

1 **Moisture-advection feedback supports strong early-to-mid Holocene monsoon**
2 **climate on the eastern Tibetan Plateau as inferred from a pollen-based**
3 **reconstruction**

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26 **Highlights**

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- 28 • pollen record from a temperature- and precipitation-sensitive site
- 29 • abrupt 3°C temperature increase/decrease at 10.4/5.0 cal kyr
- 30 • warm/wet early/mid Holocene related to moisture-advection feedback

31

32 **Abstract**

33 (Paleo-)climatologists are challenged to identify mechanisms that cause the
34 observed abrupt Holocene monsoon events despite the fact that monsoonal
35 circulation is assumed to be driven by gradual insolation changes. Here we provide
36 proxy and model evidence to show that moisture-advection feedback can lead to a
37 non-linear relationship between sea-surface and continental temperatures and
38 monsoonal precipitation. A pollen record from Lake Ximencuo (Nianbaoyeze
39 Mountains) indicates that vegetation from the eastern margin of the Tibetan Plateau
40 was characterized by alpine deserts and glacial flora after the Last Glacial Maximum
41 (LGM) (21–15.5 cal kyr BP), by alpine meadows during the Late Glacial (15.5–10.4
42 cal kyr BP) and second half of the Holocene (5.0 cal kyr BP to present) and by mixed
43 forests during the first half of the Holocene (10.4–5.0 cal kyr BP). The application of
44 pollen-based transfer functions yields an abrupt temperature increase at 10.4 cal kyr
45 BP and a decrease at 5.0 cal kyr BP of about 3°C. By applying endmember modeling
46 to grain-size data from the same sediment core we infer that frequent fluvial events
47 (probably originating from high-magnitude precipitation events) were more common
48 in the early and mid Holocene. We assign the inferred exceptional strong monsoonal
49 circulation to the initiation of moisture-advection feedback, a result supported by a
50 simple model that reproduces this feedback pattern over the same time period.

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53 **Keywords**

54 moisture-advection feedback, monsoon, Tibetan Plateau, Holocene, Last Glacial
55 Maximum, pollen-climate calibration

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58 **1. Introduction**

59 Precipitation and temperature changes in monsoonal Asia are assumed to generally
60 portray insolation changes on orbital time-scales (Prell & Kutzbach, 1992; Wang et
61 al., 2008). With respect to the last deglacial and Holocene, three proxy-based
62 findings contradict the assumption of a strict linear relationship between monsoonal
63 strength and insolation. First, though proxy-record amalgamation curves (Herzschuh,
64 2006; Wang et al., 2010) indicate that the postglacial moisture rise depicts the
65 general shape of the insolation curve, the moisture optimum lags the insolation
66 optimum by several thousand years. Second, various studies indicate that
67 temperature and moisture changes since the last glacial may not be in parallel. For
68 example, a pollen record from the south-eastern Tibetan Plateau indicates that
69 glacial deserts were replaced by wet high-cold meadows from about 15 cal kyr BP
70 while thermophile forests only started to expand at the beginning of the Holocene
71 (Kramer et al., 2010a). Third, the gradual transition of insolation is not suitable to
72 explain the observed abrupt climatic changes recorded, for example, on the Tibetan
73 Plateau (Mischke & Zhang, 2010).

74 Moisture change in monsoonal Asia during the Late Glacial has been found to
75 correlate well with temperature change in the northern high-latitudes (Wang et al.,
76 2001; Herzschuh, 2006) where the post-LGM warming trend was reversed by cold

77 events originating from the fluctuation in the strength of the Atlantic Meridional
78 Overturning Circulation (AMOC) (Shakun & Carlson, 2010). In contrast, low-latitude
79 marine records show these cold reversals with much weaker amplitude and
80 abruptness (Shakun & Carlson, 2010). Two alternative hypotheses that could explain
81 the features of the AMOC events are preserved in remote Asian monsoon areas.
82 First, AMOC events directly drive monsoonal circulation via hemispheric-wide
83 teleconnection patterns such as the transition of the westerlies (Sun et al., 2012).
84 Second, the temperature event in the North Atlantic region has, due to its
85 remoteness, only minimal direct temperature effects but triggers a feedback
86 mechanism in monsoonal Asia which in turn amplifies temperature and precipitation.
87 The self-amplifying moisture-advection feedback introduced by Zickfeld et al. (2005),
88 elaborated by Levermann et al. (2009) and applied to past monsoon transitions by
89 Schewe et al. (2012) may be a candidate to cause non-linear behavior with respect
90 to changes in the system's energy budget. It is based on the idea that, regardless of
91 the many other complex aspects of monsoon circulation, a fundamental dynamical
92 feature, the so-called moisture-advection feedback, is required for monsoon rainfall
93 to extend substantially into the continental interior throughout the summer season.
94 An atmospheric temperature gradient between a relatively cool ocean area and a
95 relatively warm continental atmosphere is necessary to drive the monsoon winds. In
96 spring, at the onset of the summer monsoon, this gradient is established due to the
97 differential heating of the land and ocean surfaces. However, during the monsoon
98 rains, the land surface cools down again, and the atmospheric temperature gradient
99 cannot be maintained through sensible heating from the ground, but only through
100 latent heating from precipitation. The monsoon winds therefore carry their own "fuel"
101 in the form of moisture which condenses over land and releases latent heat,
102 reinforcing the temperature gradient and thus the monsoon winds. This self-
103 amplifying feedback leads to a non-linear response of the monsoon strength to
104 changes in the parameters that govern the availability of energy to the circulation.

105 One such parameter is the near-surface humidity over the ocean area that acts as
106 the moisture source for the monsoon circulation. When it falls short of a critical value
107 the moisture supply (and thus the latent heating) becomes insufficient to maintain a
108 deep circulation, and an abrupt transition can occur from a conventional monsoon
109 climate to a much drier summer climate.

110 Suitable records that provide quantitative estimates of both temperature and
111 precipitation changes from the Asian interior, necessary to understand climate
112 mechanisms better, are rare but are needed to avoid problems that arise from the
113 correlation of independent chronologies. Here we present quantitative climate
114 estimates from the application of pollen-climate transfer functions to a record from
115 Lake Ximencuo on the eastern Tibetan Plateau. We consider this site to be sensitive
116 to both temperature and precipitation change due to its location at the thermally
117 dependent treeline and its intermediate position along the southeast–northwest
118 moisture gradient on the eastern Tibetan Plateau. To minimize taphonomic effects on
119 the reconstruction we applied a transfer function that was solely constructed on
120 modern pollen records from lake surface-sediments. As pollen records reflect plant-
121 relevant climate variables (a combination of air humidity, precipitation, evaporation,
122 growing season length, etc.) rather than monsoon rain strength and air temperature,
123 we verify the mean annual precipitation reconstruction with a fluvial input proxy and
124 the mean annual temperature reconstruction with a temperature-dependent
125 weathering proxy from the same core.

126 Specifically, we want (1) to extract the regional vegetation changes on the eastern
127 Tibetan Plateau since the last glacial maximum, (2) to reconstruct mean annual
128 temperature, annual precipitation and monsoon-rain events and (3) to examine the
129 validity of existing conceptual models concerning non-linear monsoon behavior by a
130 proxy-data set.

131

132

133 **2. Regional setting**

134 Lake Ximencuo is located in the Nianbaoyeze Mountains on the northeastern Tibetan
135 Plateau in the administrative districts of Jiuzhi county (Qinghai Province) and Aba
136 Tibetan Autonomous Prefecture (Sichuan Province) (Fig. 1). The mountains mainly
137 consist of a granitic dome (Lehmkuhl & Liu, 1994) that rises from the surrounding
138 plateau surface of ~4000 m a.s.l. to peaks that exceed 5000 m a.s.l. Only the main
139 peak (Nianbaoyeze, 5369 m a.s.l.) is presently covered by a small ice cap. The
140 distribution of lateral and terminal moraines in the valleys radiate from the main
141 summit. They have been found to indicate several Pleistocene glaciation phases and
142 Holocene glacier highstands (Lehmkuhl & Liu, 1994; Lehmkuhl, 1995; Owen, 2003).

143 The climate on the eastern Tibetan Plateau is dominated by the Asian monsoon
144 circulation characterized by warm and wet summers and cold and dry winters. About
145 80% of precipitation falls between May and October. Lehmkuhl (1995) collected
146 information from regional climate stations in Jiuzhi (located ~40 km northeast of Lake
147 Ximencuo at 3629 m a.s.l.) and Aba (~60 km southeast at 3277 m a.s.l.). Mean
148 annual temperature at these stations is 0.1°C and 3.3°C, and annual precipitation
149 reaches 765 mm and 711 mm, respectively. According to the station-based climate
150 model of Böhner et al. (2006), mean annual temperature at 4000 m a.s.l. is -1.6°C
151 and annual precipitation in the Nianbaoyeze Mountains decreases from south (~1000
152 mm/year) to north (~600 mm/year) and increases with elevation.

153 No detailed investigation of the vegetation in the Nianbaoyeze Mountains is available.
154 Information on vegetation composition and distribution on the eastern margin of the
155 Tibetan Plateau is given in Huang (1987), Hou (2001), Kürschner et al. (2005) and
156 Wang et al., (2006). The Nianbaoyeze Mountains are situated at the northwestern
157 and upper limit of coniferous forests of the eastern Tibetan Plateau. The nearest
158 closed montane forests, mostly composed of *Picea* (*P. purpurea*, *P. likiangensis* var.

159 *balfouriana*, *P. asperata*) and *Abies* (e.g. *A. faxoniana*), intermixed with a few
160 broadleaf elements such as *Betula*, are distributed to the south and east of the
161 Nianbaoyeze Mountains at elevations between 2500 and ~4000 m, about 40 km
162 away from Lake Ximencuo. In the Nianbaoyeze Mountains themselves, only very few
163 trees are presently found. A few relict stands of *Picea purpurea* occur on the wet
164 northern slopes in the Lake Magencuo valley up to 4250 m (Lehmkuhl, 1995), about
165 10 km away from Lake Ximencuo. Some tree-like and shrubby *Juniperus* are found
166 on southern slopes. Shrubby vegetation composed of *Artemisia*, *Sibiraea*, *Spiraea*,
167 *Lonicera*, *Salix*, *Rhododendron*, *Potentilla* and *Caragana* grow well on shady
168 northern slopes up to 4400 m. High-alpine meadows are widely distributed covering a
169 wide elevation range from 3500 to 4500 m a.s.l., and are mainly composed of
170 *Kobresia*, *Polygonum* and Poaceae. The diversity of alpine meadows on the eastern
171 Tibetan Plateau decreases with increasing elevation (Wang et al., 2006) which often
172 leads to a strong dominance of *Kobresia pygmaea* at higher elevations. Sparse
173 alpine vegetation covers the mountain debris slopes composed of several species of
174 the genera *Saussurea*, *Saxifraga*, *Aconitum*, *Androsace*, and of the families
175 Caryophyllaceae, Fabaceae and Cyperaceae.

176 The valleys in the Nianbaoyeze Mountains are traditional settlement areas of Tibetan
177 nomads (Ryavec, 1999). Herders have practiced a 'semi-nomadic' lifestyle since the
178 mid-1990s: from May to September they live in traditional Tibetan black tents in the
179 mountains close to the summer pastures (up to 4500 m a.s.l.). From autumn to
180 spring, they live in permanent houses in the valleys (around 3700 m a.s.l.) with
181 fenced winter pastures. The herds in this region are mainly composed of yak (63%),
182 sheep (31%) and some horses (6%) (Miller, 1999).

183 Lake Ximencuo lies at the northern margin of the Nianbaoyeze Mountains at 4030 m
184 a.s.l. (Fig. 1). It has an open water area of 3.6 km² and a small catchment area of ~50
185 km² (Zhang & Mischke, 2009). The lake fills a deep (maximum lake depth: 63 m),

186 glacially-eroded basin surrounded by prominent moraines to the north. The moraines
187 in the immediate vicinity of the northern half of Lake Ximencuo have been dated to
188 20.2 ka BP whereas granitic boulders near the southern margin of the lake were
189 exposed by 16.5 ka BP (Owen et al., 2003), corresponding with a multiproxy-based
190 interpretation of the lake record (Zhang & Mischke, 2009).

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193 **3. Material and methods**

194 **3.1 *Sediment core and chronology***

195 In February 2004, a core of 12.81 m length was obtained from the center of Lake
196 Ximencuo (33.3792°N/101.1035°E; 4000 m a.s.l.) at a water depth of 56.2 m using
197 an UWITEC piston corer. Sediment properties including grain-size analysis, X-ray
198 fluorescence spectroscopy, magnetic susceptibility and total organic and inorganic
199 carbon yielded three main sediment units (Zhang & Mischke, 2009). The lower part of
200 the core (1281–710 cm) is characterized by high minerogenic content. Intermediate
201 minerogenic and organic contents characterize the middle part of the core (710–350
202 cm), and the upper part of the core (350–0 cm) is characterized by high organic
203 content particularly between 350–171 cm.

204 Radiocarbon dating of 18 stratigraphic levels was performed on bulk organic matter
205 due to the lack of any macroscopic remains. The alkali soluble and insoluble fractions
206 were dated separately, though not for all horizons for the former. Within dating errors,
207 the alkali insoluble and alkali soluble fraction yielded similar results for the Holocene,
208 however, the ages of the alkali soluble fraction were significantly older than the ages
209 of the insoluble fraction for the period before, a result also known from other sites of
210 comparable lake sediments from the Tibetan Plateau (Kramer et al., 2010a). The
211 radiocarbon age of 869 ¹⁴C years BP for the core-top sample was regarded as a 'lake

212 reservoir effect' (see reasoning in Zhang & Mischke, 2009) and subtracted from all
213 other ¹⁴C ages. Relevant information on all radiocarbon dates are presented in Zhang
214 & Mischke (2009) and Mischke & Zhang (2010). For this study the age-depth model
215 (Fig. 2) was newly constructed using the Bacon package (Blaauw & Christen, 2011)
216 in R software (R Core Team, 2012) including all dated horizons (using the alkali
217 soluble fraction for the lower part of the core and the alkali insoluble fraction for the
218 upper part). We *a priori* assumed an accumulation shape of 2, a mean accumulation
219 rate of 100 yr/cm, memory strength of 20 and step-rate of 20 cm. Due to the obvious
220 glacier-related high sediment flux, a sedimentation rate of 10 yr/cm was assumed for
221 the core part below 450 cm.

222

223 **3.2 Pollen data**

224 Pollen analyses were performed on 109 core samples. Processing of pollen samples
225 (2 ml of sediment) in the laboratory included treatments with HCl (10%), KOH (10%)
226 and HF (50%; 2 h boiling), followed by acetolysis, sieving (7 mm) in an ultrasonic
227 bath, and mounting in glycerine. At least 300 terrestrial pollen grains were counted,
228 except for some samples accumulated prior to 15 cal kyr BP when pollen content
229 relative to the non-soluble minerogenic remains was extremely low. Pollen
230 identifications are based on the relevant literature (Moore et al., 1991; Wang et al.,
231 1997; Beug, 2004) and on a type slide collection of more than 600 plant species from
232 western China. The total pollen sum of arboreal and terrestrial non-arboreal taxa
233 identified in each pollen spectrum is taken as 100% for the calculation of the pollen
234 percentages. Non-pollen palynomorphs were counted from the pollen slides and their
235 abundances are expressed relative to the terrestrial pollen sum. Identification of non-
236 pollen palynomorphs is based on Kramer et al. (2010b). Only pollen taxa which occur
237 in at least five core samples are included in the numerical analyses and in the
238 construction of the pollen diagram.

239

240 **3.3 Pollen-based ordination, clustering and climate reconstruction**

241 Pollen zonation was based on the results of depth constrained clustering (Birks &
242 Gordon, 1985) using squared chord distance as a dissimilarity measure and the
243 broken-stick model for the identification of the zone number (Bennett, 1996)
244 implemented in the rioja package (Juggins, 2009) for R. Pollen percentages were
245 square-root transformed prior to further computations. Ordinations were performed to
246 identify the major groupings in pollen samples and to portray relationships among
247 pollen taxa. A preliminary detrended correspondence analysis yielded a gradient
248 length of 1.5 standard deviation units, indicating that numerical methods based on
249 linear response models are appropriate with our dataset. Accordingly, a principal
250 component analysis (PCA) was performed (setup: focusing on species correlation,
251 centering of species data). Ordinations were implemented using CANOCO software,
252 version 4.5 (ter Braak & Šmilauer, 2002).

253 To quantify annual precipitation (P_{ann}) and annual temperature (T_{ann}), we applied a
254 pollen-precipitation transfer function (based on 55 pollen taxa) to the fossil pollen
255 spectra. For that purpose the original lake sediment-based transfer function
256 presented in Herzschuh et al. (2010a) was extended by the lake surface-sediment
257 pollen spectra from the southeastern Tibetan Plateau presented in Kramer et al.
258 (2010c) and from the Qaidam Basin presented in Zhao & Herzschuh (2008).
259 Samples from sites below 3000 m a.s.l. on the southeastern plateau were excluded
260 from the dataset due to intensive anthropogenic disturbance of the vegetation. Finally,
261 our modern pollen-climate calibration dataset is based on 132 lake surface-
262 sediments from the southern, eastern and northeastern Tibetan Plateau covering a
263 wide range for P_{ann} (31–1780 mm) and T_{ann} (-6.52–5.00°C). Pollen-climate transfer
264 functions were derived using weighted-averaging partial least squares (WA-PLS) (ter
265 Braak & Juggins, 1993). Model performance was assessed by leave-one-out cross-

266 validation (Birks, 1998). A two-component WA-PLS model for P_{ann} and a three-
267 component WA-PLS model for T_{ann} were chosen to be the most parsimonious models
268 on the basis of model performance statistics (Birks, 1998). Root mean square error of
269 prediction (RMSEP) is 108 mm for P_{ann} and 1.56°C for T_{ann} , and the coefficient of
270 determination (r^2) is 0.92 and 0.64 between observed and model-predicted values.
271 RMSEP, when expressed as a percentage of the climate variable gradient, is 6.2%
272 and 13.5% for P_{ann} and T_{ann} , respectively, illustrating the good performance of the
273 models. The tree pollen content of the lower part of the core (before 15 cal kyr BP) is
274 interpreted as originating from far distances (similar to near-by Lake Nalengcuo;
275 Kramer et al, 2010a), which reaches high values at times of low regional pollen
276 productivity. For that reason, models without tree taxa were applied to these samples
277 (P_{ann} -no tree model 1st component: RMSEP of 133 mm and r^2 of 0.87; T_{ann} -no tree
278 model 2nd component: RMSEP of 1.61°C and r^2 of 0.63).

279

280 **3.4 Grain-size data and endmember modeling**

281 Grain-size analysis was performed at 8-cm intervals between the core base and 6.1
282 m core depth, at 4-cm intervals between 6.1 and 4.1 m, and at 2-cm intervals above
283 4.1 m. A Malvern Mastersizer 2000 laser granulometer was used (size range: 0.02–
284 2000 μm). Sample pretreatment included (1) adding H_2O_2 to remove organic matter
285 and soluble salts, (2) using diluted 1 M HCl to remove carbonate, and (3) using Na-
286 hexametaphosphate to disperse aggregates. General parameters obtained from
287 grain-size analyses (skewness, mean, kurtosis, sorting, grain-size fraction for sand,
288 silt and clay) are presented in Zhang & Mischke (2009). Here we present the
289 reanalyses of these grain-size data using the End-Member Modeling Algorithm
290 (EMMA) developed by Weltje (1997). This was applied in order to identify robust end-
291 members (EMs) and estimate their proportional contribution to the sediments. To
292 avoid a fixed single outcome and extract reliable robust EM(s), models were run

293 considering different numbers of end-members (between the minimum and maximum
294 numbers of potential EMs) and flexible weight transformations. The minimum number
295 of potential EMs was determined by the cumulative explained variance reaching at
296 least 95%; the maximum number of EMs was determined by the maximum value of
297 the mean coefficient of determination. We tested the robustness of the EMs and then
298 extracted the final robust EM(s) and residual member(s). All these computations
299 were made using MATLAB software. A detailed description of the EMMA method
300 applied can be found in Dietze et al. (2012).

301

302

303 **3.5 Modeling approach**

304 We construct a simple conceptual model of the summer monsoon circulation
305 (Levermann et al., 2009; Schewe et al., 2012) in which a characteristic relation
306 between monsoon precipitation and humidity over the ocean follows from the self-
307 amplifying moisture-advection feedback: below the critical threshold precipitation is
308 zero; above the threshold, precipitation increases with humidity over the ocean the
309 functional form becoming approximately linear at some distance from the threshold.
310 In order to obtain quantitative results, a number of parameters has to be known: the
311 slopes of the relations between (a) monsoon winds and atmospheric temperature
312 gradient and (b) precipitation and the terrestrial atmospheric humidity; and (c) the
313 threshold. Here, we merely intend to demonstrate that the timing and abruptness of
314 the transitions into and out of a strong monsoon phase in the first half of the
315 Holocene can be reproduced with this model. Quantitative rainfall estimates for the
316 study period would require more data and are beyond the scope of this study.
317 Therefore, we set parameters (a) and (b) to values estimated previously for the East
318 Asian region on the basis of reanalysis data (Schewe et al., 2012). The humidity
319 threshold is assumed to correspond to a sea-surface temperature (SST) threshold,

320 which is deliberately chosen to be at 26.35°C. The Mg/Ca-inferred SST time-series
321 reported by Rashid et al. (2011) for the Bay of Bengal, which is likely to have been a
322 major source of moisture for our study region, is then translated by the conceptual
323 model into a monsoon rainfall time-series.

324

325

326 **4. Results**

327 **4.1 Vegetation changes at Lake Ximencuo**

328 Overall, the non-arboreal taxa Cyperaceae, Poaceae, *Artemisia* and
329 Chenopodiaceae and the arboreal taxa *Betula*, *Picea* and *Pinus* are the most
330 abundant taxa in the fossil pollen spectra from Lake Ximencuo, summing up to >85%
331 in most samples (Fig. 3). The first two PCA axes capture 45.8% (axis 1: 30.2%, axis
332 2: 15.6%) of the total variance in the data. The correlation biplot of the first two axes
333 (Fig. 4) separates samples characterized by elements of sparse alpine vegetation
334 and alpine deserts at the positive end of the first axis and those with high values for
335 shrubs and meadow-elements at the negative end. The second axis separates
336 samples dominated by Cyperaceae and Poaceae from those rich in tree pollen and
337 *Artemisia*. In accordance with the results from depth-constrained cluster analyses
338 and from the broken-stick model the pollen diagram can be separated into five pollen
339 assemblage zones (PAZ).

340 PAZ 1 (12.74–6.30m; 20.9–15.3 cal kyr BP) is characterized by high values of
341 Chenopodiaceae, Caryophyllaceae, Brassicaceae, *Aster*-type, *Anthemis*-type,
342 *Saussurea*-type, Papaveraceae and *Hippophaë* as well as by high values of tree taxa
343 such as *Pinus*, *Picea* and *Betula* summing up to 10–20%. A decrease of tree taxa
344 and a rise of Cyperaceae to mostly >45% mark the transition to PAZ 2 (6.30–4.78m;
345 15.3–13.8 cal kyr BP). The transition to PAZ 3 represents the strongest turnover in

346 the pollen assemblage. In contrast to previous zones, PAZ 3 (4.78–3.64m; 13.8–10.4
347 cal kyr BP) has comparatively low percentages of desertic and glacial elements as
348 well as a low tree pollen fraction. It is characterized by meadow and shrubland taxa
349 such as Poaceae, *Thalictrum* and *Salix* and has in comparison to the other zones
350 intermediate values for Cyperaceae and *Artemisia*. The steep rise of tree taxa
351 particularly of *Betula* marks the transition to PAZ 4 (3.64–2.65m; 10.4–5.0 cal kyr BP)
352 which, aside from a high arboreal pollen portion, is characterized by the highest
353 *Artemisia* and lowest Cyperaceae values of the whole record. PAZ 5 (2.65–0 m; 5.0–
354 0 cal kyr BP) displays again higher values for typical meadow taxa such as Poaceae
355 and *Thalictrum*, lower *Artemisia* and tree values and higher Cyperaceae values than
356 in PAZ 4. While the samples of PAZ 1 and PAZ 3 are loosely scattered in the lower
357 right and upper central part of the PCA plot, PAZ 2 and particularly PAZ 4 and PAZ 5
358 form well-defined clouds in the upper right, lower left and upper left quadrant,
359 respectively.

360 *Glomus*, a fungal spore, is present in several samples from the lower part of the core
361 (before 14.5 cal kyr BP) and in a few late Holocene samples. *Botryococcus* and
362 *Pediastrum* are absent in PAZ 1, have low values in PAZ 2 and have high values in
363 PAZs 3, 4 and 5. The highest values are recorded after about 7.0 cal kyr BP, which
364 however, may originate from changes in the pollen influx rather than from changes in
365 algae productivity. Volvocaceae are present during the last 6.5 cal kyr and show
366 highest values during the last 2.4 cal kyr.

367

368 **4.2 Pollen-based quantitative climate reconstructions**

369 The total reconstructed range is between 316 and 883 mm for P_{ann} and 0.90
370 and -4.57°C for T_{ann} , which is covered by the modern climate range of pollen
371 analogues (Fig. 5). Transfer function-based reconstructions for the upper part of the
372 core of 759 mm for P_{ann} and -1.80°C for T_{ann} (mean of the upper two samples)

373 resemble well the climate station-based values of 791 mm and -1.62°C, indicating the
374 general applicability of the transfer function to this record.

375 Reconstructed mean annual temperature is mostly below -3°C during the Late
376 Glacial with slightly higher values between 14.7 and 12.6 cal kyr BP. A sharp
377 increase of about 3.5°C occurs during the earliest part of the Holocene and the
378 thermal optimum lasts until ~5 cal kyr BP when temperature suddenly drops by about
379 2.5° towards an overall Holocene minimum around 4 cal kyr BP.

380 Reconstructed annual precipitation is low for the lower part of the core, mostly less
381 than 500 mm, even below 400 mm for most samples from the period 17.8 and 15.1
382 cal kyr BP. Reconstructed precipitation is above average between 15.3 and 6 cal kyr
383 (mostly >800 mm), with strong increases occurring around 14 cal kyr BP and 10.5 cal
384 kyr BP. Around 4 cal kyr BP reconstructed precipitation shows a Holocene minimum.
385 It ranges around 700 mm for the late Holocene part of the core.

386

387 **4.3 Grain-size endmembers and their changes through time**

388 The grain-size spectra are dominated by the silt fractions (62.1–81.9%), with the clay
389 (13.4–25.8%) and sand (0–24.0%) fractions only accounting for a small part of the
390 total. We obtained three robust and one residual endmember when running the
391 models with different weight transformations (the percentile range of P₁-P₉₉, P₂-P₉₈,
392 P₃-P₉₇, P₄-P₉₆) and EMs from 5 and 6 endmembers (Fig. 6). The robust endmember
393 1 is dominated by a clay fraction peaking at 2 μm. It constitutes a high portion
394 throughout (Fig. 7) and it dominates during the lower part of the record between 21.0
395 and 15.3 cal yr BP. Robust endmember 2 has a bimodal frequency distribution
396 peaking at 16 μm and 125 μm. This endmember is almost absent from the lower part
397 of the record until 15.3 cal kyr BP, it sporadically appears between 15.3 and 13.8 cal
398 kyr BP and obtains its highest portion during the early-to-mid Holocene. Robust

399 endmember 3 is unimodal and dominated by silt-size grains with a peak at 62 μm ; it
400 dominates the spectrum between 15.3 and 6.0 cal kyr BP. The residual member is
401 high in the upper part of the core. Robustness analyses yielded a mean value of
402 $r^2=0.68\pm 0.22$ for grain-size scale and $r^2=0.86\pm 0.09$ for ages.

403 **4.4 Modeling results**

404 The translation of the SST record of Rashid et al. (2011) into monsoon strength
405 according to the simple conceptual model of the summer monsoon circulation
406 (Schewe et al., 2012) exhibits an abrupt increase around 11 cal kyr BP and an abrupt
407 decrease around 5 cal kyr BP (Fig. 5).

408

409

410 **5. Discussion**

411 **5.1 Vegetation changes at Lake Ximencuo in comparison to other Tibetan** 412 **records**

413 The pollen record from Lake Ximencuo indicates that a dry glacial flora dominated
414 the post-LGM vegetation which is in agreement with other records from the eastern
415 Tibetan Plateau (e.g. Herzsuh et al., 2009, Kramer et al., 2010a). However, in
416 comparison to other sites the amount of ruderal indicators and cushion plants such
417 as Papaveraceae, Chenopodiaceae, Brassicaceae, *Saussurea*-type and
418 Caryophyllaceae is particularly high at Lake Ximencuo. This indicates that fresh
419 glacier-free soils dominated in the direct vicinity of the lake which fits with the finding
420 that the glacier extended to the southern lake shore until about 16.5 ka (Owen et al.,
421 2003; Zhang & Mischke, 2009).

422 We did not observe differences in the pollen grain preservation that would indicate
423 the input of reworked pre-LGM material. Accordingly, we interpret the high amount of
424 *Pinus*, *Picea* and *Quercus* as an extra-regional signal. This is in agreement with

425 modern pollen deposition studies where sparse vegetation cover has been shown to
426 cause a high reflection of long-distance transported pollen grains (Herzschuh, 2007).
427 Our interpretation also fits with the finding that boreal forests survived at lower
428 elevations in the Yunnan-part of the Tibetan Plateau throughout the glacial as
429 revealed by high *Pinus* pollen values in Shudu Lake sediments (Cook et al., 2011).

430 Starting at about 15.5 cal kyr BP, the dry glacial flora is replaced by Cyperaceae-rich
431 high-alpine meadow (probably dominated by *Kobresia*, analogous to modern
432 vegetation) which turned into alpine meadow and shrubland roughly similar to the
433 modern vegetation by about 14 ka. The composition and timing of these Late Glacial
434 vegetation changes are in agreement with reconstructions based on the Naleng Lake
435 pollen record (Kramer et al., 2010a), a glacial lake that is likewise located at the
436 present-day treeline indicating that the Lake Ximencuo pollen record reflects typical
437 regional signals. The re-expansion of ruderal taxa, particularly of Chenopodiaceae,
438 between 12.5 and 11 cal kyr BP is, within dating errors, synchronous with the
439 Younger Dryas event.

440 Starting at about 11 cal kyr BP, *Betula*-rich *Picea/Pinus* forest became established in
441 the Nianbaoyeze Mountains and dominated the area until about 5.0 cal kyr BP
442 without major vegetation changes. The expansion of boreal forests beyond their
443 present-day western and northern limit during the early and mid Holocene is a
444 phenomenon seen in most records from the eastern Tibetan Plateau margin
445 (Herzschuh et al., 2010b; Zhao et al., 2011). In contrast to the Lake Naleng area
446 (Kramer et al., 2010c), where alpine meadows re-occur between 8 and 7 cal kyr BP,
447 only a minimal glacial flora re-expansion may have occurred at Lake Ximencuo as
448 suggested by a slight increase in the Papaveraceae record.

449 The mixture of alpine meadows and shrublands broadly similar to modern conditions
450 became established in the surroundings of Lake Ximencuo by about 4.8 cal kyr BP.
451 However, secondary vegetation (indicated by higher *Potentilla*-type and *Quercus*

452 values) reflecting human impact expanded from about 3 cal kyr BP onwards, similar
453 to findings from the Naleng Lake record (Kramer et al., 2010c).

454

455 **5.2 Pollen-based quantitative climate estimates and comparison to other** 456 **records**

457 Pollen-based climate reconstructions may be biased if human activities and CO₂
458 changes markedly affected the past vegetation composition. Hitherto archeological or
459 paleoecological evidence that humans—although present since at least the mid
460 Holocene (Altenderfer et al., 2006)—altered the vegetation beyond a local scale is
461 lacking. This is obvious when comparing a Holocene low-resolution peat pollen
462 record from the Ximencuo valley (Schlütz et al., 2009) to our pollen record from Lake
463 Ximencuo which reveals that both records differ particularly in the abundance and
464 trends of herbaceous plants. While local human impact is traceable by grazing
465 indicators at the site of the peat record since 6 cal kyr BP, on a regional scale the
466 abundance of these indicators during the Holocene does not vary in the lake pollen
467 record with its much larger pollen source area. Accordingly, we assign the mid-
468 Holocene decline of tree pollen largely to climate, as this feature is consistent with
469 records of forest decrease from all over Inner Asia (Zhao & Yu, 2012). This supports
470 the suitability of the record for quantitative climate reconstruction purposes.
471 Nevertheless, the pollen-based reconstruction may underestimate Late Glacial and
472 early Holocene precipitation as dryland vegetation is favored in times of low CO₂
473 particularly at high elevations (Herzschuh et al., 2011). Furthermore, long-term
474 processes such as soil development in the direct vicinity of the lake may have
475 caused vegetation-climate disequilibrium and thus reduce the quality of the climate
476 reconstruction, while time-lags related to small migration rates are probably irrelevant
477 in the mountainous terrain of eastern Tibet.

478 Quantitative climate estimates from Lake Ximencuo can be compared to other pollen-
479 based reconstructions from the Tibetan Plateau, although it needs be taken into
480 account that they have been derived using different reconstruction methods and/or
481 modern calibration sets based on various sedimentary origins.

482 Available reconstructions from the Tibetan Plateau (Koucha Lake, Herzschuh et al.,
483 2009; Luanhaizi, Herzschuh et al., 2010a; Kuhai Lake, Wischnewski et al., 2011;
484 Donggi Cona, Wang et al., 2013) show, like those from Lake Ximencuo, ~50% of the
485 present-day precipitation value for the pre-14 cal kyr BP Late Glacial period. In
486 particular, those records from the northeastern Tibetan Plateau (Herzschuh et al.,
487 2009, Wischnewski et al., 2011) indicate that a major increase in precipitation
488 occurred at about 15 cal kyr BP while at Lake Ximencuo a two-step increase towards
489 Holocene levels occurred (i.e. likewise about 15 cal kyr BP and at the beginning of
490 the Holocene). All available records from the southern or eastern humid plateau
491 margins (Tang et al., 1999; Lu et al., 2011, Herzschuh et al., 2010a) indicate that the
492 early-to-mid Holocene was the wettest period of the entire post-LGM phase. However,
493 absolute values and values relative to modern conditions vary markedly among these
494 sites. At all these sites (but not Chen Co) a slightly wetter mid Holocene than early
495 Holocene is recorded and all sites show a gradual reduction of reconstructed
496 precipitation from about 6 cal kyr BP. However, the Lake Ximencuo precipitation
497 pattern differs from those obtained from the arid upper northeastern Tibetan Plateau
498 where the early Holocene seems to be slightly drier than present-day conditions
499 (Herzschuh et al., 2009, Wischnewski et al., 2011) as suggested by a relatively high
500 *Artemisia* component. An early Holocene high *Artemisia* component is also visible in
501 the Lake Ximencuo record: it is the cause of the early lower than mid-Holocene
502 reconstructed precipitation. Whether this is a true reflection of the regional vegetation
503 around Lake Ximencuo lake or rather reflects the vegetation changes on the upper

504 northeastern Tibetan Plateau (and thus biases the quantitative precipitation
505 estimates) is impossible to discern.

506 Non-pollen quantitative temperature records of similar length and resolution to the
507 Lake Ximencuo record that could be used for evaluation of the obtained T_{ann}
508 reconstructions are lacking for the Tibetan Plateau. Some pollen-based
509 reconstructions such as those from the northeastern and central Tibetan Plateau
510 (Herzschuh et al., 2009, Lu et al., 2011) are questionable as they are not located at a
511 temperature sensitive position and the vegetation at these sites seems to be
512 controlled by moisture only. However, mean annual temperature reconstructions from
513 Lake Hidden (Tang et al., 1999) and Lake Luanhaizi (Herzschuh et al., 2006) that
514 originate from a similar ecoclimatic position to that of Lake Ximencuo may be useful
515 for evaluations. Both show generally similar trends and quantities of change. For
516 example, transfer function-based results from Lake Hidden show a $\sim 4^{\circ}\text{C}$ difference
517 between pre-15 cal kyr BP and the Holocene optimum and, at both sites, the
518 strongest increase of $>3^{\circ}\text{C}$ occurred at the beginning of the Holocene at about 10.0
519 cal kyr BP.

520 The reliability of the inferred temperature trends from Lake Ximencuo pollen record is
521 furthermore supported by its correlation to the Y/Al element ratios (Fig. 5), which
522 probably reflect the weathering (i.e. enrichment of Y a resistant mineral relative to Al)
523 in the catchment of Lake Ximencuo that is assumed to be high during warm phases
524 (Zhang & Mischke, 2009).

525

526 **5.3 Fluvial erosion and the tracing of heavy monsoon events**

527 Basin morphology and the lack of ancient lake shorelines indicate that Lake
528 Ximencuo most probably obtained a stable high lake level throughout its evolution
529 since the LGM. Accordingly, grain-size distribution may reflect changes in the

530 sediment input to the lake rather than distance changes between the coring position
531 and the shoreline or river mouth to the lake. Clay-dominated EM1 most probably
532 reflects the glacier input to the lake. EM3 shows a grain-size distribution typical for
533 Chinese loess (An et al., 2012) and hence may reflect aerial input. This interpretation
534 is in temporal accordance with other records from western China that find a high
535 aeolian input during the Younger Dryas (e.g. Guan et al., 2009; Liu et al., 2013). The
536 'residual' member might reflect a mixture of autochthonal productivity (e.g. of
537 diatoms) and of continuous sediment supply from streams. We interpret EM2 as
538 tracking erosion peaks related to heavy monsoon rain events that caused the
539 transport of relatively coarse-grained sediments to the lake center, accompanied by
540 in-washed sediments of smaller grain size even though other events such as
541 earthquakes may have produced a similar signal (Wang et al., 2014). This record is
542 in temporal accordance with an anoxygenic phototrophic bacteria record from
543 Qinghai Lake (Ji et al., 2009, Fig. 5) that likewise is assumed to trace freshwater
544 inputs in relation to heavy monsoon rain events. We therefore infer that heavy
545 monsoon rains were lacking before 15.3 ka, were infrequent between 15.3-10 ka BP,
546 peaked during the early and mid Holocene and decreased after 4 ka.

547

548 **5.4 The moisture-advection feedback as main climate driver**

549 The pollen-based quantitative climate reconstruction supports earlier findings that
550 postglacial monsoonal development on the Tibetan Plateau is not a simple reflection
551 of insolation changes. First, the postglacial temperature and precipitation optimum
552 phase occurred during the early and early mid-Holocene, which is several thousand
553 years after the insolation optimum as reported by several studies (Wang et al., 2011).
554 Second, the precipitation and temperature reconstructions are not entirely in parallel
555 at Lake Ximencuo, as would be expected for a monsoon climate directly forced by
556 insolation. While the major increase in precipitation occurred at ~15 cal kyr BP the

557 major temperature increase occurred at the beginning of Holocene. This supports
558 previous qualitative climate inferences from pollen records obtained from lakes at
559 similar bioclimatic positions on the Tibetan Plateau (e.g. Naleng Lake, Kramer et al.
560 2010a) and is in agreement with the finding that the major vegetation change in arid
561 areas of the Tibetan Plateau occurred at ~15 cal kyr BP rather than at the beginning
562 of the Holocene (Wischnewski et al., 2011). Third, vegetation change and thus
563 inferred climate change at Lake Ximencuo is characterized by rapid transitions (as
564 has been reported from other Tibetan records, Mischke & Zhang, 2010), which is in
565 contrast to the smooth changes of insolation. As inferred from the Lake Ximencuo
566 pollen record, temperature rapidly increases at the beginning of the Holocene and
567 decreases at about 5 cal kyr BP.

568 In our view the observed differences between climate change on the eastern Tibetan
569 Plateau and the insolation signal can be best explained by taking the moisture-
570 advection feedback into account when explaining postglacial climate development.
571 While the differential heating of land and ocean in spring is important for the initiation
572 of the monsoon, its continuation over the entire summer until September or October
573 is only possible with the release of latent heat associated with rainfall over the
574 continents, which reinforces the monsoon winds (Levermann et al., 2009). According
575 to a conceptual model of the monsoon season, this feedback leads to a non-linear
576 response of the monsoon strength to changes in boundary conditions. In particular, a
577 conventional large-scale monsoon circulation can only exist if atmospheric specific
578 humidity over the ocean (which is closely related to SST) is above a specific
579 threshold value (Schewe et al., 2012). Below the threshold, the advection and
580 release of latent heat are insufficient to sustain monsoon rainfall over land. We
581 hypothesize that this threshold was passed in most years between 10.4 and 5.0 cal
582 kyr BP and the moisture-advection feedback enabled a powerful South Asian
583 summer monsoon that extended far into the continental interior and led to marked

584 latent heat release and heavy precipitation events throughout the entire summer in
585 the study region. This hypothesis is consistent with the expansion of thermophile
586 plant taxa, the frequent occurrence of fluvial erosion events, and, compared to today,
587 an increased weathering rate in the catchment of Lake Ximencuo as inferred by this
588 study. In contrast, we argue that the moisture-advection feedback was not or only
589 rarely active during the Late Glacial and the second part of the Holocene, due to low
590 sea-surface temperatures and low humidity over the ocean. A threshold behavior
591 would also explain the observed ~1200-yr time-lag between the end of the Younger
592 Dryas in high northern latitude records (Hoek, 2008) and the major temperature
593 increase at Lake Ximencuo in the monsoonal realm (this study), although dating
594 uncertainty may also have contributed to the temporal mismatch. Our inference about
595 the importance of the moisture-advection feedback for Inner Asian climate is
596 supported by the results obtained from driving the conceptual model with Mg/Ca-
597 inferred SST from the Bay of Bengal (Rashid et al., 2011; Fig. 5). This experiment
598 shows that our hypothesis is consistent with the reconstructed changes in sea-
599 surface temperatures in the relevant ocean region, which gradually increased after
600 the LGM and reached a maximum between approximately 10 and 5 cal kyr BP. The
601 threshold behavior introduced by the self-amplifying moisture-advection feedback
602 and captured in the conceptual model leads to a rapid change in monsoon strength in
603 response to such gradual changes in boundary conditions.

604 At the same time, we found that total precipitation changes, as inferred from the
605 pollen record, were much less abrupt than temperature change, and moisture was
606 available in the eastern Tibetan Plateau both before and after this period of strong
607 monsoon. Here it is important to note that the temperature signal recorded in the
608 pollen abundance might directly reflect summer temperature, i.e. growing season
609 temperature, which is particularly relevant for vegetation, whereas, the precipitation
610 signal might integrate moisture availability over different times of the year and from

611 other sources than monsoon precipitation (e.g. continental moisture recycling and
612 supply from the westerlies). This difference could explain why abrupt changes in
613 monsoon strength may be better reflected in the temperature signal.

614

615

616 **6. Conclusion**

617 The pollen-based reconstruction of climate changes on the eastern margin of the
618 Tibetan Plateau since the end of the LGM reveals an abrupt and strong ($\sim 3^{\circ}\text{C}$)
619 temperature increase at about 10.4 cal kyr BP and a strong decrease at 5.0 cal kyr
620 BP. Reconstructed precipitation is highest during the first half of the Holocene,
621 although precipitation markedly increases already by ~ 15 cal kyr BP and changes are
622 generally more gradual than for temperature. The reconstruction of fluvial events
623 from the Ximencuo record, as well as proxy-evidence from Qinghai Lake, indicate
624 that the first half of the Holocene is not only characterized by exceptionally warm
625 conditions but also by the frequent occurrence of strong precipitation events. The
626 application of a simple conceptual model depicting the inland moisture transport
627 demonstrates that a moisture-advection feedback is initiated during this time period,
628 which may have led to the observed non-linear relationship between SST in the
629 nearby sea areas and warmth and monsoonal precipitation in the Asian interior. Our
630 study provides proxy evidence that the moisture-advection feedback is capable of
631 both translating gradual insolation changes into a stepwise monsoon activity pattern
632 and of amplifying minimal changes in tropical SST such as those originating from
633 remote climatic events (e.g. the AMOC shut down) into a monsoon signal of large
634 amplitude on nearby continents.

635

636

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817 **Figures and figure captions**

818 **Fig.1** Study area of the Nianbaoyeze Mountains, showing the location of Lake
819 Ximencuo and sites mentioned in the text. Detailed map after Lehmkuhl (1995) and
820 Wischnewski et al. (2013).

821 **Fig. 2** Recalculation of the age-depth model for Lake Ximencuo using the bacon
822 procedure of Blaauw & Christen (2011) on ^{14}C dating results published in Zhang &
823 Mischke (2009) and Mischke & Zhang (2010).

824 **Fig. 3** Pollen diagram from Lake Ximencuo sediment core. Unshaded polygons trace
825 the 10x magnifications.

826 **Fig. 4** Biplot of axes 1 and 2 of a Principal Component Analyses performed on the
827 pollen data from Lake Ximencuo.

828 **Fig. 5** Climate change (pollen-based T_{ann} and P_{ann} reconstructions from Lake
829 Ximencuo, this study; Y/Al element ratio from Lake Ximencuo, Zhang & Mischke,
830 2009) and the occurrence of heavy monsoon events on the eastern Tibetan Plateau
831 (grain-size EM1, this study; bacteria blooms in Qinghai Lake, Ji et al., 2009)
832 compared to the inland monsoonal precipitation estimates (arbitrary scale, this study)
833 from driving a minimal conceptual model on the moisture-advection feedback
834 (Levermann et al., 2009, Schewe et al., 2012) by sea-surface temperatures from the
835 Bay of Bengal (Rashid et al. 2011). Exceptionally high temperatures at Lake
836 Ximencuo and the indication of high-magnitude monsoon events coincide with—and
837 are probably causally related to—the inferred initiation of the moisture-advection
838 feedback in monsoonal Asia.

839 **Fig. 6** Loadings of grain-size based endmember modeling of Lake Ximencuo
840 sediment core. The extracted three endmembers are assumed to represent unmixed
841 grain-size distributions contributed from different sources to the lake basin i.e. EM1 –

842 glaciofluvial input, EM2 – fluvial erosion during strong precipitation events, EM 3 –
843 aerial input, residual member – mixed-sources.

844 **Fig. 7** Variation in the contribution of the three endmembers and the residual
845 member to the lake's minerogenic content throughout the sedimentation of the Lake
846 Ximencuo record. According to our interpretation, EM1 represents the glaciofluvial
847 input, EM2 the fluvial erosion during strong precipitation events, EM 3 the aerial input
848 and the residual member originates from mixed-sources.

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