Paleoclimate change in Ethiopia around the last interglacial derived from annually-resolved stalagmite evidence

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2 **annually-resolved stalagmite evidence**

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21

22 **ABSTRACT**

23 Oxygen and carbon ($\delta^{18}O/\delta^{13}C$) isotope, growth rate and trace element data are reported for a

24 U-Th dated, annually-laminated stalagmite, GM1 from Goda Mea Cave, Ethiopia. The

25 stalagmite grew intermittently around the last interglacial. The proxy records are used to

develop a conceptual growth model of the stalagmite and to assess its potential for revealing a climate signal in this climatically sensitive northeastern African region during an important period in the evolution of *Homo sapiens* and dispersal of Anatomically Modern Humans out of Africa. Speleothem deposition is of short-duration occurring at ~129 ka, ~120 ka, in an undated growth phase, and at ~108 ka; probably due to tectonic activity. $\delta^{18}O$ composition is very stable within growth phases (1σ variability < 0.76‰), as are Mg/Ca, Sr/Ca and Ba/Ca, all indicative of well-mixed source-waters. A shift to positive $\delta^{18}O$ values and increased variability in Mg/Ca, Sr/Ca and Ba/Ca prior to growth hiatuses is observed, indicating a loss of the well-mixed water source prior to growth cessation. Mean $\delta^{18}O$ composition (–3.82 to –7.77‰) is lower than published modern and Holocene stalagmites from the region. Geochemical data, statistical analyses, and a conceptual model of stalagmite growth, demonstrate that climatic conditions recorded by GM1 were wetter than the Holocene. The ~129 ka growth phase particularly presents an annual record of the relative Intertropical Convergence Zone (ITCZ) position. The GM1 record, the oldest high-resolution continental climate record from Ethiopia so far published, presents evidence that any early human migrations which occurred during MIS 5 are likely to have occurred during a wet event in northeast Africa.

1. Introduction

In Ethiopia, stalagmites provide high-resolution records of past climate and environment (Asrat et al., 2007; Baker et al., 2007; 2010). Fast-growing, annually-laminated stalagmites are ubiquitous, due to the strong seasonality of rainfall and the water balance in Ethiopia. Regular laminae, visible in hand section, can provide precise annual chronology. Annual

growth rates of these stalagmites, determined from the thickness of an annual lamina, is at the upper range of those observed in stalagmites (typically ~0.5 mm/yr). This is due to the optimal climatic conditions (high temperature and rainfall) for limestone dissolution and re-precipitation. This rapid growth facilitates the high-resolution sampling of stalagmite calcite.

In Ethiopia, the real advantage of using speleothems to provide a paleoclimate proxy record is that they contain information on past rainfall variability in the region. Several major air streams and convergence zones influence the current climate pattern in northeast Africa, whose effects are often compounded by such regional factors as topography and the proximity to the oceans (e.g., Nicholson, 1996). The relatively dry north-easterly and south-easterly monsoons and the humid and moisture-laden (rainfall generating), westerly and south-westerly air flow of the Congo air stream, generally dominate the regional wind and pressure patterns. The Intertropical Convergence Zone (ITCZ) and the Congo Air Boundary (CAB) separate these major air streams. The passage of the ITCZ (Fig. 1a) dominantly determines the rainy seasons in Ethiopia, while the topography (highland barriers separated by a rift zone) modulates the local rainfall distribution. Accordingly, Ethiopian climate has two rainy seasons, one from the northward passage of the ITCZ, called locally the 'big rains' (between June and September), which is reliable and whose maxima migrates with the position of the ITCZ. A second rainy season, the 'small rains', is less consistent and occurs between March and May with maxima in April. Dryland farming, including subsistence farming, leads to a high dependency on rains in both seasons. Failure of the 'small' rains is common and has occurred in recent years in 2013/2014 and 2015/2016, particularly in the southeastern Ethiopian lowlands bordering the current study area. The climate dynamical cause of the failure of the 'small' rains, and how this varies over time, is still poorly understood.

In addition, reliable, high-resolution climate records beyond the Holocene are scarce in the northeastern African region, one of the major candidates for the origin of *Homo sapiens* and a gateway to the "out of Africa" migration of our species during the late Pleistocene. The influence of climate on the dispersal of Anatomically Modern Humans from northeastern Africa particularly during the period ~120 to ~50 ka has been a subject of intense discussion (e.g., Tierney et al., 2017 and references therein; Lamb et al., 2018). Recent discovery of *Homo sapiens* fossils dated to 177 to 194 ka in the Misliya cave in Israel (Hershkovitz et al., 2018) indicates that the "out of Africa" migration episodes have started earlier than the previously thought period of migration (~120-50 ka). Discussions on influence of climate on human dispersal often rely on marine climate records from the Indian Ocean and Mediterranean Sea. The recently published Lake Tana record from the northwestern Ethiopian highland, largely covering the last ~150 ka (Lamb et al., 2018) is the only continental record available. In this paper, we present a high-resolution continental climate record from an Ethiopian stalagmite (GM1) that grew intermittently around the last interglacial, which is very pertinent to this discussion. Though it is not a continuous record over the whole period of the last interglacial, the growth phases of GM1 are dated at particularly important periods of the MIS 5. The GM1 record, the oldest high-resolution climate record so far published from Ethiopia and continental eastern Africa, is therefore very significant in an area where any kind of reliable continental climate records from this period are scarce.

These annual-resolution records of $\delta^{18}O$, $\delta^{13}C$, trace elements and growth rate are from the Goda Mea Cave in Ethiopia (Fig. 1b). A combination of U-Th dates and lamina counting are used to identify the timing of the growth phases. Samples milled at annual resolution were analysed for $\delta^{18}O$ and $\delta^{13}C$, and at decadal resolution for trace elements. Variogram, autocorrelation and spectral analyses of the geochemical and growth rate time series are used

4

100   to develop a conceptual model for the hydrology of the waters feeding the speleothem. The

101   time series of $\delta^{18}O$, $\delta^{13}C$, trace elements and annual growth rate are then interpreted, with

102   comparison to the published modern and Holocene stalagmites from the region and globally.

103   Such high resolution, multi-proxy approach has been proved useful in reconstructing annual,

104   in some cases seasonal, rainfall (e.g., Johnson et al., 2006).

105

106   **2. Methodology**

107   *2.1. Site Description*

108   Goda Mea Cave was explored and surveyed in 2007 and a full description can be found in

109   Gunn et al. (2009). The cave is entered from a collapse doline and after about 30 m there is a

110   large flowstone deposit that almost fills the passage. A crawl beneath opens into a NE-SW

111   oriented rift passage that is initially some 5 m wide by 1 m high but increases downstream to

112   10-15 m by 7 m. The cave ends in a 90 m x 40 m x 20 m high chamber formed by upwards

113   stoping as evidenced by abundant breakdown. Above the chamber there is ~25 m of sandy

114   limestone intercalated with some thin marl and mudstone layers towards the top, overlain by

115   a ~20 m thick calcareous sandstone, silt, carbonate rich shale and marl intercalation, which

116   extends to ~1 m of soil at the surface. The limestone and sandstone-shale-marl units above

117   the cave form a continuous hydrogeological unit, connected by network of fractures.

118   There are numerous speleothems in the chamber including stalactites, stalagmites and a

119   central column that is over 10 m high and 6 m diameter. The speleothems are mostly relict

120   with some evidence of re-solution and many of the stalagmites are fractured, most likely by

121   tectonic activity (Fig. 1c and d). Speleothem growth is generally focused along some aligned

122   zones below major fracture systems/brecciated fault traces crossing the hydrogeological unit

123   all the way up to the surface.

124   Some modern monitoring data (e.g., drip water chemistry) for the cave is presented in

125   Asrat et al. (2008). Drip water Ca^{2+} and Mg^{2+} concentrations in all analysed drip water

126   samples from this cave are 2.57±0.65 mmol/L and 1.54±1.12 mmol/L, respectively. The drip

127   water Ca^{2+} concentration in this cave is high as compared to the range of Ca^{2+} in the Mechara

128   caves (2.63±2.36 mmol/L) and falls within the range of values expected for "open system"

129   evolution (Baker et al., 2016). The high Ca^{2+} concentration can be attributed to the open

130   system calcareous sandstone/shale, marl and limestone hydrogeological unit, with the calcite-

131   cemented sandstones, carbonate rich shale, marl and limestone all contributing Ca^{2+} ions to

132   the drip waters.

133

134   *2.2. Sample description*

135   GM1 is a large broken stalagmite found in the cave chamber. The 591 mm long

136   stalagmite was sectioned into two halves, and one half polished for lamina counting (Fig. 2).

137   The polished half shows continuous laminations, alternating between dense and porous/white

138   calcite, as well as visually recognizable growth hiatuses, marked by shifts in growth axis and

139   the stalagmite morphology. The other half of GM1 was milled using a hand held dental driller

140   for oxygen and carbon isotopes at ~0.6 mm resolution (966 samples), and trace element

141   analysis at ~5.5 mm resolution (103 samples). 38 samples for U-Th dating were similarly

142   drilled using a dental driller, with samples located either side of possible growth hiatuses, and

143   regularly spaced within growth phases.

144

145   *2.3. Geochemical analyses*

146   U-Th analyses were undertaken by ICP-MS at the University of Melbourne, Australia,

147   following the method of Hellstrom (2003). Samples were dissolved in concentrated HNO_3

148   and equilibrated with a mixed $^{229}Th–^{233}U–^{236}U$ tracer. U and Th were extracted in a single

149 solution using Eichrom TRU resin before introduction to a Nu Plasma multi-collector ICP-

150 MS, where isotope ratios of both elements were measured simultaneously. The decay

151 constants of Cheng et al. (2013) were used, and detritally-corrected ages calculated using eqn.

152 1 of Hellstrom (2006) with an assumed initial [^{230}Th/^{232}Th] of 1.5 ± 1.5. Age-depth modelling

153 combined floating annual laminae chronologies and U-Th analyses were as described in

154 section 3.1.

155   Oxygen and carbon isotopes were analysed at the Stable Isotope Laboratory (SILLA),

156 University of Birmingham, UK. The calcite samples were reacted with phosphoric acid and

157 analysed using an Isoprime continuous flow mass spectrometer. By comparison with a

158 laboratory marble standard, the sample ^{18}O/^{16}O and ^{13}C/^{12}C ratios are reported as δ^{18}O and

159 δ^{13}C values in per mil (‰) versus VPDB. Analytical precisions are 0.07‰ for δ^{18}O and

160 0.04‰ for δ^{13}C on the standard marble (KCM).

161   Trace elements powders were analysed at University of New South Wales, Sydney.

162 Samples of approximately 0.05 g each were weighed directly into polypropylene vials. One

163 mL of 1-1 hydrochloric acid was added to each vial. The samples were sonicated for 15

164 minutes to ensure complete dissolution. The solutions were diluted to 10.0 mL with ASTM®

165 Type I water (Millipore® filtration system, Millipore® Corporation, Billerica, Massachusetts,

166 USA).

167   Diluted samples were analysed for Ca (317.933 nm) and Mg (285.213 nm) using the

168 PerkinElmer Optima™ 7300DV ICP-OES (PerkinElmer, Shelton, USA). Ba and Sr were

169 analysed by PerkinElmer NexION 300D ICP-MS (PerkinElmer, Shelton, USA). Both

170 instruments were coupled with an ESI SC4 FAST sample introduction system (Elemental

171 Scientific, Inc., Omaha, USA) to minimise sample carryover.

172   The ICP-OES and ICP-MS were calibrated using certified multi-element standards in a

173 matrix of 2% HCl. Wavelength and analytical mass selection took into consideration spectral

174   interferences as well as sensitivities. Internal standards were added on-line via injection valve

175   to correct for physical interferences. Quality control check standards were run at selected

176   intervals in an unattended automatic analysis run, to ensure that the instrument performance

177   remained consistent over the length of analysis.

178

179   *2.4. Time series analysis*

180   Statistical analysis on the annual growth rate time series followed the approach of

181   Mariethoz et al. (2012), which included the analysis of the first derivative of growth rate

182   (growth acceleration) to determine the flickering parameter (f), which is the magnitude of the

183   anti-correlation at lag 1. Flickering ranges between –0.5 and 0, the more negative *f* values

184   indicating stronger flickering. Negative values of *f* are indicative of a karst store filling and

185   draining, as opposed to a climate forcing, and helps identify climatically sensitive

186   speleothems. In addition, variogram analysis of the growth rate time series permits the

187   derivation of the information content (*IC*) and range (*r*) in the growth rate data, which helps

188   identify the signal: noise ratio in the data and the time over which useful information might

189   be expected. Stable isotope and annual growth rate time series data were also analysed for

190   their autocorrelation and spectral properties. As the data was evenly spaced in time, spectral

191   analysis was performed using discrete Fourier transforms, using the FFTW library within

192   Microcal Origin. Five windows were used (Bartlett, Hanning, Rectangular, Welsh and

193   Triangular) in order to investigate the extent of signal leakage.

194

195   **3. Results**

196

197   *3.1. GM-1 Chronology*

198   *3.1.1. Lamina*

199   Lamina were counted (in duplicate) and a total of 1356 lamina were identified with a

200   mean lamina thickness of 0.44±0.14 mm (ranging between 0.19 and 1.12 mm). This lamina

201   thickness compares well to those reported for stalagmites in previous studies in the region

202   (Ach-1, mean = 0.53±0.26 mm, Bero-1 = 0.45±0.23 mm, Asfa-3 = 0.32±0.11 mm; Merc-1 =

203   0.29±0.04 mm; Asrat et al., 2007; Baker et al., 2007; Baker et al., 2010). In these stalagmites,

204   the visible laminae have been demonstrated to be annual by comparison to the radiometric

205   dates. The GM1 laminae are similar in their appearance and thickness to these other

206   stalagmites. Examination of thick sections of GM1 at various levels of the growth phases

207   (Fig. 2) show continuous and regular visible laminae with alternating brownish calcite (Dark

208   Compact Laminae, DCL) and thinner white calcite (White Porous Laminae, WPL) as defined

209   by Genty & Quinif (1996) and Genty et al. (1997). The presence of fine sediments on the

210   white porous calcite, and some dissolution features at the top of the DCL, suggests some

211   seasonal infiltration variability (*cf.* Borsato et al., 2007). Overall, the regularly alternating

212   DCL/WPL laminae sequence, even without more obvious structures from infiltration

213   variability, indicate deposition under a seasonal hydroclimate regime (e.g. changes in drip

214   water supersaturation or cave air CO_2 concentration), where recharge was sufficient to

215   maintain continuous dripping to the stalagmite.

216

217   *3.1.2. Growth hiatuses*

218   There are three major growth hiatuses based on the U-Th chronology (see below), and

219   other possible minor growth hiatuses have been recognized by changes in the growth axis

220   within the growth phases (Fig. 2). Visual examination of the three major growth hiatuses on

221   the polished stalagmite and the thick sections show that the hiatuses between the four major

222   growth phases are all marked by accumulation of fine detritus and brownish material on the

223 top 2 mm sections, with no indications of dissolution. Such textural features are typical of the

224 ceasing of growth due to cessation of the drip source (Railsback et al., 2013).

225

226   *3.1.3. Stalagmite morphology*

227   The morphology of the stalagmite changes from candle stick shaped, with regular nearly

228 horizontal lamina on and off the growth axis for most of the first growth phase to upwards-

229 thinning, laterally less extensive layers with laminae rapidly changing to sub-vertical angle,

230 off the growth axis, just below the first hiatus (Fig. 2). The second growth phase above the

231 first hiatus then gets broader at its axis with rapid flowing/dripping down the sides of the

232 stalagmite forming nearly vertical lamina. The third growth phase shows similar morphology

233 to the second though it rapidly thins towards the top below the third major hiatus. The last

234 growth phase has relatively broader shape with significant deposition along its axis. The

235 morphology of the stalagmite changing with the hiatus position is a clear demonstration of

236 the changing amount and concentration of calcite in the dripping water. It shows a general

237 drying out of the drip source towards the tops of the three older growth phases, while the last

238 growth phase is marked by an increased drip rate throughout the growth period.

239

240   *3.1.4. U-Th dates and annual growth rate*

241   The 38 U-Th dates on the sample (Fig. 3, Table 1) demonstrate 4 periods of growth and

242 confirm the presence of 3 hiatuses. Several age inversions are present, and one short growth

243 phase containing 37 laminae was undated.

244   The U-Th ages were used to constrain a chronology based on the annual laminae. Firstly,

245 the longest phase of speleothem growth (from ~127 mm from the top, to the base at 591 mm)

246 contained 29 very similar U-Th ages (a mean and standard deviation of 130.0 ± 3.5 ka, with

247 an average uncertainty on individual analyses of 1.3 ka), providing further evidence that the

248   1087 laminae present in this growth phase are likely to be annual. Secondly, following

249   established methods (Asrat et al., 2007; Dominguez-Villars et al., 2012; Baker et al., 2015),

250   we compared two approaches to tie the lamina chronology to the U-Th ages. The first

251   approach was as follows: within each growth phase, each U-Th age was adjusted by using its

252   relative laminae age to obtain an equivalent U-Th age for the date of the start of each growth

253   phase. Taking the mean and standard deviation, this yielded growth phases starting at

254   129.3±2.7 ka, 120.7±1.7 ka, and 108.3±0.2 ka. We compared this approach to that calculated

255   from linear regression applied to conventional age-depth plots. In this case, we used only the

256   U-Th ages with a $[^{230}Th/^{232}Th]>1000$, presuming they would be the most accurate. This

257   approach yielded a date for the start of deposition for two of the four growth phases of

258   129.2±1.7 ka and 120.6±0.3 ka. The two approaches therefore give consistent dates for the

259   start of deposition that agree with the analytical error of individual analyses.

260   GM1 deposition periods are therefore ascribed to four phases: 1087 years commencing

261   129.3±2.7 ka, 54 years of deposition at 120.7±1.7 ka, a 37-yr long undated growth phase, and

262   176 years of deposition from 108.4±0.3 ka. The ~129.3 ka growth phase occurs, within

263   dating uncertainty, at Termination II or the early part of the last interglacial (Cheng et al.,

264   2009), and the 120.7 ka deposition immediately post-dates the full interglacial. The growth

265   phase at 108.5-108.3 ka falls within the isotope stage 5c interglacial.

266   The annual growth rate for GM1, determined from the annual lamina thickness, is

267   presented in Figure 4. Mean growth rate does not vary between growth phases: 0.43±0.14

268   mm/yr (129.3 ka), 0.41±0.17 mm/yr (120.7 ka), 0.53±0.15 (undated), and 0.47±0.14 mm/yr

269   (108.4 ka).

270

271   *3.2. Oxygen and carbon isotopes*

272   The oxygen and carbon isotope (δ^{18}O and δ^{13}C) data were ascribed to an annual lamina

273   and are presented in Figures 4. Scatter plots of δ^{18}O and δ^{13}C, including the analysis of the

274   isotopes along a lamina (the 'Hendy test') are shown in Figure S1.

275   δ^{18}O varies significantly between growth phases (–7.77±0.57‰ (129.3 ka), –

276   3.82±0.61‰ (120.7 ka), –6.05±0.76‰ (undated), and –6.31±0.59‰ (108.4 ka). Within each

277   growth phase, δ^{18}O can be described as having long periods of relatively invariant

278   composition (e.g., ±0.33‰ for the first 1000 years of the 129.3 ka deposition period, and

279   ±0.23‰ in the first 150 years of the 108.4 ka deposition period), as well as periods of rapid

280   change. For example, δ^{18}O increased from –7.8‰ to –5.3‰ in four years (and to –4.5‰ after

281   13 years) at the end of the 129.3 ka growth phase, and from –6.2‰ to –4.4‰ in nine years at

282   the end of the 108.4 ka growth period.

283   δ^{13}C is characterised by low inter-sample variability, with the presence of long-term

284   trends. For example, in the 129.3 ka growth phase, the standard deviation of δ^{13}C over any

285   50-year period is between 0.1‰ and 0.3‰, but over the whole 1087 years of deposition, δ^{13}C

286   trends from –1‰ to –4‰. This 3‰ change in δ^{13}C with the growth phase is as great as the

287   variability between growth phases.

288   Figure S1 shows the relationship between δ^{18}O and δ^{13}C, both throughout the time series

289   as well as along growth layers ('Hendy tests'). The Hendy tests suggest a 1‰ increase or

290   decrease in isotope composition is possible along a growth layer, which is greater than the

291   inter-annual variability of δ^{13}C and δ^{18}O. For stalagmite GM1, there is no evidence for near-

292   equilibrium deposition: modern and Holocene stalagmites demonstrate isotope fractionation

293   of 1-2‰ (Asrat et al., 2007; Baker et al., 2007; 2010), and similar deposition conditions

294   appear to apply to stalagmite GM1. Based on these works, we have quantified these

295   fractionation processes and confirmed that they operate in the same direction as the climate

296   forcing, which has also been observed by other works (e.g., Dorale and Liu, 2009).

297

298   *3.3. Trace elements*

299   Ba/Ca, Mg/Ca and Sr/Ca show similar patterns to $\delta^{13}C$ and $\delta^{18}O$, although sampled at a

300   lower temporal resolution. A long-term trend to lower ratios in the 129.3 ka growth phase

301   matches that observed in $\delta^{13}C$. Significant short-term changes in trace element composition

302   occurs at the same time as the increases in $\delta^{18}O$ and $\delta^{13}C$ at the end of the 129.3 ka growth

303   phase, and within the 120.7 ka growth period. At the end of the 129.3 ka growth phase, a

304   change in gradient of Mg/Ca, Sr/Ca and Ba/Ca lasted for ~170 years, indicative of a drying

305   trend. This was followed by the $\delta^{18}O$ increase of 2.2‰ that occurred over 4 years, and then

306   an increased variability in Sr/Ca, Mg/Ca and Ba/Ca (Mg/Ca increases, Mg/Ca and Ba/Ca

307   decrease) until growth stops 28 years later. In contrast, the trace element response at the time

308   of a 2‰ increase in $\delta^{18}O$ at the end of the 108.4 ka growth phase is muted and trends to lower

309   values. The greatest range in trace elements occurs in the 120.7 ka growth phase, where Sr/Ca

310   increases, and Mg/Ca and Ba/Ca have opposing increasing and decreasing trends.

311

312   *3.4. Time series analysis on stable isotope and growth rate time-series*

313   Following Mariethoz et al. (2012), and as described in section 2.4, the growth rate time

314   series variogram properties was investigated. Due to the short duration of several of the

315   growth phases, only the longest time series at 129.3 ka was analysed. The results are plotted

316   in Figure 5a, where variogram analysis on stalagmite GM1 is compared to previous published

317   stalagmite statistics on growth rate series. The autocorrelation and spectral properties of the

318   growth rate, $\delta^{18}O$ and $\delta^{13}C$ series were also investigated (Figs 5b and c).

319   Stalagmite GM1 growth at 129.3 ka has evidence of 'flickering' (f=−0.33), that is a

320   growth acceleration that flickers around a mean value (Mariethoz et al., 2012). 'Flickering'

321   has been explained as growth rate sensitivity to the filling and draining of a karst store, which

322 is trying to reach a dynamic equilibrium, with reported values between –0.24 (low flickering,

323 potential climate signal) and –0.39 (high flickering, potential karst hydrology signal).

324 Variogram analysis shows that the stalagmite has low range (r, 20.5 years) in the growth

325 rate record. Growth rate therefore has no 'memory' of previous growth rates longer than this

326 timescale, indicative that the karst store(s) that feed the stalagmite are relatively small. The

327 GM1 range is the lowest reported to date for an annually laminated speleothem. The

328 Information Content (IC) of growth rate, which is the balance of the signal in the variogram

329 and the noise, is 56%, indicating that the stalagmite growth rate record contains a greater

330 proportion of signal than noise. In comparison to previously published records (Fig. 5a),

331 stalagmite growth rate statistical properties lie in region A, with stalagmites that have a high

332 information content, relatively low flickering and range, and where growth rate has proven

333 useful in paleoclimate reconstruction.

334 Autocorrelation of growth rate, $\delta^{13}C$ and $\delta^{18}O$ time series are presented in Figure 5b for

335 the three longest growth phases (~129, ~120 and ~108 ka). Significant autocorrelation can be

336 observed for $\delta^{13}C$ and $\delta^{18}O$ for the ~129 and ~108 ka growth phases, with autocorrelation

337 >0.6 at 15-year lag. Autocorrelation for $\delta^{13}C$ is stronger than for $\delta^{18}O$, indicative of

338 additional smoothing of the $\delta^{13}C$, likely from the soil carbon store. Growth rate has very low

339 autocorrelation (<0.4 after 4 years lag), in agreement with the observed flickering of growth

340 rate. The ~120 ka growth phase has lower autocorrelation of $\delta^{18}O$ and $\delta^{13}C$ than the other two

341 growth phases, suggesting limited mixing or smoothing of these proxies during this short

342 growth phase. Low autocorrelation would agree with the observed highest variability in trace

343 elements at this time.

344 Spectral analysis was also undertaken (Fig. 5c) for the longest continuous growth phase at

345 ~129 ka. Bartlett, Hanning, Rectangular, Welsh and Triangular windows were used to

346 explore the spectral properties for $\delta^{18}O$ and growth rate time series. Growth rate has a 17-18

347 year peak that is not statistically significant at 95% confidence, and two other peaks (31-33

348 and 53-59 years), which are longer than the range (r, 20.5 years) and are likely to be

349 harmonics of the 17-18 year frequency. $\delta^{18}O$ also has only weak and insignificant spectral

350 power, not surprising given the low variability in the $\delta^{18}O$ data.

351

352 **4. Discussion**

353

354   *4.1. Holocene stalagmite records*

355   Previous cave research in Ethiopia has included limited cave drip water and climate

356 monitoring during sampling expeditions to the Mechara region of Ethiopia between 2004 and

357 2008 (Asrat et al., 2008) (Fig. 1), and the analysis of modern and Holocene stalagmite

358 samples. Modern calibration studies of stalagmite $\delta^{18}O$ of carbonate shows evidence of

359 climate sensitivity, despite deposition out-of-equilibrium (Baker et al. 2007; 2010). The latter

360 is potentially due to both rapid degassing and evaporation. $\delta^{18}O$ and growth rate correlations

361 with climate are sample-specific. Drip-specific flow-paths determine whether a stalagmite

362 has a proxy which is sensitive to the 'big' rains, or to the relative amount of rain in the 'big'

363 and 'small' rain seasons, or neither. For example, modern stalagmite $\delta^{18}O$ and growth rate

364 records were reported from two stalagmites from Rukiesa Cave (Baker et al., 2007). The

365 annual nature of the laminae was confirmed by ^{14}C analyses and comparison to the modern

366 atmospheric bomb carbon peak. In these samples $\delta^{18}O$ and growth rate were shown to have a

367 correlation with the ratio of 'small' to 'big' rainfall and total summer rainfall, respectively. A

368 sample specific climate sensitivity of $\delta^{18}O$ and growth rate was observed, which probably

369 reflects the karst hydrogeology and its effect on individual water flow-paths. A loss of

370 climate correlation was also observed in one sample during a period of high growth rates.

371    $\delta^{18}O$ and growth rate time-series both exhibit multi-decadal variability in two stalagmite

372   samples deposited in the Holocene. Such variability may be an amplification of extremes in

373   hydroclimate (e.g., drought years) or rainfall isotopic composition, due to the non-linear

374   replenishment or drainage of karst stores (Baker et al., 2013). A mid-Holocene record from

375   Achere Cave (sample Ach-1) had laminae that were demonstrated to be annual by

376   comparison to U-Th dates (Asrat et al., 2007). In this stalagmite a 18-21 yr periodicity in

377   growth rate and $\delta^{18}O$ occur. $\delta^{18}O$ has a greater variability than $\delta^{13}C$, indicative of the

378   variability being driven by variations in the extent of evaporative enrichment of $\delta^{18}O$. A

379   discontinuous Holocene record from Bero Cave had six growth phases over the last 7800

380   years (Baker et al., 2010). Mean stalagmite $\delta^{18}O$ is 1.2‰ higher than that predicted by

381   forward modelling, and a multi-decadal variability in $\delta^{18}O$ and growth rate was again

382   observed (Baker et al., 2010). $\delta^{18}O$ from this stalagmite was indicative of both rapid

383   degassing and the additional enrichment, probably due to evaporation.

384   Stalagmite growth phases are relatively short (10^3-10^4 years) due to the tectonically active

385   nature of the region, which can change water flow paths. This is observed in stalagmites from

386   all three caves, and has been explained by changes in flow regime or to the relative position

387   of a growing stalagmite caused by tectonic activity related to the East African Rift (Asrat,

388   2012). Physically anomalous laminae within an otherwise regular and visible annual laminae

389   sequence, frequent deviations from vertical growth axis, and abrupt changes in stalagmite

390   morphology, as well as the tectonically-controlled formation of the larger cave system,

391   further confirm the influence of tectonics and recorded earthquakes in the region (see Fig. 1a)

392   on the length of the growth phases (Asrat et al., 2008; Asrat, 2012).

393

394   *4.2. Conceptual growth model*

395   The combination of stable isotope and trace element geochemistry, growth rate, statistical

396   analyses, and observations of the laminae types and stalagmite shape, allow us to build a

397   conceptual model for stalagmite GM1 (Fig. 6).

398   Firstly, the stalagmite probably had continuous deposition for more than 1000 years, and

399   during this period isotope and trace element composition has low variability and high

400   autocorrelation. This homogeneity of $\delta^{13}C$, $\delta^{18}O$ and trace elements suggests a drip water

401   source which is well mixed, enough to obscure any annual to decadal scale variability in $\delta^{18}O$

402   and maintain dripping. The continuous deposition of the laminae and the candle-stick shape

403   of the stalagmite before it narrows down towards the tip of this growth phase (the last few

404   years of growth) supports a continuous drip source. We propose this water comes from

405   matrix flow of the porous sandstone and sandy limestone, which was channelled to the drip

406   source by a network of small fractures. The annual laminae are driven by this flow regime,

407   which provides the necessary seasonal variability in drip water hydro-geochemistry.

408   Combined with the evidence from flickering, we infer the variations in lamina thickness are

409   driven by the karst hydrology and not by the cave environment. In the ~129 ka growth phase,

410   these would have to maintain high levels of saturation for the initial ~1000 years of

411   deposition.

412   Secondly, the four growth phases of GM1 reflect the changing karst hydrologic regime

413   above the cave. The ~129 ka growth phase is marked by the dominance of a continuous

414   supply of water from 'matrix' flow for most of its growth period, which rapidly dried out a

415   few tens of years before the hiatus. In contrast, the ~120 ka growth phase reflects a rapid

416   'fracture' flow, following a possible tectonic event, which did not maintain growth for a long

417   time before it abruptly shut off as marked by the rapid increase in trace element ratios and

418     $\delta^{18}O$. The third, undated growth phase shows similar features to that of the second, only the

419   growth period was shorter suggesting a rapid start to dripping and subsequent exhaustion

17

420　from a 'fracture' source, as marked by the nearly vertical laminae down the sides of the

421　stalagmite and the abrupt narrowing towards its top. The ~108 ka growth phase is again

422　marked by a more 'porous' matrix flow which maintained growth for longer period, attested

423　by the regular laminae, broad stalagmite shape, generally low variability trace element ratios

424　and depleted $\delta^{18}O$.

425　　　Thirdly, growth rate varies annually, shows evidence of flickering, and has a range of ~20

426　years and a spectral peak at 17-18 years.　Given the relative homogeneity of the stable isotope

427　and trace element signals, the growth rate variability has to occur subsequent to the mixing of

428　the water. Dissolutionally-enlarged fractures or a network of small conduits would allow

429　limited water storage, permitting degassing of karst water and prior calcite precipitation

430　(PCP) as well as drip rate variations, both affecting growth rate variability. We conceive this

431　store to have a proportional volume of approximately 20 years of recharge (see later). In

432　drying conditions, water from this store could maintain dripping and deposition for short time

433　periods.

434　　　Fourth, a pre-existing fracture zone/brecciated fault trace which might have been

435　reactivated during subsequent tectonic activities, extending from the surface, through to the

436　store and the cave roof, permits fracture flow to stalagmite GM1. This would permit short-

437　duration recharge, probably after high magnitude / frequency rainfall events, in the absence

438　of saturated porous sandstone and limestone aquifer. This would explain the short duration,

439　high geochemical variability, ~120 ka growth phase. The absence of 'stored' water and

440　subsequently rapid exhaustion of the fracture flow/drip source is supported by the nearly

441　vertical lamina depositing down the sides of the stalagmite, high variability/rapid increase in

442　the $\delta^{18}O$ and trace element ratios and low autocorrelation in $\delta^{13}C$ and $\delta^{18}O$.

443　　　The varying trends between the $\delta^{18}O$ and $\delta^{13}C$, and trace element ratios from one growth

444　phase to the other suggests that a single kinetic fractionation process does not dominate our

445   proxy records, rather we infer a hydrological control based on climatic and tectonic

446   processes.

447   Our conceptual model explains other features of GM1 geochemistry and growth rate. The

448   geochemistry at the end of the 129.3 ka growth phase can be interpreted as a decline in the

449   saturation or water level in the porous media, leading to an increase in $\delta^{18}O$ and trace

450   elements as dripping is maintained just from the smaller store. Before growth stops, a change

451   in gradient of Mg/Ca, Sr/Ca and Ba/Ca indicates a drying trend which lasts for ~170 years,

452   followed by a $\delta^{18}O$ increase of 2.2‰ that occurred over 4 years, and then increased

453   variability in Sr/Ca, Mg/Ca and Ba/Ca until deposition stops 28 years later. Many of the

454   previously studied stalagmites from Ethiopia such as Ach-1 (Asrat et al., 2007) show similar

455   features, which could be attributed to the specific geological setting of the region where

456   earthquake/tectonics play a strong role in shifting the relative position or the extent of the

457   major 'fracture' flow routes for such short-phased growths, leading to growth maintained for

458   short time longer from the smaller 'matrix' flow.

459   The similarity in values for the range r (20.5 yrs), the spectral frequency f (17-18 yrs),

460   and the observation that it takes 24 yrs for the stalagmite to stop growing, all suggest the

461   presence of a water store that can hold ~20 years of recharged water. The multi-decadal

462   growth rate frequency of 17-18 yrs, although insignificant, is in agreement with that observed

463   from Holocene stalagmites in the region (Ach-1, Bero-1; Baker et al., 2010), and similar to

464   observed variability in the modern rainfall pattern and subsequent flow at the upper Blue Nile

465   (Taye and Willems, 2012). Plausible climatic forcing over this timescale includes changes in

466   Atlantic and Indian Ocean sea surface temperature and variability in the movement and

467   intensity of the ITCZ and its effect on Ethiopian rainfall (Degefu et al., 2017). However, the

468   similarity of f and r suggests that any climate forcing in GM1 growth rate variability may be

469 amplified by the size of karst water store, or be karstic rather than climatic, the latter

470 something previously observed in forward modelling studies (Baker et al., 2013).

471   Finally, the shift to higher isotope values within a growth phase can be explained by our

472 conceptual model as a change from porous flow being the dominant water source to a

473 dominance of fracture flow. It suggests that the observed 2‰ shift could be indicative of a

474 water that has undergone additional kinetic or evaporative isotope fractionation. Both

475 fractionation processes had been previously inferred as occurring in both Modern and

476 Holocene Ethiopian speleothems, and from 'Hendy tests' on GM1, to a similar extent (up to

477 1‰). The implication for the climatic interpretation of stalagmite $\delta^{18}O$ is that variability of

478 up to 2‰ cannot be directly ascribed to climatic forcing, but larger changes cannot be

479 explained by fractionation processes. Similar rapid shifts in $\delta^{18}O$ of +2‰ within a period of 6

480 years have been identified in the Hulu cave speleothems (Treble et al., 2007).

481

482   *4.2. The climate record*

483   We can compare the $\delta^{18}O$ composition for each GM1 growth phase with published

484 Holocene stalagmite data from Ethiopia (Asrat et al., 2007; Baker et al., 2007; 2010), as well

485 as the modelled solar insolation for 15°N (Laskar et al., 2004), and other archived speleothem

486 $\delta^{18}O$ records along the monsoon path and the "downstream" countries (China, Cheng et al.,

487 2017; Israel, Bar Matthews et al., 1999; 2003). This comparison is shown in Figure 7. In GM-

488 1 the $\delta^{18}O$ composition (–7.77±0.57‰ (129.3 ka), –3.82±0.61‰ (120.7 ka), –6.05±0.76‰

489 (undated), and –6.31±0.59‰ (108.4 ka)) is generally isotopically more negative compared to

490 both modern (Merc-1: –1.22±0.31‰; Asfa-3: –1.37±0.37‰; Baker et al. 2007), and

491 Holocene (Bero-1: –3.42±1.45‰, Baker et al., 2010; Ach-1: –3.20±0.35‰, Asrat et al.,

492 2007) samples from the region. Even allowing for kinetic fractionation and non-equilibrium

493 deposition of up to 2‰ in all samples, GM1 $\delta^{18}O$ composition at 129.3 ka, 108.4 ka, and an

494 undated growth phase, is more negative than any Holocene stalagmites from the region.

495 Combined with our conceptual understanding of GM1 deposition, we can be certain that

496 these growth phases and lower isotope composition are indicative of wetter conditions and

497 sustained recharge.

498   The GM1 record is the first high-resolution last interglacial continental climate record

499 and among the few climate records of any resolution from Ethiopia so far published. A deep

500 seismic and near-continuous core record of the last 150,000 years from Lake Tana on the

501 Northwestern Ethiopian highlands used geochemical proxies (sediment Ca/Ti ratio) for

502 climate-driven lake level fluctuations (Lamb et al., 2018). The oldest cave sediment records

503 from the Southeastern Ethiopian highlands goes back only to 63±7 ka (Tribolo et al., 2017).

504   The four phases of the GM1 record are dated at particularly important periods of the last

505 interglacial. Noting the quantified age uncertainties (see section 3.1.4), they provide high

506 resolution snapshots from some critical time-windows. The two long growth phases at ~129

507 ka and ~108 ka, which we conceptualise as being dominated by a sustained porous/matrix

508 flow regime, match maximum summer insolation at 15°N. This suggests that though internal

509 growth variability may be dominated by karst hydrology above the cave, the GM1 growth as

510 a whole and the geochemical proxies were responding to climate forcing.

511   Comparison of the GM1 $\delta^{18}O$ record with the China composite $\delta^{18}O$ record (Hulu and

512 Dongge caves; Cheng et al., 2016), the Soreq cave (Israel) $\delta^{18}O$ record (Bar-Mathews et al.,

513 1999; 2003) (Fig. 7) suggests a similar relationship for all three sites, with wet conditions

514 (lower $\delta^{18}O$ and peak summer insolation) during the ~108 ka growth period, and dry or

515 drying conditions (higher $\delta^{18}O$ and low summer insolation) at the ~120 ka growth period.

516 The Soreq cave $\delta^{18}O$ record from central Israel in the Levant has been shown to indicate

517 enhanced rainfall (Bar-Mathews et al., 1999) and could be a "downstream" indicator of a

518 stronger northeast African monsoon (Tierney et al., 1999). At ~129 ka, the lower $\delta^{18}O$ and

519   wetter conditions in Ethiopia occur at the summer insolation maxima, but may occur before

520   the isotope response observed in Israel and China. Dating uncertainty in the GM1 record

521   prevents a more precise interpretation, but it does raise the possibility of the intensification of

522   the East African Monsoon before other northern hemisphere monsoon systems at the start of

523   the last interglacial. The Lake Tana (Northwestern Ethiopian highlands) sediment Ca/Ti

524   record (Fig. 7) indicates an abrupt increase in moisture even earlier at ~132 ka, leading to

525   stable high lake level conditions during the period ~132 ka to ~95 ka, with some brief dry

526   episodes (Lamb et al., 2018). This period is defined by a generally flat trend of Ca/Ti,

527   truncated by very brief high Ca/Ti excursions. The major wet conditions during the ~129 ka

528   and 108 ka growth phases of GM1 are generally consistent with this predominantly high lake

529   level phase, although they do not particularly correspond to the lowest values of Ca/Ti.

530   As speleothem records of the Holocene from the region (Fleitmann et al., 2007) show,

531   decreasing $\delta^{18}O$ values during the early Holocene indicate a rapid northward migration of the

532   summer ITCZ and intensification of the rain belt of the Indian Summer Monsoon. On the

533   other hand, the southward migration of the ITCZ during the middle to late Holocene, marked

534   by increasing $\delta^{18}O$ values in speleothems, led to weakening of the associated summer

535   monsoon. Similar studies from Madagascar (e.g., Voarintsoa et al., 2017) have also shown

536   the link between speleothem $\delta^{18}O$ values and the ITCZ migration. The ~129 ka growth phase

537   of GM1 could therefore be effectively considered as an annual record of relative ITCZ

538   position ~129 ka BP.

539   Ethiopia/northeast Africa has been considered as the origin of Anatomically Modern

540   Humans (White et al., 2003; McDougall et al., 2005) and possibly the region out of which

541   Anatomically Modern Humans dispersed during MIS 5-3 or during earlier migration episodes

542   (*cf*. Hershkovitz et al., 2018). Considering the significant debate about the role climate

543   variability played in human evolution and dispersal of Anatomically Modern Humans "Out of

544 Africa" (e.g., Tierney et al. 2017 and references therein; Lamb et al., 2018), high-resolution

545 speleothem records from Ethiopia and northeastern Africa such as those of GM1 can shed

546 light on this debate. The GM1 record for instance shows that the earliest human migration

547 during MIS 5, confirmed by the presence of ~110-80 ka old Anatomically Modern Human

548 fossils in Israel (Grün et al., 2005), occurred during a major wet event in northeast Africa.

549 The Tana Lake record confirms that the northwestern Ethiopian highlands experienced

550 relatively stable moist climate during MIS 5c-e (Lamb et al., 2018). This supports earlier

551 conclusions that human migration occurred during humid conditions, as such conditions

552 provided humans "green corridors" to overcome inhospitable deserts (e.g., Timmerman and

553 Friedrich, 2016). However, the major episode of human migration occurred during 50-75 ka

554 (Nielsen et al., 2017), and marine records from the Gulf of Aden show that this migration

555 event occurred during a sustained dry condition in northeast Africa (Tierney et al., 2017),

556 while the Lake Tana record shows more complex climate variability during this period (Lamb

557 et al., 2018). Future research is required on the speleothems from Goda Mea and Aynage

558 caves, some of which have been dated to the critical period of 120-50 ka.

559

560 **5. Conclusions**

561 Stalagmite GM1 was deposited discontinuously around the time of the last interglacial, at

562 ~129 ka, 120 ka, an undated growth phase, and ~108 ka. Variogram analysis of growth rate

563 shows a low range (20.5 years), some flickering (–0.33) and good information content (56%),

564 indicative of a stalagmite fed by a karst water store of limited volume. Oxygen and carbon

565 isotopes and trace elements generally have low variability, indicative of a second, well-mixed

566 water source feeding the stalagmite. Stalagmite GM1 provides a high-resolution insight into

567 stalagmite hydrogeochemical responses to environmental change prior to growth hiatuses.

568 Multi-decadal variability of frequency 17-18 years, though statistically not significant at 95%

569 confidence, is present, but only in the growth rate time series, and is slightly less than the

570 range in the growth rate record. A climatic or karstic forcing of this spectral frequency cannot

571 be determined.

572   Our conceptual model for the stable isotope, trace element and growth rate records in

573 GM1 allows the interpretation of the stalagmite geochemical time series. Importantly, all

574 three proxies were necessary to adequately understand the processes forcing them, and

575 whether they contained a climatic or karstic signal. Only through this approach were we able

576 to confirm that low $\delta^{18}O$ at ~129 ka and ~108 ka can be attributed to wetter climatic

577 conditions. These two growth phases occur at the same time as solar insolation maxima for

578 15°N, and suggest a direct solar forcing on rainfall in Ethiopia at these times, influencing the

579 northward migration of the ITCZ and the associated rain belt of the Indian Summer

580 Monsoon, of potential relevance for early modern human migration out of the region.

581

References

601   Asrat, A., Baker, A., Umer, M., Leng, M.J., van Calsteren, P., Smith, C.L., 2007. A high-

602   resolution multi-proxy stalagmite record from Mechara, Southeastern Ethiopia:

603   Paleohydrological implications for speleothem paleoclimate reconstruction. Journal of

604   Quaternary Science 22, 53-63.

605   Asrat, A., Baker, A., Leng, M.J., Gunn, J., Umer, M., 2008. Environmental monitoring in the

606   Mechara Caves, Southeastern Ethiopia: implications for speleothem paleoclimate studies.

607   International Journal of Speleology 37, 207-220.

608   Asrat, A., 2012. Speleoseismicity in the Mechara karst, Southeastern Ethiopia. In: Fairchild,

609   I.J., and Baker, A. (2012) Speleothem Science: from Process to Past Environments. Wiley-

610   Blackwell.

611   Baker, A., Asrat, A., Fairchild, I.J., Leng, M.J., Wynn, P.M., Bryant, C., Genty, D., Umer,

612   M., 2007. Analysis of the climate signal contained within $\delta^{18}O$ and growth rate parameters

613   in two Ethiopian stalagmites. Geochimica et Cosmochimica Acta 71, 2975-2988.

614   Baker, A., Asrat, A., Fairchild, I.J., Leng, M.J., Thomas, L.E., Widmann, W., Jex, C., Dong,

615   B., Calsteren, P.V., Bryant, C., 2010. Decadal scale rainfall variability in Ethiopia

616   recorded in an annually laminated, Holocene-age, stalagmite. The Holocene 20, 827-836.

617   Baker, A., Bradley, C. and Phipps, S.J., 2013. Hydrological modelling of stalagmite $\delta^{18}O$

618   response to glacial-interglacial transitions. Geophysical Research Letters 40, 3207–3212.

619   Baker, A., Hellstrom, J., Kelly, B. F. J., Mariethoz, G., Trouet, V., 2015. A composite

620   annual-resolution stalagmite record of North Atlantic climate over the last three millennia.

621   Scientific Reports 5, 10307. http://doi.org/10.1038/srep10307.

622   Baker, A., Flemons, I., Andersen, M. S., Coleborn, K., Treble, P. C., 2016. What determines

623   the calcium concentration of speleothem-forming drip waters? Global and Planetary

624   Change 143, 152–161.

625   Bar-Matthews, M., Ayalon, A., Kaufman, A., Wasserburg, G.J., 1999. The Eastern

626   Mediterranean paleoclimate as a reflection of regional events: Soreq cave, Israel. Earth

627   and Planetary Science Letters 166, 85–95.

628   Bar-Matthews, M., Ayalon, A., Gilmour, M., Matthews, A., Hawkesworth, C.J., 2003. Sea-

629   land oxygen isotopic relationships from planktonic foraminifera and speleothems in the

630   Eastern Mediterranean region and their implication for paleorainfall during interglacial

631   intervals, Geochimica et Cosmochimica Acta 67, 3181-319.

632   Borsato, A., Frisia S., Fairchild I.J., Somogyi, A. Susini, J., 2007. Trace element distribution

633   in annual stalagmite laminae mapped by micrometer-resolution X-ray fluorescence:

634   implications for incorporation of environmentally significant species. Geochimica et

635   Cosmochimica Acta 71, 1494–1512.

636   Cheng, H, Edwards, R. L., Broecker, W. S., Denton, G. H., Kong, X., Wang, Y., Zhang, R.,

637   Wang, X., 2009. Ice Age Terminations. Science 326, 248-252.

638   Cheng, H., Edwards, R. L., Chuan-Chou Shen, C-C., Polyak, V. J., Asmerom, Y., Woodhead,

639   J., Hellstrom, J., Wang, Y., Kong, X., Spötl, C., Wang, X., Alexander Jr., E. C., 2013.

640   Improvements in ^{230}Th dating, ^{230}Th and ^{234}U half-life values, and U–Th isotopic

641   measurements by multi-collector inductively coupled plasma mass spectrometry. Earth

642   and Planetary Science Letters 371–372, 82–91.

643 Cheng, H., Edwards, R. L., Sinha, A., Spötl, C., Yi, L., Chen, S., Zhang, H., 2016. The Asian

644   monsoon over the past 640,000 years and ice age terminations. Nature 534, 640-646.

645 Degefu, M.A., Rowell, D.P., Bewket, W., 2017. Teleconnections between Ethiopian rainfall

646   variability and global SSTs: observations and methods for model evaluation. Meteorology

647   and Atmospheric Physics 129, 173-186.

648 Dominguez-Villar, D., Baker, A., Fairchild, I.J., Edwards, R.L., 2012. A method to anchor

649   floating chronologies in annually laminated speleothems with U-Th dates. Quaternary

650   Geochronology 14, 57-66.

651 Dorale, J. A., Liu, Z., 2009. Limitations of hendy test criteria in judging the paleoclimatic

652   suitability of speleothems and the need for replication. Journal of Cave and Karst Studies

653   71, 73–80.

654 Fleitmann, D., Burns, S. J., Mangini, A., Mudelsee, M., Kramers, J., Villa, I., Neff, U., Al-

655   Subbary, A. A., Buettner, A., Hippler, D., Matter, A., 2007. Holocene ITCZ and Indian

656   monsoon dynamics recorded in stalagmites from Oman and Yemen (Socotra), Quaternary

657   Science Reviews 26, 170–188.

658 Genty, D., Quinif, Y., 1996. Annually laminated sequences in the internal structure of some

659   Belgian stalagmite-importance for palaeoclimatology. Journal of Sedimentary Research

660   66, 275–288.

661 Genty, D., Baker, A., Barnes W.L., 1997. Comparison of annual luminescent and visible

662   laminae in stalagmites. Comptes Rendus Academie Sciences Paris, Serie II 325, 193–200.

663 Johnson, K. R., Hu, C. Y., Belshaw, N. S., Henderson, G. M., 2006. Seasonal trace-element

664   and stable-isotope variations in a Chinese speleothem: The potential for high-resolution

665   paleomonsoon reconstruction. Earth and Planetary Science Letters 244, 394-407.

666   Grün, R., Stringer, C., McDermott, F., Nathan, R., Porat, N., Robertson, S., Taylor, L.,

667   Mortimer, G., Eggins, S., McCulloch, M., 2005. U-series and ESR analyses of bones and

668   teeth relating to the human burials from Skhul. Journal of Human Evolution 49, 316–334.

669   Gunn, J., Baker, A., Asrat, A., 2009. Caves of Mechara: The millennium expedition to

670   Ethiopia. National Speleological Society Journal, June 2009, 4-7.

671   Hellstrom, J., 2003. Rapid and accurate U/Th dating using parallel ion-counting multi-

672   collector ICP-MS. J. Anal. At. Spectrom. 18, 1346-1351.

673   Hellstrom, J., 2006. U–Th dating of speleothems with high initial ^{230}Th using stratigraphical

674   constraint. Quaternary Geochronology 1, 289–295.

675   Hershkovitz, I., Weber, G. W., Quam, R., Duval, M., Grün, R., Kinsley, L., Ayalon, A., Bar-

676   Matthews, M., Valladas, H., Norbert Mercier, N., Arsuaga, J. L., Martinón-Torres, M., de

677   Castro, J. M. B., Fornai, C., Martín-Francés, L., Sarig, R., May, H., Krenn, V. A., Slon,

678   V., Rodríguez, L., García, R., Lorenzo, C., Carretero, J. M., Frumkin, A., Shahack-Gross,

679   R., Mayer, D. E. B-Y., Cui, Y., Wu, X., Peled, N., Groman-Yaroslavski, I., Weissbrod, L.,

680   Yeshurun, R., Tsatskin, A., Zaidner, Y., Weinstein-Evron, M., 2018. The earliest modern

681   humans outside Africa. Science 359, 456-459.

682   Lamb, H.F., Bates, C.R., Bryant, C.L., Davies, S. J., Huws, D. J., Marshall, M. H., Roberts,

683   H. M., 2018. 150,000-year palaeoclimate record from northern Ethiopia supports early,

684   multiple dispersals of modern humans from Africa. Scientific Reports 8, 1077, DOI:

685   10.1038/s41598-018-19601-w.

686   Laskar, J., Robutel, P., Joutel, F., Gastineau, M., Correia, A.C.M., Levrard, B., 2004. A long

687   term numerical solution for the insolation quantities of the Earth. Astronomy and

688   Astrophysics 428, 261-285.

689   Mariethoz, G., Kelly, B., Baker, A., 2012. Quantifying the value of laminated stalagmites for

690   paleoclimate reconstructions. Geophysical Research Letters 39, L05407.

691   McDougall, I., Brown, F. H., Fleagle, J. G., 2005. Stratigraphic placement and age of modern

692   humans from Kibish, Ethiopia. Nature 433, 733-736.

693   Nicholson, S.E., 1996. A review of climate dynamics and climate variability in Eastern

694   Africa, in: T.C. Johnson, E.O. Odada (Eds.), The Limnology, Climatology and

695   Paleoclimatology of the east African Lakes, The International Decade for the east African

696   Lakes (IDEAL), Gordon and Breach, Newark, NJ, 25-56.

697   Nielsen, R., Akey, J.M., Jakobsson, M., Pritchard, J.K., Tishkoff, S., Willerslev, E., 2017.

698   Tracing the peopling of the world through genomics. Nature 541, 302–310.

699   Railsback, L.B., Akers, P.D., Wang, L., Holdridge, G.A., Voarintsoa, N.R., 2013. Layer-

700   bounding surfaces in stalagmites as keys to better paleoclimatological histories and

701   chronologies. International Journal of Speleology 42, 167-180.

702   Taye, M.T., Willems, P., 2012. Temporal variability of hydroclimatic extremes in the Blue

703   Nile basin. Water Resources Research 48, W03513.

704   Tierney, J. E., de Menocal, P. B., Zander, P. D., 2017. A climatic context for the out-of-

705   Africa migration. Geology, doi:10.1130/G39457.1

706   Timmermann, A., Friedrich, T., 2016. Late Pleistocene climate drivers of early human

707   migration: Nature 538, 92–95.

708   Treble, P.C., Schmitt, A.K., Edwards, R.L., McKeegan, K.D., Harrison, T.M., Grove, M.,

709   Cheng, H., Wang, Y.J., 2007. High resolution Secondary Ionisation Mass Spectrometry

710   (SIMS) $\delta 18O$ analyses of Hulu Cave speleothem at the time of Heinrich Event 1.

711   Chemical Geology 238, 197–212.

712   Tribolo, C., Asrat, A., Bahain, J.-J., Chapon, C., Douville, E., Fragnol, C., Hernandez, M.,

713   Hovers, E., Leplongeon, A., Martin, L., Pleurdeau, D., Pearson, O., Puaud, S., Assefa, Z.,

714   2017. Across the gap: geochronological and sedimentological analyses from the Late

715   Pleistocene-Holocene sequence of Goda Buticha, southeastern Ethiopia. PLoS One 12,

716   e0169418.

717 Voarintsoa, N. R. G., Railsback, L. B., Brook, G. A., Wang, L., Kathayat, G., Cheng, H., Li,

718   X., Edwards, R. S., Rakotondrazafy, A. F. M., Razanatseheno, M. O. M., 2017. Three

719   distinct Holocene intervals of stalagmite deposition and nondeposition revealed in NW

720   Madagascar, and their paleoclimate implications. Climate of the Past 13, 1771–1790.

721 White, T. D., Asfaw, B., Degusta, D., Gilbert, H., Richards, G. D., Suwa, G., Howell, F. C.,

722   2003. Pleistocene *Homo sapiens* from Middle Awash, Ethiopia. Nature 423, 742–747.

723

Table Caption

726 Table 1. U and Th isotope data and age determinations (in depth order) for stalagmite GM1.

727   Square brackets indicate activity ratios. Ages shown are corrected for an initial

728      $[^{230}Th/^{232}Th]$ of 1.5 and a 100% uncertainty, which is incorporated into the age

729   uncertainty.

730

731 **Figure Captions**

732

733 Fig. 1. (a) Regional structural setting of Ethiopia showing the location of Mechara. Lake

734   Tana, and the epicentres of the major earthquakes in the Main Ethiopian Rift and the

735   adjoining highlands are marked (Note that earthquake epicentres in the northern Afar

736   depression are not represented). Insets show the mean position of the ITCZ in July

737   (Boreal summer) and January (winter) over Africa; and the mean monthly rainfall (mm)

738   and mean monthly temperature of the Mechara region, at the Bedesa Meteorological

739   Station (1994-2014 data from the Ethiopian Meteorological Agency). Location of (b) is

740   marked by a broken triangle around the location of Mechara (modified from Asrat et al.,

741   2008); (b) The topography, geology, structure and drainage system of the Mechara karst

742   area and locations of the entrances to the caves mentioned in the text; (c) Goda Mea cave

743   (surveyed according to BCRA Grade 3 using tape, compass and clinometer); (d) a

744   photograph of the main chamber of the Goda Mea cave interior showing the collapse

745   chamber on which grew several speleothems following a major fracture system

746   (photograph by J. Gunn). Figures (a) and (b) modified from Asrat et al., 2008; Fig. (c)

747   modified from Gunn et al., 2009.

748   Fig. 2. GM1 hand-section in both scanned image (left) and sketch (middle), showing the four

749   major growth phases, locations of the major and minor growth hiatuses, and sampling for

750   isotopes, trace elements and U-Th analyses. Right: photomicrographs of thick sections

751   from across the major growth hiatuses showing a clear evidence of growth stoppage with

752   no apparent dissolution.

753   Fig. 3. U-Th data for stalagmite GM1. (a) Corrected U-Th ages vs depth for all analyses. (b)

754   ^{238}U concentration vs depth (c) $[^{230}Th/^{232}Th]$ vs depth (d) Initial $[^{234}U/^{238}U]$ vs depth (e)

755   U-Th ages vs depth for samples with $[^{230}Th/^{232}Th]$ >1000. In all plots, the three major

756   hiatuses are shown as vertical dashed lines.

757   Fig. 4. GM1 times series for the geochemical proxies. From top: growth rate, δ^{13}C, δ^{18}O,

758   Sr/Ca, Mg/Ca, Ba, Ca. Note the axis breaks on the x-axis, which permit equal scaling of

759   data on the time axis.

760   Fig. 5. (a) Scatterplot of variogram parameters range, information content and flickering for

761   the 129.3 ka growth phase; (b) Autocorrelation of growth rate, δ^{13}C and δ^{18}O time series

762   for the three U-Th dated growth phases; (c) Spectral analysis on the growth rate and δ^{18}O

763   times series for the 129.3 ka growth phase. The 17-18 years peak, though statistically not

764   significant is marked.

765   Fig. 6. Conceptual model for the deposition of stalagmite GM1.

766   Fig. 7. Comparison of climate proxy records. (a) Insolation at 15°N (Laskar et al., 2004); (b)

767   Chinese composite stalagmite δ^{18}O record (Cheng et al., 2016); (c) Soreq Cave δ^{18}O

768   record (Bar-Matthews et al., 1999; 2003); (d) Lake Tana sediment Ca/Ti record (Lamb et

769   al., 2018); (e) Ethiopian stalagmite δ^{18}O composite. Box-plots show median, inter-

770   quartile range and range for each stalagmite; shading represents different caves (grey –

771   Bero Cave; Green – Rukiesa Cave; Orange – Achere Cave; Cyan – Goda Mea Cave).

772   Note the x-axis break; vertical shading aligns the Ethiopian records to the other time

773   series.

774   Fig. S1. Scatter plot of $\delta^{13}C$ vs $\delta^{18}O$ for all growth phases. 'Hendy tests' along growth

775   laminae are shown in colour (Lines 1-7): their location is shown in Fig. 2 (HL1-HL7).

776