from the BoB to the southwestern Indian Ocean, c) that seasonal runoff changes of the Ganges-Brahmaputra river system impact the isotopic composition of the BoB surface water, thus accounting for 20% to 50% of the depletion of $\delta^{18}O$ in subsequent rainfall over Meghalaya, and d) that disequilibrium conditions between surface water and water vapor during stormy monsoonal weather can further enhance the isotopic depletion.

Stable isotopes measured in palaeoclimatic archives, such as stalagmites, fed by moisture from the Indian Ocean are likely to record seasonal changes in ISM strength. Our proposed mechanisms for the trends over 2007–2008 should be valid over the Holocene epoch and are probably valid on glacial–interglacial timescales. Future rainfall reconstructions from speleothem archives will benefit from a detailed understanding of the history of BoB surface waters over the same time period.

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