1   Climatic and in-cave influences on $\delta^{18}O$ and $\delta^{13}C$ in a stalagmite from northeastern

2   India through the last deglaciation

3

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22

23

Abstract

25 Northeastern India experiences extraordinarily pronounced seasonal climate,

26 governed by the Indian Summer Monsoon (ISM). The vulnerability of this region to

27 floods and droughts calls for detailed and highly resolved paleoclimate

28 reconstructions in order to assess the recurrence rate and driving factors of ISM

29 changes. We use stable oxygen and carbon isotope ratios ($\delta^{18}O$ and $\delta^{13}C$) from

30 stalagmite MAW-6 from Mawmluh Cave to infer climate and environmental

31 conditions in northeastern India over the last deglaciation (16-6 kyr BP). We interpret

32 stalagmite $\delta^{18}O$ as reflecting ISM strength, while $\delta^{13}C$ appears to be driven by local

33 hydroclimate conditions. Pronounced shifts in ISM strength over the deglaciation are

34 apparent from the $\delta^{18}O$ record, similarly to other records from monsoonal Asia. The

35 ISM is weaker during the late glacial and the Younger Dryas, and stronger during the

36 Bølling-Allerød and Holocene. Local conditions inferred from the $\delta^{13}C$ record appear

37 to have changed less substantially over time, possibly related to the masking effect of

38 changing precipitation seasonality. Time series analysis of the $\delta^{18}O$ record reveals

39 more chaotic conditions during the late glacial, and higher predictability during the

40 Holocene, likely related to the strengthening of the seasonal recurrence of the ISM
41 with the onset of the Holocene.

42

43 **INTRODUCTION**

44 The Asian Monsoon (AM) is characterized by the seasonal reversal of circulation
45 between ocean and landmasses throughout South-East Asia, resulting in pronounced
46 hydroclimate seasonality in the affected regions. Stalagmite oxygen isotope ratio
47 ($\delta^{18}O$) records from the AM realm have provided crucial information on past climate
48 conditions in this densely populated region (e.g., Wang et al., 2005; Cheng et al.,
49 2016a; Eroglu et al., 2016). Pronounced glacial-interglacial variations in AM strength
50 are found to be strongly influenced by Northern Hemisphere Summer Insolation
51 (NHSI) (Cheng et al., 2009, 2016a; Kathayat et al., 2016), and closely related to
52 changes in the North Atlantic (Wang et al., 2001; Yuan et al., 2004). The AM is
53 therefore a highly dynamic system susceptible to external and internal forcings,
54 calling for precise paleoclimate reconstructions throughout monsoonal Asia to infer
55 on future developments under climate change scenarios.

56

57 The majority of precisely dated high-resolution reconstructions of past glacial-
58 interglacial AM variation stem from Chinese caves, providing unprecedented insight
59 in monsoonal dynamics over the past 640,000 years (Cheng et al., 2016a).
60 Information from the Indian subcontinent, particularly at high temporal resolution and
61 chronological precision, is still relatively scarce over these timescales (e.g., Sinha et
62 al., 2005; Govil and Divakar Naidu, 2011; Zhisheng et al., 2011; Menzel et al., 2014;
63 Kathayat et al., 2016). The Indian Summer Monsoon (ISM), the branch of the AM
64 that delivers moisture from the Arabian Sea and Indian Ocean to the Indian
65 subcontinent, as well as to the Arabian peninsula (Burns et al., 2002; Fleitmann et al.,
66 2007), and China (Zhisheng et al., 2011; Baker et al., 2015), delivers ~80% of the
67 annual rainfall of these regions, and dominates their hydrological cycle (Sinha et al.,
68 2007, Breitenbach et al. 2010) (Fig. 1A).

69

70 Stalagmite $\delta^{18}O$ is a widely applied proxy for monsoonal strength in Asia, and is
71 interpreted as reflecting the $\delta^{18}O$ of precipitation (Breitenbach et al. 2010, 2015;
72 Pausata et al., 2011; Cheng et al., 2016b; Eroglu et al., 2016). As the $\delta^{18}O$ signal is

governed by a multitude of factors, their relative importance at a specific location needs to be carefully assessed in order to correctly interpret paleoclimatic data. Isotopic composition of the moisture source, transport length, and the amount of precipitation at the site can all affect precipitation $\delta^{18}O$ in monsoonal regions, and taken together they provide information on monsoonal strength (Breitenbach et al., 2010; Baker et al., 2015; Eroglu et al., 2016). Cave monitoring efforts and simultaneous study of different stalagmite geochemical proxies in parallel often allow to more clearly determine the controls on local climate conditions, leading to more accurate paleoclimate reconstructions (Baldini, 2010; Oster et al., 2012; Breitenbach et al., 2015; Baldini et al., 2016; Cheng et al., 2016b). Stable carbon isotope ratios ($\delta^{13}C$) are routinely measured together with $\delta^{18}O$, but have so far rarely been reported for records from monsoonal Asia. This is partly due to the more complicated and site-specific interpretation of stalagmite $\delta^{13}C$, which necessitates thorough understanding of the local conditions; $\delta^{13}C$ can be influenced by vegetation composition (i.e., C_3 vs. C_4 plants), soil processes, open vs. closed conditions in the karst during carbonate dissolution, and isotope fractionation in or above the cave (Fairchild and Baker, 2012). However, carefully evaluated stalagmite $\delta^{13}C$ time series can provide important climate information to supplement and extend the interpretation of $\delta^{18}O$ records, often resulting in a more in-depth understanding of past climate conditions (Genty et al., 2003; Cosford et al., 2009; Ridley et al., 2015; Cheng et al., 2016b). Particularly interesting is the difference in spatial scale between both proxies: while $\delta^{18}O$ generally reflects large-scale atmospheric circulation processes (Breitenbach et al., 2010; Baker et al., 2015), $\delta^{13}C$ is a proxy for local processes and therefore more sensitive to changes at local to regional level (Ridley et al., 2015; Cheng et al., 2016b).

Here we present new sub-decadally resolved $\delta^{18}O$ and $\delta^{13}C$ data from a stalagmite from northeastern (NE) India that covers the interval of the last deglaciation and early Holocene (~16-6.5 ky BP). The last deglaciation was a period of rapid and substantial global climate change, driven by a ~3.5°C increase in global temperatures (Shakun et al., 2012). This resulted in large-scale reorganizations of circulation and weather patterns globally, with important repercussions in monsoonal Asia (Dykoski et al., 2005; Cheng et al., 2009; Ma et al., 2012). This analysis follows long-term studies of

106   the controls on precipitation $\delta^{18}O$ (Breitenbach et al., 2010), as well as detailed cave

107   microclimate monitoring schemes (Breitenbach et al., 2015), which allow us to

108   disentangle local and regional responses to climate change over the last deglaciation.

109

110   **GEOGRAPHICAL AND CLIMATOLOGICAL SETTING**

111   Mawmluh Cave is located at 25°15'44"N, 91°52'54"E, 1320 m above sea level on the

112   Meghalaya plateau in NE India (Fig. 1). The cave developed in a Tertiary limestone

113   butte at the southern fringe of the plateau (Ghosh et al., 2005; Gebauer, 2008), and is

114   today mainly covered by grassland. Mean annual air temperature inside the cave

115   (18.5°C) is very similar to that recorded at the meteorological station Cherrapunji

116   (17.4°C) and in the nearby Mawmluh village (19.1°C).

117

118   Hydroclimate in Meghalaya is extremely seasonal, with ~80% of annual precipitation

119   falling during the ISM season (June-October; Breitenbach et al., 2010). The

120   Meghalaya Plateau is the first morphological barrier for northward-moving moisture

121   from the Bay of Bengal (BoB), inducing intense orographic rainfall. Thus, Meghalaya

122   is a major water source for the Bangladesh plains, a region frequently flooded during

123   summer, e.g., in 1998, when ~60% of the country was inundated (Murata et al., 2008;

124   Webster, 2013). Despite having the highest rainfall amount in the world (Prokop and

125   Walanus, 2003), low retention capacity, due to the geological conditions on the

126   southern Meghalaya Plateau, results in frequent water shortage during the dry season

127   (November-May).

128

129   **MATERIALS AND METHODS**

130   **Stalagmite MAW-6**

131   The 21 cm long stalagmite MAW-6 was found broken in Mawmluh Cave in 2009 and

132   its original position is known only approximately (Fig. 1). The stalagmite displays

133   complex brown-grey color variations, with bands up to a few mm wide, but no annual

134   laminations. At least three white layers can be discerned which span from the growth

135   axis towards the sides of the stalagmite. To verify the mineralogy in MAW-6, three

136   samples were analyzed by X-ray diffraction (XRD) using a powder XRD

137   diffractometer (Bruker, D8 Advance), equipped with a scintillation counter and an

138   automatic sampler at ETH Zurich, Switzerland.

U-series dating and chronology development

After cutting the stalagmite lengthwise using a diamond stone saw, 24 U-series samples with weights between 88 and 311 mg were milled using an ethanol-cleaned stainless steel bit, and subsequently analyzed by multi-collector inductively coupled plasma mass spectrometry using a Thermo-Finnigan Neptune in the Minnesota isotope laboratory, University of Minnesota. The chemical procedures used to separate uranium and thorium for U-series dating are similar to those described in Edwards et al. (1987). Uranium and thorium isotopes were analyzed on the multiplier behind the retarding potential quadrupole in peak-jumping mode. Instrumental mass fractionation was determined by measurements of a $^{233}U/^{236}U$ spike. The detail techniques are similar to those described in (Cheng et al., 2000, 2009) and half-life values are those reported in (Cheng et al., 2013).

The age model for MAW-6 was computed for each growth segment by applying a cubic interpolation procedure using the COPRA software (Breitenbach et al., 2012). COPRA computed 2000 ensemble realizations for both the $\delta^{18}O$ and $\delta^{13}C$ records, from which the median, i.e., the central age for a defined sample depth, was calculated. The uncertainty in the age model is defined by the 95% confidence intervals, derived using the $\pm 2\sigma$ deviation from the median (Breitenbach et al., 2012).

Stable isotope analysis

1050 samples for stable isotope analysis were milled continuously at 200 μm resolution using a semi-automated high-precision drill (Sherline 5400 Deluxe) at ETH Zurich. Nine Hendy tests were performed over the length of the stalagmite to look for signs of kinetic isotope fractionation effects and, if present, to evaluate potential changes in the intensity of kinetic fractionation through time (Suppl. fig. 1). For this, carbonate samples were drilled point-wise along a single layer of the stalagmite using a 0.3 mm diameter drill bit.

Samples were analyzed for $\delta^{18}O$ and $\delta^{13}C$ on a Delta V Plus mass spectrometer coupled to a ThermoFinnigan GasBench II carbonate preparation device at the Geological Institute, ETH Zurich (Breitenbach and Bernasconi, 2011). An in-house

172   carbonate standard (MS2), which is well linked to NBS19 (Breitenbach and

173   Bernasconi, 2011), was used to evaluate the runs. All values are expressed in permil

174   (‰) and referenced to the Vienna PeeDee Belemnite (V-PDB) standard. The external

175   standard deviation (1σ) for both, $\delta^{18}O$ and $\delta^{13}C$ analyses on the carbonate is smaller

176   than 0.07‰.

177

178   Because the MAW-6 record covers the period of the last deglaciation, the contribution

179   of changes in both sea surface temperature (SST) and sea level due to the melting of

180   continental ice sheets to stalagmite $\delta^{18}O$ must be considered. To estimate this

181   contribution, a linear interpolation of seawater $\delta^{18}O$ values ($\delta^{18}O_{seawater}$) reconstructed

182   from a sediment core from the BoB (Rashid et al., 2011) was performed to fit the

183   MAW-6 data points. The $\delta^{18}O_{seawater}$ record was subsequently subtracted from the

184   measured $\delta^{18}O_{calcite}$ in MAW-6, to yield an ice volume corrected ($\delta^{18}O_{IVC}$) record

185   (Suppl. Fig. 2). It should be noted that this procedure can introduce artefacts, as the

186   records are irregularly sampled, and the results should be interpreted with care.

187

188   **Recurrence quantification analysis**

189   To infer possible changes in the dynamical regime of the ISM between late glacial

190   and early Holocene conditions, we performed a statistical analysis considering the

191   deterministic nature of the underlying process (the ISM), encoded by the recurrence

192   properties of the $\delta^{18}O_{IVC}$ record. We use a measure of complexity, called *recurrence*

193   *determinism* (DET), which is derived from a recurrence plot, a graphical, binary

194   representation of pairs of times of similar values (actually states) within the time

195   series (see Supplemental information for further details; Marwan et al., 2007; Ozken

196   et al., 2015; Eroglu et al., 2016). DET reveals high values for deterministic processes

197   and regular (e.g., cyclic, periodic) variations, whereas more stochastic (i.e., random)

198   dynamics lead to low DET values. Moreover, the recurrence analysis is combined

199   with the pre-processing TACTS technique that allows detrending regularization of

200   irregularly sampled time series (see Supplemental information; Ozken et al., 2015).

201

202   The $\delta^{18}O_{IVC}$ record was divided into two periods, the late glacial (LG, 16-13 kyr BP)

203   and the early Holocene (EH, 9-6.5 kyr BP) and the DET measure were calculated for

204   both periods separately (Fig. 8). A statistical test based on a bootstrap approach was

205 performed to evaluate the significance of the variations in the DET measure. This test

206 provides a cumulative probability distribution of DET measures corresponding to the

207 null-hypothesis that there is no change in the dynamics of the underlying climate

208 process. From this test distribution the upper 95% confidence limit can be defined.

209

210 **RESULTS**

211 **Petrography and Mineralogy**

212 XRD analysis reveals that stalagmite MAW-6 consists of calcite. The white layers

213 described above were identified as dirt layers in the stalagmite. They are well visible

214 at the fringes of the respective layers, while the stalagmite tip has been washed clean

215 by impinging water droplets (Fig. 1C). The deposition of silty material on the

216 stalagmite surface at these depths might indicate burial of MAW-6 by sediment re-

217 deposition in the cave, inhibiting further growth. Several buried stalagmites have been

218 located in the cave (Suppl. Fig. 3) and sediment migration within the cave passage

219 appears to be an important process during high-discharge events of the cave stream.

220 However, caution must be applied with this interpretation, since dirt layers can also

221 originate from other processes, such as aerosol and dust deposition.

222

223 **Age model**

224 MAW-6 grew between ~16 and 6.5 kyr BP. The age model is based on 20 U-series

225 dates, with analytical errors between ±16 and ±264 years (Fig. 2 and Table 1). Four

226 dating samples contained high amounts of detrital thorium and were excluded from

227 the final age model (shown in red in Fig. 2).

228

229 Three hiatuses were identified in the depth-age relationship, coinciding with the white

230 dirt layers in the stalagmite. User-specified hiatus depths of 69.46 mm, 111.86 mm

231 and 147.26 mm from the stalagmite top allowed COPRA to split the age model

232 construction into independent age models (before and after the hiatuses respectively).

233 This procedure yielded a segmented depth-age chronology for the stalagmite, with

234 hiatuses at 10.4-9.6 kyr BP, 11.6-10.8 kyr BP and 13-12.4 kyr BP. The details for the

235 age modeling procedure can be found in Breitenbach et al. (2012).

236 Using the COPRA procedure, the age uncertainties of the MAW-6 record can be

237 transferred from the age to the proxy domain (Breitenbach et al., 2012), which results

238   in a 95% confidence interval of possible proxy values at a given point in time. As a

239   consequence, it is not possible to determine the high-frequency variations within the

240   bounds of the confidence interval (Fig. 3A). Comparative discussions to other

241   paleoclimate records from the AM realm are therefore restricted to the median proxy

242   values in MAW-6 derived from the COPRA Monte Carlo modeling and to long-term

243   centennial changes. We still show the original MAW-6 isotope data for a tentative

244   comparison with other available records in Fig. 7, as this kind of uncertainty is

245   common to all paleoclimate records.

246

247   **The MAW-6 δ^{18}O and δ^{13}C records**

248   The δ^{18}O profile varies from -8.3‰ to -2.8‰ (Fig. 3A), with heavier values found in

249   the oldest part of the record (end of the last glacial), and lighter values in the youngest

250   part (the Holocene). Both, the Bølling-Allerød (B-A) interstadial, beginning at ~14.5

251   kyr BP with -1.5‰ shift, and the Younger Dryas (YD), between 12.6 and 11.6 kyr BP

252   and featuring the heaviest values of the entire record (~-3.5‰), are clearly

253   demarcated. However, the exact beginning and the end of the YD in MAW-6 cannot

254   be defined, since the interval is bracketed by two hiatuses. The 850-year long hiatus

255   that masks the end of the YD is followed by a substantial 4.5‰ decrease in δ^{18}O,

256   marking the transition into the Holocene (~9.6 kyr BP). This decrease occurs in two

257   rapid stages, characterized by hiatuses, interrupted by an interval (~10.8 - 10.2 kyr

258   BP) of relatively constant intermediate values (~-6.5‰). Lowest δ^{18}O values (~-8‰)

259   are found during the early Holocene (~9.6 kyr BP), slightly increasing towards the

260   youngest part of the record. Our sea-level corrected δ^{18}O$_{IVC}$ record shows that ice

261   volume and SST changes affect the isotope signature mainly before the YD, with only

262   minor impacts during the Holocene, accounting for ~1/4 (1‰) of the shift between

263   deglaciation and Holocene (Fig. 3A). We are therefore confident that the larger part

264   (3‰) of the variation in δ^{18}O during the deglaciation is attributable to changes in ISM

265   strength. For the following discussion, only the δ^{18}O$_{IVC}$ record is considered.

266

267   Compared to the δ^{18}O profile, the δ^{13}C profile is much more uniform, with variations

268   ranging between -1.2 and -6.6‰ and without clear trends over time. The heaviest

269   values are found during the early part of the record (late glacial-B-A, average -4‰).

270   In contrast to δ^{18}O, the YD period is characterized by slightly lighter values than

271 during the late glacial-B-A section (average -4.3‰), whereas the transition into the

272 Holocene leads to the most negative values (average -5.2‰ post ~9.6 kyr BP). The

273 high-frequency variations in $\delta^{18}O$ and $\delta^{13}C$ are remarkably similar, but shifts in $\delta^{13}C$

274 are generally much more pronounced than in $\delta^{18}O$. The similarity between the two

275 records is also reflected by their high correlation (during all periods $r > 0.55$, Fig. 3B).

276

277 A crossplot of $\delta^{13}C$ vs. $\delta^{18}O_{IVC}$ reveals four distinct clusters (Fig. 4). These clusters

278 are mainly influenced by the average $\delta^{18}O_{IVC}$ during the different periods, therefore

279 we distinguish a Holocene, intermediate (10.8-10.2 kyr BP), YD, and B-A/late glacial

280 cluster. The boxplot representation of the datasets in Fig. 4B allows the quantification

281 of temporal and proxy-related differences. While a clear distinction of the different

282 time periods is apparent in the $\delta^{18}O_{IVC}$ dataset, a much larger spread in $\delta^{13}C$ values is

283 found (Fig. 4B). The YD cluster is characterized by the heaviest $\delta^{18}O_{IVC}$ values of the

284 entire record, while $\delta^{13}C$ is slightly lighter than during the late glacial. A trend

285 towards progressively lighter $\delta^{18}O_{IVC}$ values is found between the YD, intermediate,

286 and Holocene clusters, while intermediate $\delta^{13}C$ values are slightly heavier than during

287 the YD (Fig. 4B).

288

289 The Hendy tests carried out throughout MAW-6 show evidence for kinetic

290 fractionation during the YD and late glacial, while (near-)equilibrium conditions seem

291 to have prevailed during the B-A and the Holocene (Suppl. Fig. 1). Kinetic effects are

292 identified by strong correlations between $\delta^{18}O$ and $\delta^{13}C$, as well as enrichment in the

293 heavy isotopes with increasing distance from the growth axis (Hendy, 1971).

294

295 **Determinism of the $\delta^{18}O$ record**

296 The analysis reveals distinctly different DET measures for the LG (DET=0.663 and

297 inside the confidence interval) and the EH (DET=0.736, and outside the confidence

298 interval) (Fig. 8). A high DET measure indicates a more predictable, i.e., a less

299 chaotic, regime, while the opposite holds true for low DET measures.

300

301 **DISCUSSION**

302 **The influence of karst processes on stalagmite stable isotopes in Mawmluh Cave**

303   Karst processes at Mawmluh Cave are driven by the seasonal cycle in regional

304   hydrology (Fig. 5). Precipitation $\delta^{18}O$ becomes increasingly lighter during the ISM

305   months, and reaches the most negative values during the late and post ISM (August-

306   October) (Breitenbach et al., 2010). A direct amount effect can therefore be ruled out,

307   as maximum precipitation occurs earlier in the ISM season (July-August). Instead,

308   precipitation $\delta^{18}O$ in Meghalaya is controlled by: i) the travel distance of the air

309   masses, which increases throughout the ISM season, promoting stronger Rayleigh

310   fractionation during transport and lighter $\delta^{18}O$ at the site (Breitenbach et al., 2010), ii)

311   stronger contribution from isotopically depleted freshwater delivered to the BoB

312   during the late ISM (Sengupta and Sarkar, 2006; Singh et al., 2007; Breitenbach et al.,

313   2010), and iii) isotopic depletion of rainwater during large rainstorms (amount effect

314   *sensu* Dansgaard, 1964) (Lawrence et al., 2004; Breitenbach et al., 2010; Baker et al.,

315   2015). These mechanisms all drive precipitation $\delta^{18}O$ in the same direction, resulting

316   in lighter $\delta^{18}O$ during and after the ISM, and heavier values during dry season months

317   (Breitenbach et al., 2015; Myers et al., 2015). At Mawmluh Cave, infiltration is

318   strongly skewed towards the summer months, and consequently dripwater $\delta^{18}O$ is

319   biased towards the ISM season (Fig. 5). Still, a clear seasonal cycle in dripwater $\delta^{18}O$

320   is observed, with the lightest values occurring during the late ISM months, indicating

321   rapid (<1 month) fluid transfer into the cave (Breitenbach et al., 2015) (Fig. 5).

322   Dripwater (and stalagmite) $\delta^{18}O$ at Mawmluh Cave can therefore be used as a reliable

323   ISM strength proxy.

324

325   Dripwater $\delta^{13}C$ can be influenced by changes in vegetation type above the cave (C_3

326   vs. C_4 plants; Denniston et al., 2001), soil activity (Genty et al., 2006; Scholz et al.,

327   2012), bedrock dissolution and open vs. closed system conditions in the karst (Genty

328   et al., 2001), and prior calcite precipitation (PCP) and fractionation processes in the

329   cave (Griffiths et al., 2012; Ridley et al., 2015). At Mawmluh Cave, precipitation and

330   consequently vegetation and soil activity (microbial activity and root respiration) are

331   at a maximum during the summer months (June-October), resulting in highest relative

332   humidity and soil pCO_2 during this period. The extremely high amounts of rainfall

333   delivered at Mawmluh Cave during the ISM season (max. 13,472 mm between June

334   and September; Breitenbach et al., 2015) lead to waterlogging of the soil and karst

335   overlying the cave (Breitenbach et al., 2015), most likely resulting in more closed

336   system conditions. Therefore, prior carbonate precipitation (PCP) in epikarst and cave

337   is minimized (or even completely absent) during the ISM season. Strong seasonal

338   variations in cave air pCO_2 are observed as a consequence of seasonal ventilation

339   changes (Breitenbach et al., 2015) (Fig. 5). During the dry season months, low cave

340   air pCO_2 due to strongly reduced rainfall amount above the cave and intensified

341   ventilation, leads to enhanced degassing of CO_2 from the solution, enriching

342   dripwater in ^{13}C (Breitenbach et al., 2015). Moreover, open system conditions prevail

343   in the overlying soil and karst, due to seasonal aridity, resulting in low soil activity

344   (less input of isotopically light organic carbon to soil water) and promoting PCP (Fig.

345   5). All factors taken together, conditions during the dry season result in heavier

346   dripwater $\delta^{13}C$ in Mawmluh Cave, whereas the opposite holds true for the ISM

347   months. Dripwater and stalagmite $\delta^{13}C$ is therefore strongly influenced by effective

348   infiltration in the soil and tightly connected to local climate conditions.

349

350   **Interpretation of the MAW-6 isotope records**

351   We find large variations in $\delta^{18}O_{IVC}$ in stalagmite MAW-6 over the period of the last

352   deglaciation, with the heaviest values recorded during the late glacial and YD (Fig. 3).

353   Considering the controls on precipitation and dripwater $\delta^{18}O_{IVC}$ at Mawmluh Cave, we

354   interpret the late glacial and YD portions of the record as periods of weaker/shorter

355   ISM, accompanied by changes in the circulation regime (i.e., a more proximal

356   moisture source), whereas stronger ISM and longer moisture transport paths prevailed

357   during the B-A and the Holocene. Changes in both sea surface temperature (SST) and

358   sea level due to the melting of the continental ice sheets during the last deglaciation

359   resulted in substantial alteration of the isotopic composition of the surface ocean

360   water, as well as affecting evaporation and convection from the sea surface (Gadgil,

361   2003). The changes in the moisture source affect precipitation and stalagmite $\delta^{18}O$. In

362   the BoB, the moisture source for the ISM, a ~3.2-3.5°C increase in SST between the

363   Last Glacial Maximum and the Holocene, and a +1.4°C SST shift between the YD

364   and the Holocene, have been documented (Rashid et al., 2007, 2011; Govil et al.,

365   2011). In this region, additional depletion of seawater ^{18}O occurred most likely due to

366   freshening of the BoB by increased runoff from precipitation and glacier melt in the

367   Himalaya and Tibet.

368

369   The millennial-scale average in MAW-6 $\delta^{13}C$ shows much lower variability than

370    $\delta^{18}O_{IVC}$ over the last deglaciation, but the centennial-scale variations are remarkably

371   similar (Fig. 3). It is likely that changes in vegetation density and composition

372   occurred between cold/dry glacial and warm/humid interglacial periods. However,

373   vegetation changes above the cave as the primary cause for the high frequency

374   variation in $\delta^{13}C$ can probably be ruled out, as these would require longer time

375   periods and would likely be more gradual than the rapid decadal-scale shifts we find

376   in MAW-6. Karst processes, namely PCP and kinetic fractionation in the cave, can

377   best explain the observed variation in MAW-6 $\delta^{13}C$. We find heavier $\delta^{13}C$ values

378   during weak ISM periods, as identified in the $\delta^{18}O_{IVC}$ record, indicating enhanced PCP

379   and kinetic fractionation stemming from drier summer and/or longer winter seasons.

380   Periods of strong ISM, on the other hand, are characterized by lighter $\delta^{13}C$ values,

381   which is in line with more closed system conditions during the wet summers, higher

382   cave air pCO_2, which subdues kinetic fractionation, and more active vegetation and

383   soil.

384

385   It is possible that kinetic processes affect stalagmite $\delta^{18}O$ as well, precluding

386   quantitative rainfall reconstructions, but still allowing qualitative interpretation of

387   monsoon strength. In fact, kinetic fractionation would drive stalagmite $\delta^{18}O$ towards

388   more positive values, as prolonged degassing and possibly evaporation enrich the

389   precipitating solution in the heavy isotope, thus increasing the sensitivity of the

390   speleothem to record dry periods. Modern dripwater $\delta^{18}O$ values directly reflect

391   precipitation $\delta^{18}O$ values at the site, lending additional confidence to the interpretation

392   of stalagmite $\delta^{18}O$ as a monsoon strength proxy. Periods of enhanced kinetic

393   fractionation in the past can be detected using the Hendy tests. Evidence for kinetic

394   fractionation is observed during the YD and the late glacial, periods that we interpret

395   as drier, while (near-)equilibrium conditions seem to have prevailed during the B-A

396   and the Holocene, when conditions were wetter (Suppl. Fig. 1). These results have to

397   be interpreted with care, however, as sampling along a single growth layer is

398   extremely difficult when no annual laminae are present.

399

400   We can use the complementary information of $\delta^{18}O_{IVC}$ and $\delta^{13}C$ in stalagmite MAW-6

401   to interpret climate variations on supra-regional as well as local scale over the last

402 deglaciation. Clear shifts in average $\delta^{18}O_{IVC}$ are apparent during different time periods

403 (clusters in Fig. 4), indicating changing ISM strength, related to the moisture source

404 and composition upstream of the study site. While shifts in $\delta^{13}C$ are less strongly

405 expressed, it is still possible to distinguish periods of local aridity/humidity related to

406 the amount of effective infiltration in the karst and cave ventilation dynamics.

407 Positive correlation between $\delta^{18}O_{IVC}$ and $\delta^{13}C$ indicates that in general, weaker ISM

408 conditions are reflected as locally drier conditions at the study site, either due to

409 reduced summer rainfall, or a prolonged dry season (Fig. 4). "Weak-ISM" periods

410 (late glacial, YD) are also characterized by a tendency towards heavier $\delta^{13}C$,

411 suggesting drier conditions at the cave site, whereas Holocene $\delta^{18}O_{IVC}$ and $\delta^{13}C$

412 clearly cluster at lighter values for both proxies, indicating strong ISM and humid

413 conditions at the study site.

414

415 However, a detailed analysis of the relationship between $\delta^{18}O_{IVC}$ and $\delta^{13}C$ suggests

416 that the connection between local climate and large-scale ISM dynamics might be

417 more complex. The YD is clearly defined as the cluster with heaviest $\delta^{18}O_{IVC}$ values,

418 suggesting a more proximal moisture source with little freshwater influence from

419 riverine runoff, and an overall weakened ISM circulation (Fig. 4B). Local

420 hydroclimate conditions (indicated by $\delta^{13}C$), on the other hand, appear to have been

421 rather similar to those during the preceding B-A, but more arid than during the

422 succeeding Holocene. This apparent inconsistency (weaker ISM, without increased

423 aridity at local level) reflects the different controls on $\delta^{18}O_{IVC}$ and $\delta^{13}C$, where

424 changes in the moisture source and composition do not necessarily always influence

425 local infiltration directly (Cheng et al., 2016b). While $\delta^{18}O_{IVC}$ is influenced primarily

426 by the ISM during summer months, $\delta^{13}C$ is more sensitive to dry conditions, i.e., the

427 arid winter months. However, the dry season months in Meghalaya are characterized

428 by very dry conditions at present, and it is unlikely that conditions during the YD

429 were much different (as drier than dry is impossible). It is thus likely that a change in

430 precipitation seasonality during the YD led to a weaker ISM with a more proximal

431 rainfall source during the summer months (i.e., heavier $\delta^{18}O_{IVC}$), and at the same time

432 a more even distribution of rainfall over the year, resulting in reduced seasonality and

433 little effective change in karst processes (i.e., lighter $\delta^{13}C$).

434

Comparison to other records from Mawmluh Cave

To test if MAW-6 indeed reflects climate variations and not just local effects we compare the MAW-6 $\delta^{18}O_{IVC}$ record to the KM-A record (Berkelhammer et al., 2012) and the MWS-1 record (Dutt et al., 2015) from the same cave (Fig. 6). We find good visual replication between the three records on centennial time scale when recalculating the age models for KM-A and MWS-1 using COPRA (Fig. 6). The absolute difference in $\delta^{18}O$ values, especially pronounced between MAW-6 and MWS-1, is likely related to varying degrees of isotopic fractionation at different drip sites in the cave (similar to e.g., Stoll et al., 2015). For more quantitative information, the three time series were interpolated to annual resolution and low-pass filtered in order to only consider centennial time scale variations. Correlations were then calculated by downsampling the data to 50-year resolution. With this approach, we find high positive correlations between MAW-6 and KM-A during the period 6.9-9 kyr BP ($r = 0.93$), as well as between 9-12.4 kyr BP ($r = 0.78$, discarding the time periods corresponding to hiatuses in MAW-6). Similarly, correlation between MAW-6 and MWS-1 is positive ($r = 0.89$). All relationships are highly significant ($p < 10^{-9}$). Overall, this comparison corroborates our interpretation that variations in $\delta^{18}O_{IVC}$ in MAW-6 are driven by climate. The high resolution and the precise chronology of our record could significantly improve the available data from the ISM realm.

Comparison to other AM records

We chose three high-resolution and precisely dated records from Chinese caves (Dongge, Dykoski et al., 2005; Yamen, Yang et al., 2010; Kulishu, Ma et al., 2012), the MWS-1 record from Mawmluh Cave, and the NGRIP ice core record from Greenland (Andersen et al., 2004) to compare with our MAW-6 $\delta^{18}O_{IVC}$ record (Fig. 7). Comparison of the MAW-6 $\delta^{18}O_{IVC}$ record to these reconstructions reveals very similar centennial-millennial scale trends over the last deglaciation, further corroborating our interpretation of the record as a proxy for ISM strength (Fig. 7). However, more subtle differences are apparent as well. For example, whereas the other AM reconstructions indicate the weakest summer monsoons during the last glacial (until ~14.5 kyr BP), reflecting the pattern found in Greenland ice cores, MAW-6 records weakest ISM conditions during the YD (Fig. 7). This is partly related to the adopted correction for ice volume and SST, which results in lighter $\delta^{18}O_{IVC}$

468   during the last glacial, but the pattern is also apparent in the original $\delta^{18}O$ record (Fig.

469   3). This is possibly a reflection of changes in regional seasonality in NE India, with a

470   less vigorous ISM fed from proximal moisture sources, together with a wider spread

471   of precipitation over the entire year during the YD. Testing this hypothesis requires

472   seasonally resolved time series with highly robust chronologies.

473

474   In addition, differences appear when comparing the B-A interstadial in MAW-6 to

475   other records. Whereas the AM records considered here all show a relatively rapid

476   transition at the beginning and the end of the interval, with a plateau of lighter $\delta^{18}O$

477   values during the B-A (attributed to increasing insolation; Ma et al., 2012), MAW-6

478   shows a pattern of rapid isotopic depletion at ~14.5 kyr BP followed by a gradual

479   increase towards YD values that is more similar to the transition recorded in

480   Greenland ice cores (Fig. 7). This might hint towards a close connection between NE

481   India and the North Atlantic realm, driven by the Westerlies. Evidence from

482   paleoclimate records from Central Asia suggests that the AM and Westerly climates

483   are tightly connected over glacial-interglacial cycles (Cheng et al., 2016b). Mawmluh

484   Cave is located close to the Tibetan Plateau, with frequent influence of dry air masses

485   from the Tibetan High during the winter season, and a closer connection to the

486   Westerly climate than found at the Chinese cave sites is thus plausible. However, this

487   interpretation needs to be cautiously evaluated, due to limited replication with the

488   other $\delta^{18}O$ record from Mawmluh Cave covering the B-A interval (MWS-1; Dutt et

489   al., 2015), possibly related to chronological uncertainties in both records at this time.

490

491   The transition into the Holocene in MAW-6, although interrupted by a hiatus, shows

492   substantially lighter $\delta^{18}O$ values and thus ISM strengthening over time, similar to the

493   record from Yamen Cave (Yang et al., 2010) and well-replicated in MWS-1.

494   Conversely to the gradually lighter $\delta^{18}O$ values found in the Yamen and Dongge cave

495   records, however, both reconstructions from Mawmluh Cave show a short plateau of

496   intermediate values between ~10.2 and 10.8 kyrs BP. In MAW-6, this interval is

497   demarcated by two hiatuses, and therefore direct comparison to other records is

498   difficult. This feature in the $\delta^{18}O$ records could reflect slow retreat or even a short-

499   lived advance of the Himalayan glaciers, related to the increase in moisture and

500   precipitation (strengthening ISM) at the onset of the Holocene (Meyer et al., 2009).

The glaciation in the mountain range and related cold air outflow from the Himalaya mountain range would have hampered the intrusion of the ISM somewhat longer in NE India. This explanation remains hypothetical however, especially due to the scarcity of data from the region.

Dynamical changes in the ISM

Dynamical regime changes in the ISM between late glacial and early Holocene were investigated using recurrence quantification analysis (Ozken et al., 2015; Eroglu et al., 2016). We find a significant regime transition between LG and EH ISM, with more chaotic conditions during the LG, but higher predictability during the EH (Fig. 8). Disruption and weakening of the ISM during the LG, with frequent influence from Westerly air masses and the Tibetan High, very likely result in less predictable conditions. This is similar to findings from complex network analysis of the AM, where weaker supra-regional links were found during the cold/dry Little Ice Age (100-400 yr BP), suggesting that a weaker ISM is less predictable on a regional scale (Rehfeld et al., 2012). During the EH, on the other hand, the strong seasonality induced by the ISM would lead to more regular annual cycles in precipitation, and to a higher predictability.

CONCLUSIONS

Stalagmite MAW-6 provides new paleoclimate data from Mawmluh Cave in NE India, covering the last deglaciation. We combine decadal-scale $\delta^{18}O$ and $\delta^{13}C$ measurements on MAW-6 to unravel climate change at regional and local scale over this period. A substantial post-glacial shift towards more negative $\delta^{18}O$ values is interpreted as strengthening of the ISM, with maximum expression during the early Holocene. This pattern is in agreement with other reconstructions from Mawmluh Cave and the AM realm. Both the B-A and YD periods are clearly demarcated in the record as stronger and weaker ISM, respectively. $\delta^{13}C$ is interpreted as reflecting local hydroclimatic conditions, and is generally similar to $\delta^{18}O$, suggesting that a weak/strong ISM results in drier/wetter conditions at the study site. An intriguing exception to this rule is the YD, where combined $\delta^{18}O$ and $\delta^{13}C$ analysis suggests a reduction in precipitation seasonality, together with weakening of the ISM. Statistical time series analysis of the $\delta^{18}O$ record reveals a significant regime transition over the

last deglaciation, with less predictable ISM during the late glacial, and higher predictability during the Holocene, which we relate to the build-up of strong precipitation seasonality induced by the ISM.

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804

805   **Tables**

806   Table 1: U-series dating results for stalagmite MAW-6. The errors given are 2σ.

807    $^*\delta^{234}U = ([^{234}U/^{238}U]_{activity} -1)\times1000$. $^{**}\delta^{234}U_{initial}$ was calculated based on ^{230}Th age (T),

808   i.e., $\delta^{234}U_{initial} = \delta^{234}U_{measured} \times e^{\lambda234xT}$. Corrected ^{230}Th ages assume the initial $^{230}Th/^{232}Th$

809   atomic ratio of $4.4\pm2.2 \times 10^{-6}$. Those are the values for a material at secular

810   equilibrium, with the bulk earth $^{232}Th/^{238}U$ value of 3.8. The errors are arbitrarily

811   assumed to be 50%. ***B.P. stands for "Before Present" where the "Present" is defined

812   as the year 1950 A.D. Values are indicated at one decimal place more than significant,

813   to avoid rounding errors. Ages excluded from the final chronology are shown in

814   italics.

815

816   **Figures**

817

818   Figure 1: A) Map with summer climatological conditions in the broader study area.

819   The location of Mawmluh Cave in NE India is indicated by the black dot. Other

820   discussed cave locations are indicated by the grey dots and arrows (D: Dongge Cave,

821   Y: Yamen Cave, K: Kulishu Cave). The dashed line indicates maximum northward

822   extent of the Intertropical Convergence Zone (ITCZ), which drives monsoonal

823   circulation. Brown arrows delineate dominant ISM circulation patterns, Asian

824   Summer Monsoon (ASM) winds are shown in green. B) Map of Mawmluh Cave.

825   Stalagmite MAW-6 was found broken in the West Stream (map courtesy of Daniel

826   Gebauer). C) Scan of cut and polished stalagmite MAW-6.

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829   Figure 2: Age-model of MAW-6 (constructed using cubic interpolation in COPRA).

830   The median of the age model is shown in blue, with the 95% confidence intervals in

831   light grey. The U-series ages used to construct the age model are shown in black,

832   while the excluded ages are in red. Hiatuses are indicated by dashed black lines.

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834

835   Figure 3: A) $\delta^{18}O$ and $\delta^{13}C$ records with median and 95% confidence intervals. The

836   time periods discussed in this study are indicated at the bottom of the figure (IM:

837   Intermediate period, 10.8-10.2 kyr BP, YD: Younger Dryas, B-A: Bølling-Allerød,

838   LG: late glacial). Major controls on $\delta^{18}O$ (ISM strength) and $\delta^{13}C$ (amount of in-cave

839   fractionation due to cave air pCO_2 and drip rate) are indicated by the bars.

840   B) Cross-correlation between $\delta^{18}O$ and $\delta^{18}O_{IVC}$ vs. $\delta^{13}C$, estimated using kernel-based

841   cross correlation analysis (Rehfeld and Kurths, 2014) with the toolbox NESTool

842   (http://tocsy.pik-potsdam.de/nest.php).

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845   Figure 4: A) $\delta^{13}C$ vs. ice volume corrected $\delta^{18}O$ ($\delta^{18}O_{IVC}$) relationship in stalagmite

846   MAW-6. The record can be subdivided into clusters corresponding to different time

847   periods: Holocene (Hol), Intermediate (IM), YD, and B-A and late glacial (B-A/LG).

848   All clusters show high degrees of correlation between $\delta^{13}C$ and $\delta^{18}O_{IVC}$, indicated by

849   the corresponding r values (same values as in Fig. 3B). Arrows indicate the direction

850   of the main forcings (dry season aridity and ISM strength). B) Boxplots for $\delta^{18}O_{IVC}$

851   and $\delta^{13}C$ (top and bottom panel respectively). Boxes are defined by the median (red

852   line) and delimited by the 1st and 3rd quartile. Whiskers define the lowest and highest

853   value within 1.5 times the inter quartile range of the cluster. Outliers are indicated by

854   red crosses.

855

856

857   Figure 5: Schematic of the factors influencing isotope signals at Mawmluh Cave. Data

858   is derived from monitoring studies at the cave site (Breitenbach et al., 2015), and

859   precipitation data is from the Indian Meteorological Department Station Cherrapunji.

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861

862   Figure 6: Comparison of stalagmite $\delta^{18}O$ records MAW-6 (blue), MWS-1 (orange,

863   Dutt et al., 2015) and KM-A (purple, Berkelhammer et al., 2012). Proxy uncertainties

864   (95% confidence intervals), as calculated by COPRA, are shown in light shading.

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866

867 Figure 7: Comparison of MAW-6 $\delta^{18}O_{IVC}$ to $\delta^{18}O$ records from the ISM and broader

868 AM regions, as well as to the NGRIP Greenland ice core record. MAW-6 reflects and

869 corroborates previous reconstructions from the Asian monsoon region showing the

870 weakest ISM after the deglaciation occurring during the Younger Dryas, and stronger

871 ISM during the preceding Bølling-Allerød, as well as during the Holocene.

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874 Figure 8: Results of the TACTS analysis on MAW-6 $\delta^{18}O$. The cumulative

875 probability distribution established through 5000 random realizations of DET measure

876 is shown by the blue line, with gray shading indicating the 95% confidence interval.

877 The late glacial (LG) and early Holocene (EH) are characterized by distinct DET

878 measures (0.663 and 0.736, respectively). While the LG is within the 95% confidence

879 interval, the EH is outside, highlighting the high predictability of the ISM during this

880 period.

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