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1 Quantification of Asian monsoon variability from 68 ka BP through 2 pollen-based climate reconstruction 3

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19

20 **ABSTRACT**

21 The glacial-interglacial variability of precipitation and its driving mechanism in monsoonal
22 regions has long been a subject of debate. However, there are few records of quantitative
23 climate reconstruction dating to the last glacial cycle in areas dominated by the Asian summer
24 monsoon. Here, using a pollen-based quantitative climate reconstruction based on three sites
25 in areas exposed to the Asian summer monsoon, we demonstrate that climate has undergone
26 great variability over the past 68 ka. The differences between the last glacial and the Holocene
27 optimum could have been as much as 35%–51% for precipitation, and 5–7 °C for mean annual
28 temperature. Our findings also reveal regional heterogeneity during the abrupt climate events
29 of Heinrich Event 1 and Younger Dryas, that drove drier conditions in southwestern China
30 dominated by the Indian summer monsoon, and a wetter climate in central eastern China. The
31 pattern of variation in reconstructed precipitation, exhibiting strong glacial-interglacial variability,
32 is broadly consistent with the stalagmite d18O records from Southwest China and South Asia.
33 Our results of reconstruction quantify the sensitivity of the MIS3 precipitation to orbital insolation
34 changes, and highlight the prominent influence of interhemispheric temperature gradients on
35 Asian monsoon variability.

36 Comparison with transient simulations and major climate forcings has shown that the mode
37 of precipitation variability during the transition from the last glacial maximum to the Holocene
38 has been significantly modulated by weak or collapsed Atlantic meridional overturning
39 circulation events in addition to insolation forcing.

40

41 **INTRODUCTION**

42 The Asian summer monsoon, comprising the Indian Summer Monsoon (ISM) and the East
43 Asian Summer Monsoon (EASM), is a critical climate system in the understanding of orbital-
44 scale climate variability driven by changes in the land-sea thermal contrast and atmospheric
45 pressure [1]. Among the climate variables, precipitation reconstruction is a key process in
46 revealing the summer monsoon's evolution and variability, while stalagmite d18O records have
47 been considered to be a significant proxy indicator [2,3]. In the Asian monsoon region, a great
48 number of stalagmite d18O records have revealed the close relationship between monsoon
49 variations and solar insolation on orbital-millennial scales, showing a significant precession
50 cycle [4]. The synthesized Chinese stalagmite data show that d18O records are a valid indicator
51 of EASM intensity or integrated moisture transport, and the d18O variations of Chinese
52 stalagmites cannot be simply interpreted as local precipitation, but rather as a mean state of
53 summer monsoon intensity [5,6]. However, the temporal and spatial consistency of the d18O

54 records with strong precession cycles are notably different from the significant 100 ka cycles
55 derived from the marine and loess records [7]. Such contradiction leads to the driving
56 mechanism for glacial-interglacial cycles and the role of internal forcings such as ice sheet size,
57 CO₂ concentration, and millennial scale perturbations remains unclear [8,9]. Moreover, the high
58 consistency of Chinese stalagmite δ¹⁸O records as a mean state of summer monsoon
59 variability cannot directly reflect the regional heterogeneity of rainfall

60 in the Asian Monsoon Domain (AMD) [10]. In this respect, the application of quantitative
61 climate reconstructions based on pollen data can reflect regional heterogeneity and validate
62 the difference between monsoon intensity and rainfall amount. Although quantitative
63 precipitation reconstruction studies at a global or continental scale exist [11], few have focused
64 on quantitative data from the last glacial period and the Holocene in monsoonal China. In recent
65 decades, efforts have been made towards quantitative climate reconstruction in eastern Asia,
66 but most studies have been confined to the Holocene and located in marginal areas influenced
67 by the summer monsoon, such as northern China and the Tibetan Plateau [12,13]. One study
68 based on the Dajiuhe pollen record from east-central China covering the last 16 ka shows that
69 the annual precipitation has varied from about 900 mm during the last deglaciation (16–10 ka
70 BP) to 1256 mm during the Holocene optimum (10–7 ka BP) [14]. On the Tibetan Plateau, where
71 the summer monsoon influence is relatively weak, the pollen-based precipitation reconstruction
72 revealed extremely dry conditions at the Last Glacial Minimum (LGM), being only 32% of the
73 present-day rainfall [13]. However, high-resolution climate reconstruction within the humid
74 monsoonal area is still lacking. This greatly hinders our understanding of the driving
75 mechanisms responsible for orbital and abrupt climate changes, and our ability to identify the
76 possible existence of phase-relationships in the monsoon climate system.

77 Here, we present quantitative reconstructions of precipitation and temperature for the past
78 68 ka using representative pollen records from three locations along the summer monsoon
79 pathway in southern China. Two records are located directly in the moisture pathway of the ISM
80 in southwestern (SW) China, while one is located in central eastern China where climate is
81 influenced not only by the ISM, but also by the EASM (Fig. 1). Using a novel statistical approach
82 to pollen-based climate reconstruction, we obtained very consistent results that allowed us to
83 quantify precipitation variability, decipher regional heterogeneity, and evaluate the speleothem
84 proxy records. Our high-resolution climate reconstruction, associated with simulation using a
85 transient model, provides robust evidence for a better understanding of the spatiotemporal
86 hydroclimate changes in the Asian monsoon system.

87

88 **Fig. 1. Location map of the three pollen records (red dots) and the cave sites along the monsoon trajectory**
89 **(green dots).** The three strategic pollen sites in red dots used for quantitative climate reconstruction are TCQH, CH
90 and DJH, of which monthly climate parameters are shown on the right. The green dots mark the caves of eastern
91 coastal Bay of Bengal (CBoB) (Liu, 2020), Xiaobailong (XBL) (Cai, 2016), southeastern Yunnan (SEY) (Liu, 2020),
92 Dongge (DG) (Wang, 2001), Yangzi (YZ) (Wu, 2020), Sanbao (SB) (Wang, 2008) and Hulu (HL) (Cheng, 2016). The
93 blue dot under the red TCQH indicates the δD_{wax} record from the same lake (Zhao, 2020). The white arrows display
94 the atmospheric circulation systems and June-July-August mean 850-hPa streamlines for the period 1970-2000.

95

96 **MATERIAL AND METHODS**

97 **Regional setting and core sites**

98 The quantitative climate reconstruction strategy we followed was based on three pollen
99 records. The longest one, covering the last 68 ka, is Tengchong-Qinghai (TCQH) [20] from a
100 volcanically dammed lake (25.1312°N; 98.5731°E; 1885 m) situated to the west of the

101 Gaoligong Mountains in Yunnan Province of SW China. The local climate is dominated by
102 Indian monsoon circulation, with humid conditions from May to October, and a dry season from
103 November to April. The mean annual temperature (MAT) of the core locality is approximately
104 15.4 °C, and the mean annual precipitation (MAP) is approximately 1532 mm, with a peak in
105 July (>300 mm). The second core was obtained from the protected Caohai (CH) lake of western
106 Guizhou Province (26.8140°N; 104.3405°E; 2165 m). This site, also in the southeastern margin
107 of the Tibetan Plateau of SW China, experiences weaker monsoon rainfall intensities. However,
108 together with the TCQH record, this site provides information on summer rainfall levels along
109 the ISM moisture trajectory. The MAT in the study locality is approximately 10.9 °C, and MAP
110 is approximately 974 mm, with a peak in June and July (>160 mm).

111 The Dajihu (DJH) record [21] originates from a karst lake in the Shennongjia mountains
112 near the Middle Yangtze River of central China (31.4833°N; 110.0666°E; 1751 m), close to the
113 northern boundary of the subtropical zone. The DJH wetland is 1770–1800 m above sea level
114 and surrounded by mountains with heights in the range 2200–2600 m. The regional climate is
115 characterized by a transition between subtropical and warm temperate regimes. Under the
116 strong influence of the EASM, the area has short, warm and wet summers, and long, cold, dry
117 winters. Presently, the MAT is 7.4 °C and the annual precipitation is 1379 mm, 40%–50% of
118 which falls in summer. Annual evaporation varies from 500 to 800 mm, indicating relatively high
119 humidity.

120

121 **Fig. 2. Pollen-derived quantitative MAP (blue lines) and MAT (red lines) reconstructions and their significant**
122 **tests of the three records.** The spatial distribution maps of MAP and MAT in eastern Asia and location of the study
123 sites (TCQH, CH and DJH) are also shown. Root mean square error (RMSE) values for each reconstructed climate
124 variable are shown (gray lines). In the histograms on the right, the brown bars indicate proportions of variance explained
125 by 999 calibration functions trained with random environmental data. The proportions of the variance (95%) and the
126 first axis of a PCA of the record are marked respectively with red and black dotted lines.

127

128 **Age models and pollen records**

129 The TCQH core was drilled using a gravity piston corer under water 2.13 m deep. The core was
130 13.2 m long and the sediments were relatively uniform, except for a possible volcanic ash layer
131 at 6.1–6.6 m. In total, 261 samples at 4–5 cm intervals were collected for pollen analysis. The
132 chronology of the core was determined using accelerator mass spectrometry radiocarbon
133 (AMS14C) dating of 14 samples of plant fragments and bulk organic sediments by Beta Analytic
134 Inc., Miami, USA (Fig. S1a and Table S1 online). However, there were few dating samples for
135 the glacial period, particularly for the interval older than 30 ka BP. In order

136 to make up for this deficit, the paleomagnetic datapoints were used to provide additional age
137 control points for the lower section of the core [20]. The relative paleomagnetic intensity (RPI)
138 curve of the TCQH core was compared with curves obtained from analyses of South China Sea
139 sediments [22] and the global RPI stacks [23]. Four recognized RPI lows correlated with those
140 from Sint-200, at 16, 30, 40, and 62 ka, respectively corresponding to four geomagnetic
141 excursions. The first two were in good agreement with the 14C dating results. Therefore, the
142 two deeper age controls (#40 and 62 ka BP) were used for our chronological framework [20].
143 The resulting age-depth model suggested that the core covered the last 68 ka BP,
144 corresponding to marine isotope stages MIS4 to MIS1. The average temporal sampling
145 resolution was about 250 years.

146 In 2013, a 4.05 m continuous sediment core was taken from the CH wetland. All sediment was

147 sampled at 1–2 cm intervals, with a total of 226 samples prepared for pollen analysis. Nine
148 samples were taken for AMS 14C dating. Based on the calibrated ages, the time span covered
149 by the CH record was approximately 22 ka BP, with a sampling resolution of about 94 years.
150 A continuous and undisturbed 6.0 m core was collected from a peat bog at the DJH site using
151 a Russian-type peat corer [21]. A total of 153 sediment samples were taken at 2–5 cm intervals
152 from this core. To obtain a chronology for the sequence, seven bulk peat and wood samples
153 were taken for AMS 14C dating. The dating results showed that the core covered the last 42 ka
154 BP, with a sampling resolution of about 270 years. The two ages (10.03 and 13.37 ka BP) at
155 depths of 163 and 193 cm basically defined the range of Younger Dryas (YD) event. In addition,
156 a published pollen-based quantified precipitation of the same lake based on more dating results
157 [14] was used to compared with our data. At the three study sites described above, all 14C
158 dates were calibrated to calendar ages using the Bacon software package and IntCal20
159 calibration curves [24]. The three records include a total of 639 pollen samples, corresponding
160 to a centennial resolution.

161 For all pollen samples from the three cores mentioned above, 1–2 g of sediment was weighed
162 and processed by heavy liquid separation [25]. Briefly, after adding 15% HCl to remove
163 carbonates, followed by 10% KOH to remove humic substances, each sample was subjected
164 to heavy liquid flotation, then mounted in glycerin jelly and examined under a light microscope.
165 On average, 521 terrestrial pollen grains (with proportionally corresponding numbers of fern
166 and algal spores) per sample were counted. All percentages of pollen grains and spores (of
167 trees, shrubs, terrestrial herbs, wetland herbs (Cyperaceae), ferns, and algae) quoted here are
168 based on the total sum of pollen from terrestrial taxa. The data were graphed
169 using Tilia (Figs. S2–S4 online) [26].

170 **Pollen-based quantitative climate reconstruction**

171 To identify the climate variables that most strongly influence the regional vegetation, six climate
172 variables, including MAT, mean temperatures of warmest month (MTwa) and coldest month
173 (Mtco), MAP, mean precipitations of warmest month (MPwa) and coldest month (MPco), were
174 initially selected to characterize the climate at each modern pollen sampling site. A correlation
175 analysis of these six climate variables was carried out for the three study cores using the
176 modern pollen datasets (Fig. S5 online). Result showed that the best variables of the three
177 cores were MAT and MAP (Table S2 online), and the seasonal climate parameters had a high
178 correlation with them. Therefore, MAT and MAP were considered appropriate variables for the
179 development of calibration models.

180 The continuous improvement of modern pollen databases in East Asia provides a good
181 foundation for this quantitative paleoclimate study [27]. The selection of the distribution range
182 of modernpollen samples is a key step in the pollen-based quantitative climate reconstruction.
183 Based on the chosen climate variables and the zonal vegetation types in the regions, the
184 modern pollen sites with 9 different radiuses (from 700 to 1500 km at 100 km spacing) were
185 compared. This comparison showed that the results of quantitative reconstruction did not
186 improve linearly with the spatial increase of selected distribution area. Along with the stepwise
187 increase of the spatial extent of the calibration subset when the radius is increased from 700 to
188 1500 km, values of R2 peaks and the values of RMSEP are relatively low when the radii are
189 1000, 1100, and 1200 km, especially for the 1100 km radius subset, suggesting that the dataset
190 ($r = 1100$ km) exhibits the best model performance for all the three sites (Table S2 online).

191 Therefore, in this study, we used the reference datasets within a radius of 1100 km around the
192 focal sites to provide high-quality information on pollen-climate relationships to enable
193 quantitative paleoclimate reconstruction (Fig. S6 and Table S2 online). In addition, we tested
194 the suitability of several calibration models for the selected modern pollen database. The
195 Locally Weighted-Weighted Average Partial Least Squares (LW-WAPLS) calibration model was
196 ultimately chosen because it yielded a significantly lower root mean square error (RMSE) than
197 the other tested transfer function methods (Table S3 online). The results of leave-one-out cross
198 validation of the LW-WAPLS model are summarized in Table S4 (online). It was found that the
199 model's RMSE values were lowest and its R2 values were highest when the reconstruction was
200 based on #30 analogs (k = 32, 30 and 28). The "Rioja" package in "R" was used to establish
201 and check the pollen-climate models, and the "analog" package was used for the Locally
202 Weighted-Weighted Average (LW-WA) and LW-WAPLS models [28].

203 **Model simulations**

204 The National Center for Atmospheric Research Community Climate System Model (NCAR-
205 CCSM) is a global, coupled ocean/atmosphere/sea-ice/land-surface climate model [29]. For the
206 present study, we used the result of a transient simulation in a fully coupled climate model,
207 version 3 (NCAR-CCSM3) at low resolution [30], which can simulate the climate evolution of
208 the last 300 ka. The atmospheric model, Community Atmospheric Model 3 (CAM3), and the
209 land model have a #3.75° latitude-longitude resolution (T31). The ocean model is the NCAR
210 implementation of the Parallel Ocean Program (POP), with a #3.6° longitudinal resolution.

211 The sea-ice model is the NCAR Community Sea-Ice Model (CSIM), a dynamic-thermodynamic
212 model that includes a subgrid-scale ice thickness distribution. Several 3000-year-long transient
213 simulations were carried out, in which the orbital forcing (ORB), green-house gases
214 concentration (GHG), and ice-sheet volume (ICE), differently combined (ORB, ORB + GHG,
215 ORB + GHG + ICE), were ascribed to the last 300,000 years (300 ka BP to present day, 0 ka
216 BP), but with an acceleration factor of 100. The low resolution NCAR-CCSM3 generates glacial-
217 interglacial variability comparable to modern observed records [30] as well as to some key
218 proxy records since the LGM [31]. In this study, we used the simulated
219 results of the monsoon region in eastern Asia (between 20°–55°N and 90°–120°E) over the last
220 70 ka for comparison with the pollen-based quantitative climate reconstruction.

221

222 **RESULTS**

223 The three records studied were from the Tengchong–Qinghai volcanic lake (TCQH), the Caohai
224 mountain wetland (CH), and the Dajihu mountain wetland (DJH), covering the last 68, 22, and
225 42 ka respectively (Fig. 1 and Fig. S1 online). At the two ISM sites (TCQH and CH), the
226 vegetation changes from the glacial to Holocene were characterized by the transformation from
227 a sclerophyllous forest dominated by Tibetan evergreen oak, named *Quercus* (T/E), to a
228 broadleaved forest dominated by *Castanopsis*/*Lithocarpus* (Figs. S2 and S3 online) [20]. In
229 contrast, the DJH record reflected a change from a mixed alpine coniferous/sedge meadow to
230 temperate deciduous forest due to its more northerly latitudinal position and the climatic
231 dominance of the EASM (Fig. S4 online) [21].

232 Two climate variables—MAP and MAT—were quantitatively reconstructed (Fig. 2). The
233 predicted quantitative results of these climate variables, using LW-WAPLS calibration models,
234 gave high correlation coefficients between observed and predicted values ($R^2 > 0.8$). The

235 reconstructed precipitation records for TCQH show a relatively low level of precipitation
236 (average 1026 mm) in the earliest interval (68–57 ka BP) corresponding to MIS4, with
237 significantly more precipitation (average 1258 mm) during MIS3 between 57 and 43 ka BP. All
238 three records consistently indicate that precipitation fell to a minimum (average 927 mm in
239 TCQH, 542 mm in CH, and 903 mm in DJH) at 28–16 ka BP, which coincides perfectly with
240 the timing of the global LGM [33]. The variability in precipitation between the LGM and the
241 Holocene in the three cores ranged by as much as 500–940 mm, the largest being in the case
242 of CH. It is worth noting that the precipitation reconstructed with the DJH core clearly increased
243 during the period around 18–16 ka BP (H1), which is highly consistent with the Humification
244 result (Fig. S7 online) and previous high-resolution reconstruction results
245 of the same site [14].

246 In order to understand better the relationship between reconstructed MAP and simulated annual
247 and seasonal climate changes in each study site, we compared the simulated precipitation (in
248 summer, winter and annual) using the TraCE21 model for the locations of TCQH, CH, and DJH
249 (Fig. S8 online). The result shows that the pollen-based MAP matches the simulated
250 precipitation in summer and the whole year better than that in winter, indicating a strong
251 summer monsoon rainfall signal in pollen assemblages. This implies that the pollen proxy is
252 also a good summer monsoon indicator and comparable with stalagmite records. However,
253 there is a clear decoupled variation between the pollen proxy and the model's result during the
254 LGM in all three sites. Such inconsistent changes may imply, on the one hand, that the
255 simulated local precipitation for the LGM was largely overestimated (particularly in TCQH and
256 DJH) or, on the other hand, decreased temperature at LGM might interfere, to some extent,
257 with the result of pollen-based precipitation reconstruction because thermal condition is one of
258 the main factors of vegetation transformation. We also quantitatively reconstructed MAT, which
259 yielded compelling results because the modern pollen dataset covers a large area with high
260 spatial resolution. Its overall variation trends were found to be in good agreement with the
261 results derived from biomarkers (brGDGTs) [18] for the same lake (TCQH), although there were
262 some differences in amplitude due to seasonal biases [34] (Fig. S9 online). The two southwest
263 records show much lower temperatures than at present at the LGM i.e., 5.7 °C (TCQH) and
264 7.2 °C (CH). In central China (DJH), where the climate is dominated by the EASM, it was
265 #4.7 °C lower than at present, which is close to the model's results [11,35]. The greater lowering
266 of temperature at the LGM observed at the TCQH and CH sites in SW China than at DJH in
267 central China are likely due to the extent of mountain glaciers along longitudinal valleys in the
268 Hengduan mountain ranges on the southeastern edge of the Tibetan Plateau. We note that the
269 elevations of our study sites are all above 1700 m, and suggest that these results probably
270 reflect a steeper lapse rate in low-latitude mountains during the LGM [36] due to elevation-
271 dependent temperature amplification by surface albedo caused by changes in mountain
272 glaciers [37]. Temperature and rainfall quantified in the Holocene Thermal Maximum (HTM),
273 concentrated between 9 and 5 ka BP, were basically in phase and approximately
274 contemporaneous at the three different sites.

275

276 **DISCUSSION AND CONCLUSION**

277 **Hydroclimate changes and regional heterogeneity of the Asian monsoon system**

278 In areas where the current annual precipitation exceeds 1300 mm (e.g., TCQH and DJH), the

279 absolute reductions in precipitation during the LGM were quite similar (about 35% lower). The
280 larger decrease in precipitation (51%) at the CH site, being spatially consistent with the present-
281 day low rainfall, suggests that the winter monsoon had a stronger influence during the LGM in
282 this location. The orbital-scale variation of precipitation derived from the three study records
283 from SW and central eastern China are basically coupled, although their absolute amounts
284 differ (Fig. 2 and Table S5 online). This is broadly in agreement with the hypothesis derived
285 from orbital-scale $\delta^{18}O$ data from stalagmites, which suggests that variations in monsoon
286 strength are driven by large-scale atmospheric dynamics associated with changes in solar
287 radiation, rather than precipitation at the regional scale [38]. However, some reconstructed and
288 modeled monsoon rainfall in southeastern (SE) China suggests the strongest EASM intensity
289 occurred in the middle or late Holocene [39,40], which differs from speleothem $\delta^{18}O$ data. Our
290 data, with relatively low resolution for the Holocene interval, suggest that precipitation was
291 generally high in the middle Holocene over the study areas, while the late Holocene saw a more
292 pronounced drop in DJH rainfall in central China than in SW China. This indicates that there
293 was interregional diversity in the Holocene rainfall patterns of southern China. This may be
294 attributed to the position of the westerly jet stream [40], and/or a stronger Western Pacific
295 Subtropical High (WPSH) and southerly monsoon wind and moisture divergence during mid-
296 to-late Holocene producing less moisture divergence and reduced rainfall in SE China [10].
297 In addition, we observed clear indications of the abrupt climate changes associated with the
298 Heinrich event 1 (H1) and YD. However, these millennial-scale abrupt climate perturbations in
299 the DJH records of central eastern China were characterized by increased rainfall rather than
300 the reduced rainfall observed in SW China, indicating precipitation in antiphase. In particular,
301 DJH records show that precipitation increased from #860 mm during the LGM to #1020 mm at
302 the H1 event, exhibiting a clear antiphase with the lowest stand of rainfall amount over the ISM
303 region (i.e., in the TCQH and CH records). Such hydroclimatic heterogeneity over the East
304 Asian region is consistent with the modern-day spatial pattern of summer rainfall variability over
305 East Asia. This is represented by a meridional “tripole” pattern of precipitation, with increased
306 rainfall over northeastern and southwestern China, and decreased rainfall over central eastern
307 China [32,41]. Our age-depth model of the DJH record compared with another record with more
308 ^{14}C dating data [34] produces an accurate framework for capturing the YD event, which shows
309 a coincidence between an abrupt colder climate and higher precipitation at the YD. However, it
310 should be pointed out that both records from DJH wetland lack direct dating results close to the
311 H1 event, so there may be age uncertainties due to relatively low time-resolution. This needs
312 to be confirmed in the future.

313 Our findings agree with the results from the multiproxy speleothem records for the middle
314 Yangtze region that indicate a wetter central eastern China during North Atlantic cooling events,
315 despite the oxygen isotopic records suggesting a weaker monsoon [32]. Our quantitative result
316 is compelling because it demonstrates that a weakened Asian summer monsoon led to drier
317 conditions in SW China, while central China was relatively wetter during these abrupt events.
318 The jet transition hypothesis concerning shifts in the meridional temperature gradient and
319 seasonal displacement of the westerlies has been proposed as an explanation for the observed
320 antiphase of rainfall and for the East Asian hydroclimate simulation that predicts a slowdown of
321 the Atlantic Meridional Overturning Circulation (AMOC) that is representative of the H1 event
322 and YD [32,42].

323 **Pollen-based climate reconstruction in comparison with speleothem d18O records**

324 The highly coupled orbital-scale variation of precipitation evident from the three pollen records
325 in southern China shows a significant variability from the last glacial period to the Holocene. In
326 general, its pattern agrees well with d18O records from most stalagmites in eastern Asia which
327 have been considered as the index of monsoon intensity. The overlap between the speleothem
328 records and the reconstructed MAP is broadly coherent in most records from regions both ISM
329 and EASM dominances, despite some biases during the glacial period (Fig. 3). Taking into
330 account the significant difference between the d18O records from southwest and southeast of
331 Chinese mainland [10,17], we compared the speleothem d18O records separately from regions
332 of ISM and EASM domains (Fig. 3). On glacial-interglacial timescales, the records from the
333 Xiaobailong (XBL) [17] and southeastern Yunnan (SEY) (unnamed cave) [16] in SW China and
334 the CBoB caves (Hin Thun No.2 cave, Padamyca cave, Phabaung Gu cave, and Ya Thea Phyan
335 Gu cave) near the Bay of Bengal [16] displayed distinct differences in d18O values between
336 the last glacial period (MISs 4–2) and the Holocene, indicating relatively weak ISM during the
337 glacial period.

338 This is in good agreement with our quantified precipitation and previously observed evidence
339 from the Chinese loess [43]. In particular, the integrated d18O data from the CBoB records
340 located near the Bay of Bengal [16] are highly coherent with our reconstructed precipitation
341 (Fig. 3e), indicating that the CBoB caves are located at a key location in the main moisture
342 transport path from the Indian Ocean to the study areas. It has been found that the records
343 close to the monsoon moisture source demonstrate larger glacial-interglacial variability, which
344 then decreases landward because of the moisture pathway effect in the stalagmite d18O data
345 [16]. It is presumably correct that the massive shelf exposure influence (such as Sunderland)
346 on tropical Indo-Pacific hydroclimate was large at LGM, affecting both regional ocean and
347 atmosphere circulations, and amplifying the glacial-interglacial cyclicity [44].

348 In SE China, the speleothem d18O records from Dongge (DG) and Hulu (HL) caves [2,3] show
349 greater differences from our data than in SW China. A climate model with an embedded oxygen-
350 isotope model shows that the moisture advection path from Indian Ocean to eastern China
351 moved equatorward and strengthened over the southern Bay of Bengal during the LGM,
352 contributing only #30% and 15% of the LGM precipitation at DG and HL, respectively [45]. A
353 recent simulation using an isotope-enabled general circulation model has revealed that LGM
354 summer precipitation decreased substantially over the Indian Ocean and South Asia, but
355 increased in East Asia (around east of 105°E) [17]. These modeling results have shed light on
356 how glacial-interglacial changes in monsoon rainfall are manifested in isotopically different
357 ways between SW and East China. According to the Rayleigh distillation model, temperature is
358 one of the factors responsible for the differences in d18O between the LGM and the present on
359 land in Southeast Asia [16].

360 Our estimate of MAT between the last glacial period and the Holocene is about 5–7 °C at the
361 ISM-dominated mountain sites on the southeastern margin of the Tibetan Plateau, which is
362 consistent with previously reported temperature estimates for the Tibetan Plateau [46] and the
363 modeled temperature for SW China [47]. Pronounced glacial climate changes may have
364 resulted in greater isotopic distillation than today and thus progressively lower d18O values for
365 precipitation as one moves downstream along the trajectory of the ISM [6].

366 For orbital and millennial timescales, a comparison of our quantified precipitation with

367 speleothem d18O records suggests that the variations in d18O records may reflect strong
368 (weak) summer monsoon events as a whole, rather than amount of local rainfall at an individual
369 site [8,17]. Despite the variation in d18O of precipitation indicating a mean state of summer
370 monsoon intensity, the records from the locations near the moisture source area (e.g., CBoB)
371 may exhibit a closer variability with local amount of precipitation as shown by our pollen-based
372 precipitation reconstruction. The coupled relationship with stalagmite d18O data from SW China
373 and the coastal area near the Bay of Bengal demonstrates the high reliability of pollen-based
374 precipitation reconstruction, which can be used as a standard precipitation mode in monsoonal
375 region dominated by ISM.

376

377 **Fig. 3 Comparison of the quantitative climate reconstructions with the proxy records from ice cores, deep-sea**
378 **marine sediments of Indo-Pacific oceans and stalagmites of Chinese caves.** (a) Stalagmite $\delta^{18}\text{O}$ records from
379 Dongge & Hulu caves (Cheng, 2016) and summer mean insolation (June-July-August) at 30°N (Laskar, 2004); (b-c)
380 Pollen-based climate reconstructions for MAP (b) and MAT (c) (data integration using Z-score standardized method)
381 of the three sites (this study); (d) SST_{Mg/Ca} in MD05-2901 (the western SCS), ODP1145 (the northern SCS) (Wang
382 2012); (e) Marine $\delta^{18}\text{O}$ *G. ruber* records of NIOP905 (Jung, 2009) and SST from core SK237-GC09 (Raza, 2017); (f)
383 CH₄ values of the EDML ice core (EPICA Community members, 2006) and atmospheric CO₂ concentrations (Ahn,
384 2008); (g) $\delta^{18}\text{O}$ values of the EDML ice core ((EPICA Community members, 2006). The yellow backgrounds indicate
385 the high value periods of boreal summer insolation at orbital scale.

386

387 **Model-proxy comparison and climate forcings**

388 A standardized approach was used to integrate quantitative precipitation data in eastern
389 Asia, including our three sites and Gonghai lake (GH), the latter on the edge of the Chinese
390 Loess Plateau near the EASM boundary [12]. In the two sub-systems (i.e., ISM and EASM) of
391 Asian summer monsoon, we found that the general trends of precipitation variation are broadly
392 coupled. The only difference between the two monsoon domains is that the absolute rainfall
393 amount was much higher in SW China (ISM) than in central-east (DJH) and northern China
394 (GH) dominated by EASM (Fig. 3). In addition, the variation pattern of MIS3 rainfall is better
395 matched with the summer insolation at 60°N than at 30°N, likely indicating that high-latitude
396 forcing was more important in driving the hydroclimate changes in the Asian monsoon domain
397 (Fig. 4A, b). Several proxy data from loess and stalagmite records also suggest that solar
398 radiation at high latitudes may dominate the precipitation variability of the Asian summer
399 monsoon [54]. The traditional theory postulates that changes in the high-latitude summer
400 insolation drove variations in glacial-interglacial ice sheet extents which, in turn, triggered
401 changes in other parts of the global climate system [55]. Over recent decades, however,
402 another hypothesis has highlighted that monsoon variation is largely driven by inter-tropical
403 insolation differences [56]. It was found that the summer insolation at 60°N high-latitude and
404 the July 21 insolation differential between tropics 30°N and 30°S were virtually identical in terms
405 of their phases [8], both of which are in good agreement with our reconstructed precipitation
406 variations (Fig. 4). In general, June insolation at high-latitude of the Northern Hemisphere is
407 considered as the critical forcing driving changes in the ice sheet extent, while insolation
408 differential between interhemispheric tropics triggers monsoon variations on the orbital-scale,
409 such as the Asian summer monsoon [57]. The consistency of our data with the “Summer Inter-
410 Tropical Insolation Gradient” (SITIG) first proposed by Reichert [56] likely supports the
411 hypothesis that the insolation differential between interhemispheric tropics drives Asian
412 monsoon variations on the orbital-scale [8]. Compared with the traditional theory, “SITIG”
413 appears to be a more integrated forcing of monsoons, because the insolation difference
414 involves both “pull” and “push” forcings from two hemispheres that drive monsoon changes.

415 Our evidence, tightly coupled with the SITIG, highlights the influence of interhemispheric
416 temperature gradients on monsoon variability. The consistency with the marine proxy records
417 from the Indian Ocean, the main source area of Asian summer monsoon rainfall, also agrees
418 with the mechanism triggered by the inter-tropical insolation gradient [49,50].

419 In matching our data and the insolation cycles, we found some obvious time lags in the
420 reconstructed precipitation. For instance, during the last glacial termination, there was a distinct
421 time lag (4–5 ka) between the lowest amount of precipitation (corresponding to H1 event) and
422 the lowest boreal summer insolation. The drier conditions that prevailed during H1 have been
423 found in many records in eastern Asia and across the world [58]. To understand further the
424 forcings of precipitation changes in AMD, we used the transient simulation in a fully coupled
425 climate model of low resolution [59], which can simulate the climate evolution of the last 300
426 ka. The National Center for Atmospheric Research Community Climate System Model 3
427 (NCAR-CCSM3) forcing experiments under orbital (ORB), greenhouse gases (GHGs), and ice
428 sheet (ICE) were created individually. The orbital modulation of glacial-interglacial precipitation
429 variability shown in the models may be useful for a correct interpretation of monsoonal changes
430 in high-resolution pollen records. As expected, the solo-forcing experiments clearly show the
431 dominance of orbital effect on AMD precipitation changes over the last glacial-interglacial cycle,
432 which are basically consistent with our quantitatively reconstructed MAP pattern over the last
433 68 ka (Fig. 5). Although the NCAR-CCSM3 produces precipitation variability that is generally
434 comparable to our combined proxy records, it still has a 4–5 ka difference between them at the
435 lowest values during the LGM.

436

437 **Fig. 4 Comparison of pollen-based precipitation reconstructions with speleothem $\delta^{18}\text{O}$ records from mainland**
438 **Southeast Asia over the past 40 kyr. (a) $\delta\text{D}_{\text{wax}}$ record of TCQH (Zhao, 2020), Stalagmite $\delta^{18}\text{O}$ values from Dongge**
439 **& Hulu (b) (Wang, 2001; Cheng, 2016), YZ (c) (Wu, 2020), SEY (d) (Liu, 2020) and CBoB (e) (Liu, 2020); The overlap**
440 **of speleothem records obtained from the CBoB on the reconstructed MAP based on pollen records from TCQH of CH**
441 **of ISM dominance (f) and DJH of EASM dominance (g) with curves of light purple (Sun, 2019) and dark purple (this**
442 **study).**
443

443

444 The time lag of pollen-based precipitation behind the transient model and/or solar radiation
445 during the last glacial termination (H1), together with the YD event, can be explained by the
446 high and low latitude climate teleconnections, where the abrupt input of fresh water to the North
447 Atlantic led to reductions in the AMOC and resultant lowering of sea surface temperatures
448 (SSTs) [49,60].

449 The AMOC dynamics and ice volume forcing are the key processes which link changes in
450 high-latitude temperature to low-latitude monsoon intensity [8,61]. The $^{231}\text{Pa}/^{230}\text{Th}$ records
451 from a highly resolved marine sedimentary record in the deep western North Atlantic show that
452 only those Heinrich events that occurred close to glacial maxima coinciding with massive
453 expansions of marine-based and land-based ice sheets (H1 and H2) were able to trigger the
454 off mode of AMOC [51,62], and deeply affect the Asian monsoon intensity.

455 Investigations of marine d^{18}O records derived from planktonic foraminifera and (SST)
456 estimates in the tropical Indian Ocean [49] also suggest a close relationship between the warm
457 basins of the Indo-Pacific region and high-latitude ice sheet changes linked by the global
458 thermohaline circulation (Fig. 4). A simulated Northern Hemisphere terrestrial climate dataset
459 for the past 60 ka produced a result similar to our data, that the inclusion of abrupt events
460 represents millennial-scale monsoon variability driven by fluctuations in the strength of the

461 AMOC (Fig. 5) [11]. North Atlantic water-hosing experiments suggest a dynamic response of
462 the monsoonal wind and precipitation in East Asia to a reduction of the AMOC, and the northern
463 westerlies may play a role in transmitting the signal from the North Atlantic to the Asian
464 monsoon regions [63].

465

466 **Fig. 5.** Model-proxy comparisons based on transient simulations over the AMD and reconstructed MAP from four pollen
467 records. Spatial pattern of climate simulation (a), and the corresponding temporal variation of Empirical Orthogonal
468 Function (EOF) analysis of simulated MAP over the AMD under different forcings (ORB (red), ORB + GHG (orange),
469 ORB + GHG + ICE (light blue) (b). The blue curve on (b) is the July 21 30°N–30°S insolation [48]. (c) Stacked curves
470 of quantitative precipitation reconstruction from the records of this study (TCQH, CH, and DJH) and GH [12].
471

472 In the Holocene, the heterogeneous spatiotemporal patterns of rainfall as inferred from
473 previous proxy-based reconstructions and modeling for China indicate that the mechanisms
474 behind the EASM remain poorly understood [40]. Our quantification data suggest that the
475 maximum rainfall amount occurred at slightly different periods of the Holocene in the three
476 locations, but no earlier than 8–9 ka ago, which was behind the peak of summer insolation and
477 the result derived from the transient model. Such a result is broadly consistent with the
478 synthesized Chinese Holocene stalagmite d18O records showing that the maximum strength
479 and stability of the EASM occurred during the middle Holocene at 6–8 ka BP [3], in accordance
480 with the classical view of a middle Holocene Optimum in China [10]. During the late Holocene,
481 the quantified precipitations from around 3 ka BP on a centennial-millennial scale show some
482 sharp fluctuations and a slight declining trend. Analysis of dominant pollen taxa in the late
483 Holocene shows that the key pollen taxa (e.g., Poaceae, Pinus, Alnus, etc.) differ from those in
484 the LGM (dominated by deciduous trees or Tibetan Sclerophyllous Quercus and alpine conifers).
485 The increase in grassland and Alnus or Pinus forest may be related to the expansion of
486 secondary forest due to more frequent fires [64]. It implies that pollen-based precipitation
487 reconstruction of the late Holocene may be influenced by anthropogenic disturbance, leading
488 to significant errors within reconstruction and bias against simulation results.

489 **CONCLUSION**

490 The pollen-based quantitative climate reconstruction of the last 68 ka highlights a significant
491 climate variability in southern China, which is characterized by a 35%–51% reduction in MAP,
492 and 5–7 °C lowering in MAT during the last glacial period. Quantification of temperature
493 changes suggests a steeper altitudinal lapse rate in the low-latitude alpine regions during the
494 LGM, which is consistent with climate simulation in the marginal mountains of the Tibetan
495 Plateau. The quantitative precipitation reconstruction broadly agrees with the d18O records
496 from most stalagmites in East Asia.

497 On glacial-interglacial timescales, the records from the caves in SW China and near the Bay
498 of Bengal are more consistent with our data than those from East China. This is in accordance
499 with the studies on synthetic Chinese d18O records and modeling results, showing that glacial-
500 interglacial changes in monsoon rainfall are manifested in isotopically different ways between
501 SW and East China. In particular, the best matching of our data with the CBoB records may
502 indicate that the location of the caves closer to the monsoon moisture source demonstrate
503 larger glacial-interglacial variability. In addition, shelf exposure influence on tropical Indo-Pacific
504 hydroclimate can largely affect regional ocean and atmosphere circulations, and thus amplify
505 the glacial-interglacial cyclicity. By comparing with climate forcings and transient climate
506 simulations, we found that the Asian monsoon precipitation was mainly driven by the insolation

507 differential between interhemispheric tropics on the orbital-scale, but its variation during the
508 transition from the LGM to the mid-Holocene was largely modulated by the abrupt changes
509 caused by the input of fresh water to the Atlantic and Southern Oceans during events such as
510 H1 and YD. We suggest that composite interactions between external (solar radiation) and
511 internal forcings (e.g., shelf exposure, SST and so on) significantly affected the variability of the
512 Asian summer monsoon. Our evidence also supports the hypothesis that there was a spatial
513 antiphase of precipitation between the records from SW and central China during the H1 and
514 YD events, indicating regional heterogeneity in the Asian monsoon systems. The present study
515 provides a robust standard pattern for precipitation variations in the humid regions of the Asian
516 summer monsoon, which can be used to validate the unquantified paleoclimate data. This study
517 provides a useful analog for the behavior of the Asian summer monsoon in the coming centuries.

518

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638