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1 **Proto-monsoon rainfall and greening in Central Asia due to extreme early**
2 **Eocene warmth**

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31

32 **The dry continental interior of Asia has remained arid throughout most of its**
33 **geological history, yet the future of this unique ecosystem remains unclear.**
34 **Here, we use palynological and isotopic records to track vegetation and**
35 **moisture throughout the warm early Eocene (57 to 44 million years ago) as an**
36 **analogue for extreme atmospheric CO₂ scenarios. We show that rainfall**
37 **temporarily doubled and replaced the regional steppe by forested ecosystems.**
38 **By reconstructing the season of pedogenic carbonate growth we constrain the**
39 **soil hydrologic regime and show that most of this rainfall occurred during the**
40 **summer season. This humid event is therefore attributed to an inland expansion**
41 **of monsoonal moisture following the massive greenhouse gas release of the**
42 **Paleocene-Eocene Thermal Maximum (PETM) as identified by a negative carbon**
43 **isotope excursion. The resulting abrupt greening of the Central Asian steppe-**
44 **desert would have enabled mammal dispersal and could have played a role in**
45 **carbon cycle feedbacks by enhancing soil organic carbon burial and silicate**
46 **weathering. These extreme Eocene proto-monsoons, albeit different from the**
47 **topography-driven Asian monsoon today, highlight the potential for abrupt**
48 **shifts in Central Asian rainfall and ecosystems under future global warming.**

49

50 The vast Asian continental interior is home to a unique and diverse steppe-desert
51 biome that has existed for millions of years¹, yet its hydroclimatic future and response
52 to extreme atmospheric CO₂ concentrations (pCO₂) remains poorly understood².
53 Sedimentary archives of past warming events together with climate model simulations
54 provide valuable insights³, especially regarding the Eocene period (56-34 Ma) which
55 was characterized by short-lived hyperthermal events of abrupt warming^{4,5}. The
56 largest and most pronounced of these is the Paleocene-Eocene Thermal Maximum
57 (PETM; 56 Ma) lasting several 10 ka, during which pCO₂ reached more than 6× the
58 pre-industrial atmospheric CO₂ concentration⁶⁻⁸. Records of such hyperthermal events
59 affecting inland Asia are rare and hard to decipher because we commonly lack
60 unambiguous rainfall proxies with respect to the various regional moisture sources,
61 thereby limiting our understanding of the hydroclimatic response to climate forcing,
62 including future projections under higher pCO₂³.

63

64 The scarce precipitation reaching inland Asia today mostly originates from monsoons
65 during summer or the mid-latitude westerlies during winter to spring (Fig. 1 A;
66 Extended Data Fig. 1)^{9,10}. Paleoclimate models and geological records indicate that
67 growth of the Tibetan-Himalayan orogen and retreat of the Paratethys Sea modulated
68 monsoons and westerlies since the Eocene (Fig. 1 A)¹¹⁻¹⁴. Proto-monsoons existed
69 already before significant surface uplift of the Tibetan Plateau, but it remains debated
70 whether these were similar to today¹⁵⁻¹⁷, less intense¹⁸ or only restricted to the tropical
71 low latitudes^{13,14,19} except during extreme orbital conditions^{20,21}. Eocene
72 paleobotanical data show that summer monsoons transported abundant precipitation
73 to the East Asian coast^{17,22}, while regional soil respiration derived from carbon isotope
74 ratios ($\delta^{13}\text{C}$) of pedogenic carbonates suggest that monsoons may have reached

75 further inland¹⁰. Oxygen isotope ratios ($\delta^{18}\text{O}$) on the other hand, suggest that the mid-
76 latitude westerlies provided winter to springtime precipitation to Central Asia,
77 especially before the Neogene surface uplift of the Tien Shan and Pamir mountains,
78 which block most of this moisture today^{9,10,23}. However, the isotopic composition of
79 this westerly precipitation is difficult to distinguish from summer monsoonal moisture
80 reaching the continental interior¹⁰.

81

82 Key in distinguishing between these different Asian moisture regimes is therefore to
83 identify not only the isotopic composition and amount of precipitation, but also the
84 seasonal timing, which remains difficult to assess from geological records. Here, we
85 address this issue using a novel approach of combining fossil sporomorphs (pollen
86 and spores) and soil organic matter as proxy for precipitation²⁴ together with clumped
87 isotope (Δ_{47}) temperatures of co-occurring pedogenic carbonates which mostly form
88 during the dry season^{25,26}. By comparing these Δ_{47} -derived soil temperatures with
89 seasonal climate model estimates we distinguish whether soil dewatering occurred
90 predominantly during a winter-time dry season after a wet summer monsoon, or during
91 summer drying of westerly-derived soil moisture^{27,28}. Today, the monsoons and
92 westerlies meet along the northeastern margin of the Tibetan Plateau, which includes
93 the continental Xining Basin (China) and is ideally located to track the evolution of
94 these moisture sources (Fig. 1). Previous soil temperature reconstructions throughout
95 the middle Eocene and Oligocene show that the Xining Basin was at low elevation²⁷,
96 while steppe-desert pollen, presence of evaporites and reduced soil respiration
97 indicate aridity^{1,23,29}. Dating of this sedimentary record has now been extended to the
98 early Eocene based on magnetostratigraphy and radiometric ages^{23,30} allowing us to
99 investigate the Central Asian steppe response to hyperthermal conditions.

100

101 **Increased rainfall during a carbon isotope excursion**

102

103 Early Eocene hyperthermals were driven by the periodic injection of greenhouse gases
104 into the atmosphere from low- $\delta^{13}\text{C}$ reservoirs. This resulted in negative carbon isotope
105 excursions (CIE) across the globe that have been detected in various terrestrial
106 proxies such as plant waxes ($\delta^{13}\text{C}_{\text{wax}}$ of C_{29} to C_{31} *n*-alkanes) or bulk soil organic
107 matter ($\delta^{13}\text{C}_{\text{SOM}}$)⁶. In the Xining Basin, $\delta^{13}\text{C}$ values are relatively high throughout most
108 of the Eocene, but we identify an abrupt and large CIE of -6‰ in $\delta^{13}\text{C}_{\text{SOM}}$ and -8‰ in
109 compound specific $\delta^{13}\text{C}_{\text{wax}}$ (Fig. 2). This is significantly larger than the CIEs related to
110 early Eocene hyperthermals which typically attain magnitudes of up to -4‰^{4,5}, but very
111 similar in both shape and magnitude to continental records of the PETM^{6,32,33}.

112

113 The Xining CIE is preceded by alluvial mudflat deposits at the basin margin (Ledu
114 section) and organic-rich gypsum in the basin center (Bingling Shan and Caijia
115 sections) indicating dry conditions. This is corroborated by the predominance of
116 steppe-desert shrubs in the palynological record and high $\delta^{13}\text{C}$ values which today are
117 only found in extremely dry ecosystems^{24,34} (Fig. 2). The negative CIE starts abruptly
118 in the uppermost gypsum bed and spans ~30 meters into the overlying organic-rich
119 alluvial mudflat deposits³⁰. Steppe-desert pollen decrease during this interval and the
120 vegetation becomes dominated by temperate forests and ferns indicating a shift to
121 wetter conditions. Less water stress experienced by C_3 vegetation and the
122 corresponding shift in biome typically results in lower $\delta^{13}\text{C}$ values^{34,35}. If combined with
123 a dramatic drop of atmospheric carbon isotopic composition ($\delta^{13}\text{C}_{\text{atm}}$), such as seen

124 during hyperthermals^{6,35,36}, these ecological changes could produce the large size of
125 the CIE observed in the Xining Basin.

126

127 Organic matter recorded in soils provides a natural average of the aboveground
128 biomass and can be used to reconstruct the amount of rainfall when corrected for
129 changes in $\delta^{13}\text{C}_{\text{atm}}$ (Fig. 2 and Extended Data Fig. 2)²⁴. Even when assuming an
130 extreme change of $\delta^{13}\text{C}_{\text{atm}}$ such as observed during the PETM^{6,35}, reconstructed mean
131 annual precipitation (MAP) values suggest at least a doubling of MAP in the Xining
132 Basin during the CIE (~574 mm/year) compared to the evaporites below (~177
133 mm/year). Using the $\delta^{13}\text{C}_{\text{atm}}$ value of lower magnitude Eocene hyperthermals would
134 result in an even larger increase in MAP (~984 mm/year) and using the early Eocene
135 baseline value would imply an unrealistic MAP of >4000 mm/year (Supplementary
136 Data 4). Throughout the remainder of the Eocene Xining record, $\delta^{13}\text{C}_{\text{SOM}}$ values
137 remain relatively high²⁹ indicating aridity which fits the predominant redbeds,
138 evaporites and steppe-desert shrubs observed in the record^{1,23}. The CIE in the Xining
139 Basin is thus a unique window where both palynological assemblages and $\delta^{13}\text{C}$ values
140 show a significant greening of the otherwise arid Asian interior. In the following, we will
141 use soil carbonates to test whether this greening was driven by increased westerly
142 moisture or monsoons (Fig. 1).

143

144 **Clumped isotope evidence for summer rainfall**

145 Well-developed pedogenic carbonate nodules are rare in the Xining Basin record but
146 were identified in three horizons of the Ledu section. In consequence, the carbonate
147 component of the alluvial mudrocks was assessed as well using Scanning Electron

148 Microscopy (SEM), X-Ray Diffraction (XRD; Fig. 2; Extended Data Fig. 3) and both
149 optical and cathodoluminescence (CL) microscopy (Extended Data Fig. 4). The strata
150 between 50 Ma and 45 Ma contain mostly calcian dolomite (6-20%), while the intervals
151 above and below are dominated by calcite (1-14%). This calcite is mostly dull
152 luminescent micrite and is interpreted as vadose-grown cement with no evidence of
153 phreatic overprinting (Extended Data Fig. 4 A and B). These cements precipitate either
154 from rising shallow groundwater or percolating soil water and are similar to pedogenic
155 carbonates³⁷. In contrast, the calcian dolomite may have formed either by syn-
156 sedimentary precipitation from saline brines or late diagenetic replacement of calcite
157 cements³⁸. Hence, we focus our isotope investigations solely on the vadose-grown
158 calcite and micritic pedogenic nodules (Extended Data Fig. 4 C) to avoid diagenetic
159 overprinting.

160 Clumped isotope thermometry measures the relative abundance of calcite-derived
161 CO₂ isotopologues of mass 47 compared to the predicted stochastic distribution (Δ_{47}).
162 Since the bonding between the heavy isotopes (¹³C-¹⁸O) is temperature-dependent,
163 this measurement allows reconstructing soil temperatures at the time of carbonate
164 growth without prior knowledge of the isotopic composition of the soil water³⁹. Most
165 Holocene pedogenic carbonates provide Δ_{47} temperatures that are warmer than mean
166 annual temperatures (MAT) because intense soil dewatering and carbonate formation
167 is typically biased towards the summer season^{25,26}. So far, the only location where soil
168 Δ_{47} temperatures have systematically yielded winter temperatures is the monsoonal
169 tropical domain, where winters are dry and warm enough for the accumulation of
170 pedogenic carbonates while summers are typically too wet²⁸. Colder-than-MAT Δ_{47}
171 temperatures can thus be interpreted as a signature for a summer rainy season,

172 especially during Eocene times when winter temperatures were overall warmer,
173 facilitating carbonate growth in the winter²⁸.

174

175 Overall, the pedogenic and vadose-grown carbonates in the Xining strata show a wide
176 spread of Δ_{47} temperatures ranging from 9 ± 3 °C to 38 ± 4 °C (Fig. 2). This range could
177 be due to high evaporative demand throughout the year, resulting in soil drying shortly
178 after erratic rainfall events under variable temperature conditions²⁶. In contrast, the Δ_{47}
179 temperatures during the wetter CIE interval (Fig. 2) are remarkably stable at 20 ± 3 °C
180 and surprisingly low, especially given the hyperthermal conditions. When compared
181 with simulated temperatures in the Xining Basin from the various climate models of
182 the Deep-Time Model Intercomparison Project (DeepMIP)⁴⁰ these simulated
183 temperatures indicate that carbonate growth during the CIE was biased towards the
184 colder winter months (Fig. 3 A). Especially when considering that the soil temperature
185 might have been 1-4°C warmer than the simulated surface air because of ground
186 heating effects⁴¹.

187

188 The Δ_{47} temperatures during the CIE therefore suggest that soil drying in the Xining
189 Basin occurred during the winter season, while the increased rainfall recorded in the
190 palynological assemblage and $\delta^{13}\text{C}$ values must have occurred mostly during the
191 summer. This is similar to the modern monsoonal setting in which pedogenic
192 carbonates near Xining are formed during the fall and after the wet season⁴².
193 Reconstructed soil water $\delta^{18}\text{O}$ ($-6 \pm 1.5\text{‰}$; Extended Data Fig. 5 and 6) is similar to
194 coeval records of both summer monsoonal and westerly moisture in the region¹⁰, but
195 the cold Δ_{47} temperatures in combination with increased MAP indicate an
196 intensification of summer monsoons deep within the Asian continent shifting the timing

197 of carbonate formation to the winter season. Cold Δ_{47} temperatures are not unique to
198 the CIE and reappear episodically in the Xining Basin ~10 million years later and
199 persist into the Oligocene (Fig. 3 A), possibly due to the combined effects of global
200 cooling and an increased role for summer monsoonal moisture²⁷. However, the
201 palynological records show that the region remained extremely dry throughout the late
202 Eocene to Oligocene¹ and unlike the intensified monsoon seen during the CIE (Fig.
203 2).

204

205 **Proto-monsoon intensification in hyperthermal conditions**

206

207 The simulated temperatures fit well with global proxy compilations⁴³ and the climate
208 models show mutual agreement⁴⁰ (Fig 3 A), however, differences between models are
209 large for simulated precipitation³ (Fig. 3 B). The ‘drier’ models show aridity in the Xining
210 Basin throughout the year (e.g. COSMOS and IPSL), while the ‘wetter’ models reveal
211 a summer rainy season (e.g. CESM and HadCM3) not unlike pre-industrial conditions
212 (Extended Data Fig. 1). Simulated MAP is relatively low and ranges from 25 to 469
213 mm/year (Fig. 3 B), which fits with the inland evaporative basins in the region⁴⁴ and
214 our proxy estimates for the Paleocene and Eocene background conditions in the
215 Xining Basin. However, even the ‘wetter’ simulations are lower than our reconstructed
216 MAP of 574 mm/year for the intensified monsoon during the CIE.

217

218 The distribution of monsoonal domains⁴⁵ in the simulations shows that the East Asian
219 monsoon was less prevalent in the Eocene compared to today (Fig. 4; Extended Data
220 Fig. 7) and that southern Asia was influenced by the Intertropical Convergence Zone
221 (ITCZ) as suggested by paleobotanical data¹⁹. Intriguingly, some of the Eocene

222 simulations also show a monsoonal domain along the East Asian coast (CESM,
223 COSMOS and GFDL), in particular in the 6x and 9x pCO₂ simulations, which would
224 have resembled the early Eocene atmosphere⁸ (Fig. 4; Extended Data Fig. 7). This
225 East Asian proto-monsoon agrees well with paleobotanical compilations showing a
226 humid coast line^{17,22}, but none of the DeepMIP models reproduce monsoons extending
227 far into the continental interior to reach the Xining Basin.

228

229 Other modeling studies have shown that Eocene monsoons can be triggered in
230 extreme orbital settings with amplified northern hemisphere heating^{20,21}. The PETM
231 and other hyperthermal events correspond to high eccentricity⁷ and these unique
232 conditions combined with extreme greenhouse warming likely conspired to intensify
233 the proto-monsoons to reach the continental interior of Asia. The shape, large
234 magnitude and thickness of the Xining CIE, and its occurrence within a long reversed
235 polarity zone correlated with chron C24r (57.1 Ma to 53.9 Ma) allow us to attribute the
236 CIE to the PETM (56 Ma), which also fits within the radiometric age constraints (Fig.
237 2; Extended Data Fig. 8). Alternatively, it could be ETM2 (54 Ma) but this hyperthermal
238 is considerably smaller in size, shorter and consists of two excursions (H1 and H2)
239 that have been observed in both marine and terrestrial records^{4,5,32} but are absent in
240 the Xining Basin. Furthermore, negative CIEs with a shape and size similar to the
241 Xining CIE have been observed in the continental Nanyang and Fushun Basins in
242 eastern China and are also linked to reduced water stress during the PETM³⁶. As the
243 largest hyperthermal, the PETM is therefore the most likely driver for this unique event
244 of monsoon intensification and greening in Central Asia.

245

246 Finally, we note that the Eocene proto-monsoons were significantly different from the
247 modern-day East Asian monsoon, which is mostly driven by interaction between the
248 subtropical jet stream and the high topography of the Tibetan Plateau¹². Eocene proto-
249 monsoons occur even in the DeepMIP simulations with no Tibetan Plateau at all (Fig.
250 4) and are driven by a distinct low atmospheric pressure anomaly developing during
251 summer over the vast and flat Asian continent (Extended Data Fig. 9). This suggests
252 that the proto-monsoons were similar to a large sea breeze driven by land-sea thermal
253 contrasts¹⁸ and explains their sensitivity to hyperthermal conditions. Global cooling
254 after the early Eocene reduced regional thermal contrasts, dampening the ability of
255 proto-monsoons to episodically expand inland and could explain the overall
256 aridification trend of Central Asia throughout the Eocene and culminating at the
257 Eocene-Oligocene Transition (EOT; 33.9 Ma)^{27,29,46}. Only until the subsequent surface
258 uplift and northward expansion of the Tibetan Plateau during the Neogene would the
259 modern-day East Asian monsoon have returned to the Xining Basin and the Asian
260 interior¹¹⁻¹⁴.

261

262 Our study thus reveals a prominent role for Eocene proto-monsoons in occasionally
263 greening the otherwise dry Asian interior under hyperthermal conditions. The PETM,
264 but also other hyperthermals such as the Middle Eocene Climatic Optimum (MECO)
265 are associated with major mammal dispersals in and out of Asia^{47,48} and future work
266 should consider the role of proto-monsoons in shifting biomes⁴⁹ and opening dispersal
267 corridors within continental Asia. Replacing the steppe-desert by forested ecosystems
268 could have also played a major role in buffering the massive carbon release during
269 hyperthermals by enhancing soil organic carbon burial and silicate weathering⁵⁰.
270 Finally, our work provides paleoclimate evidence highlighting the abrupt and non-linear

271 response of Asian monsoons to extreme greenhouse warmth. Although the Eocene
272 Asian paleogeography and hydroclimate were markedly different from today, the
273 mechanisms leading to abrupt greening of the Asian interior should be further tested
274 and simulated because of their significance for predicting future rainfall patterns.

275

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293 climate model simulation data.

294

295 **Author contributions**

296 N.M., A.L. and G.D.-N. designed the study. N.M., A.L., A.W., A.R., A.S. and G.D.-N.
297 conducted the field work and collected the samples. N.M. performed the XRD and
298 petrographic analyses. A.W., F.R.-C. and C.H. prepared and analysed the
299 palynological samples. A.R. and M.T.H. performed the *n*-alkane analyses. N.M., A.L.,
300 A.J.S. and A.M. conducted the stable isotope analyses. N.M., M.T. and J.F. performed
301 the clumped isotope analyses. J.B. and F.D.K. analysed and plotted the climate model
302 simulation data. N.M. wrote the manuscript with major input from A.L., G.D.-N. and
303 A.M. All co-authors contributed to interpreting the data and writing the manuscript.

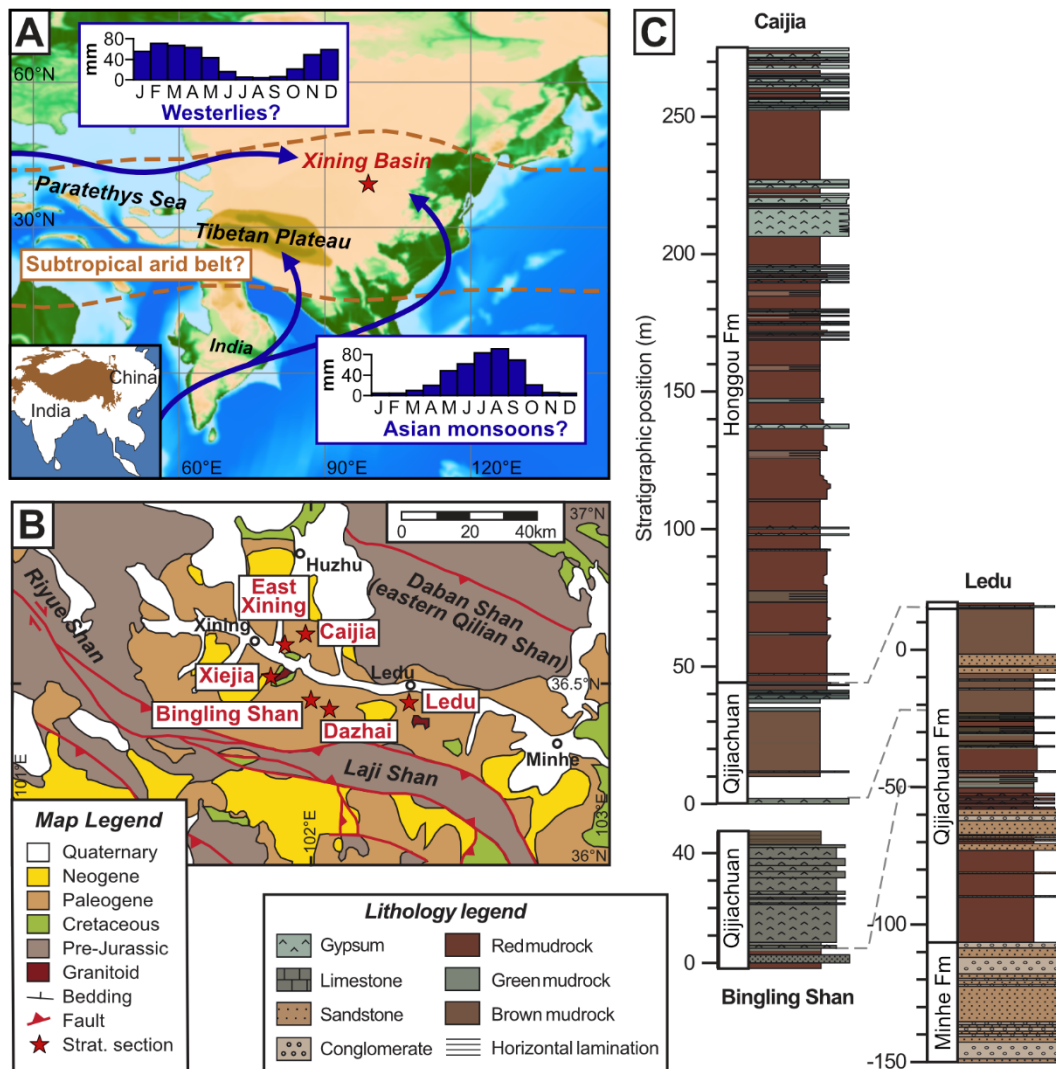
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305 **Competing Interests Statement**

306 The authors declare no competing interests.

307

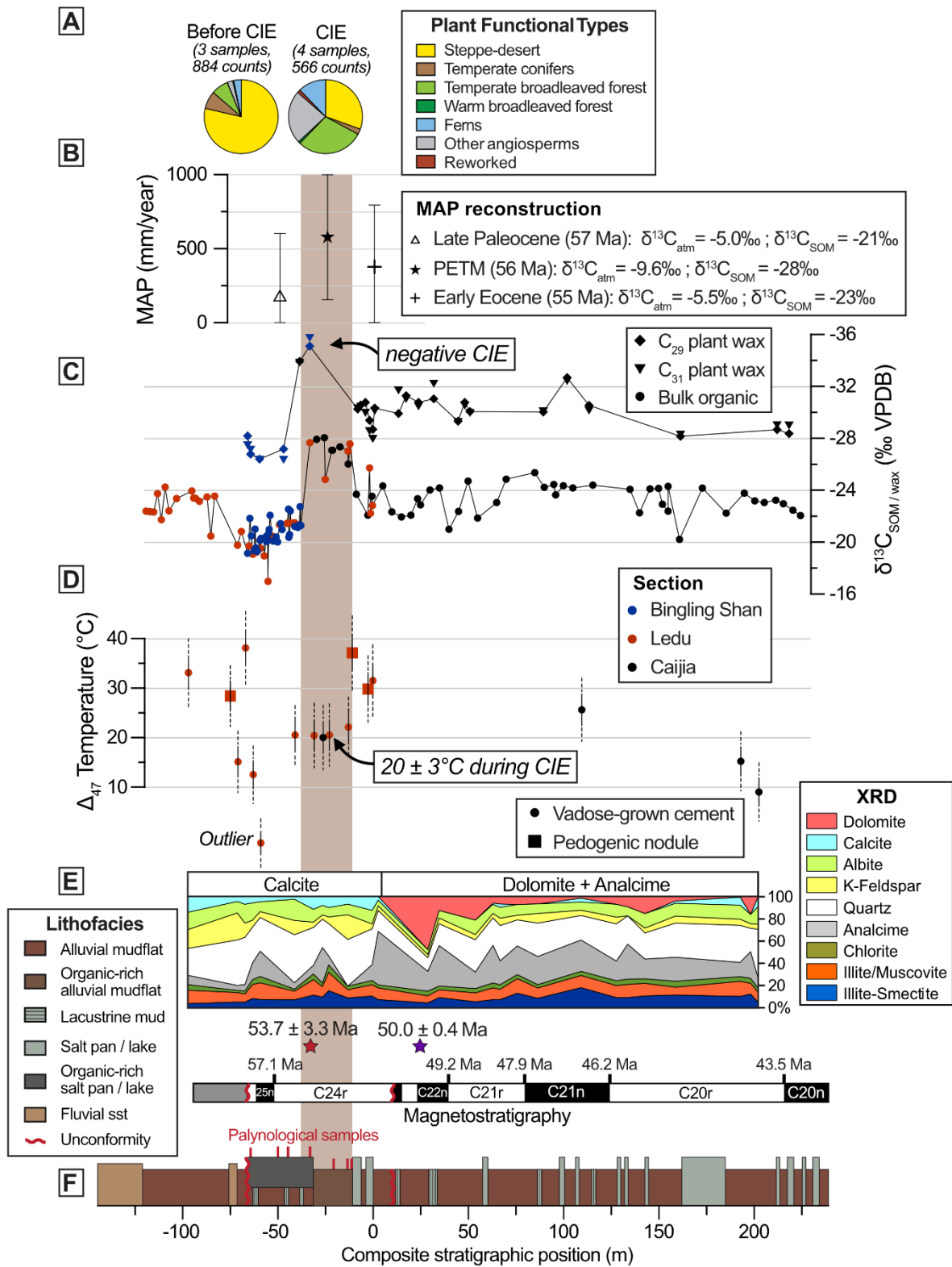
308 **Figures**



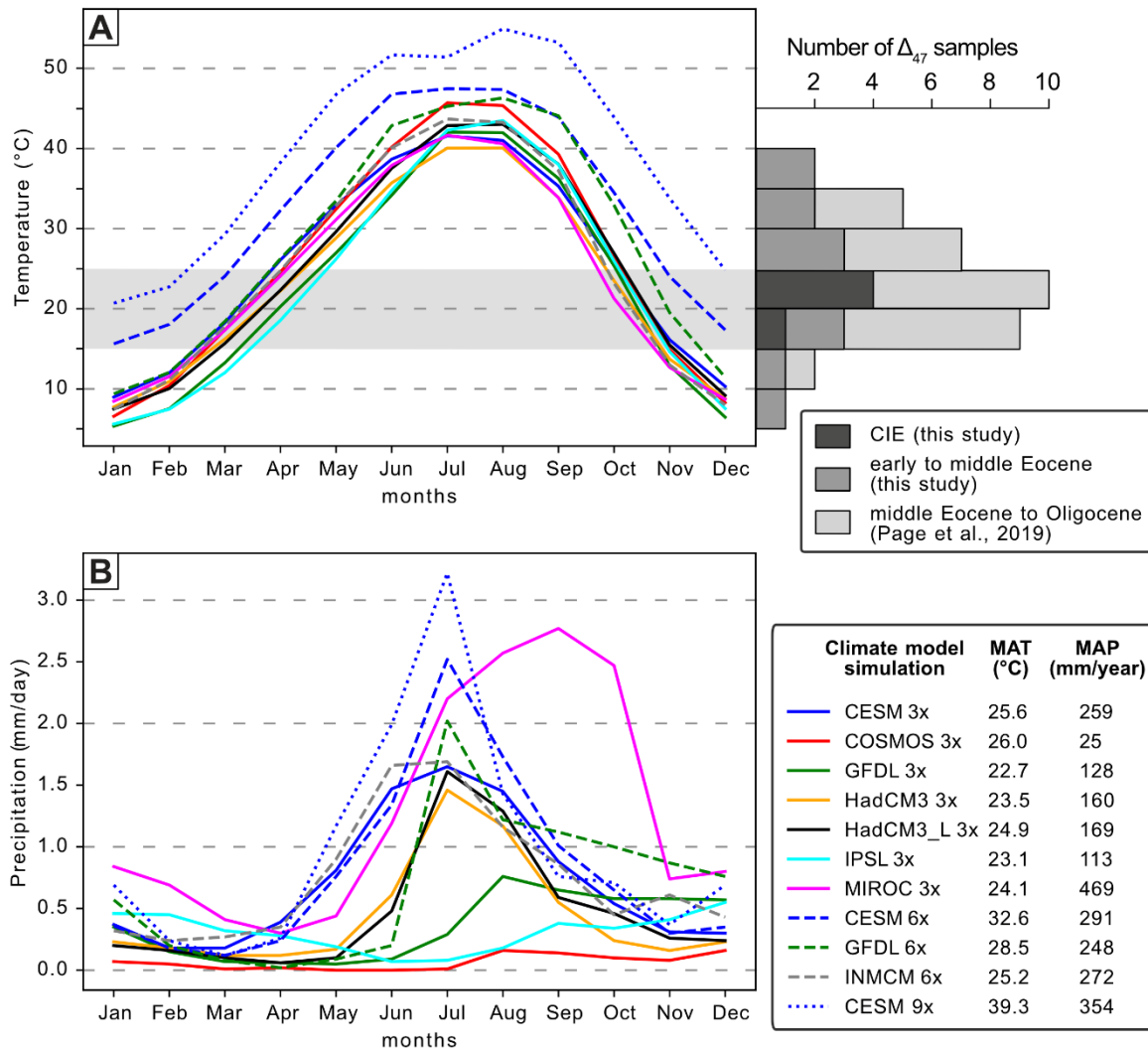
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310 *Fig. 1. Geological setting showing hypothesized Eocene hydroclimates in Asia*
 311 *including a subtropical arid belt, westerly or monsoonal domain. A) Paleogeographic*
 312 *map of Asia at 50 Ma (million years ago)³¹ showing the location of the Xining Basin.*
 313 *Blue arrows show potential moisture sources. Histograms reflect the seasonal*
 314 *distribution of this rainfall based on modern-day observations in monsoonal Xining and*
 315 *westerly-dominated Tashkent ([https://www.ncei.noaa.gov/products/wmo-climate-](https://www.ncei.noaa.gov/products/wmo-climate-normals)*
 316 *[normals](https://www.ncei.noaa.gov/products/wmo-climate-normals)).* *Climate model simulations indicate a similar seasonal distribution of rainfall*
 317 *in these domains during the Eocene^{9,20}. Inset shows present-day geographical*
 318 *location. B) Geological map of the Xining Basin (modified from ref³⁰) showing the*

319 locations of the studied sections and palynological sampling. C) Lithostratigraphy of
 320 the Caijia, Bingling Shan and Ledu key sections^{23,30}.



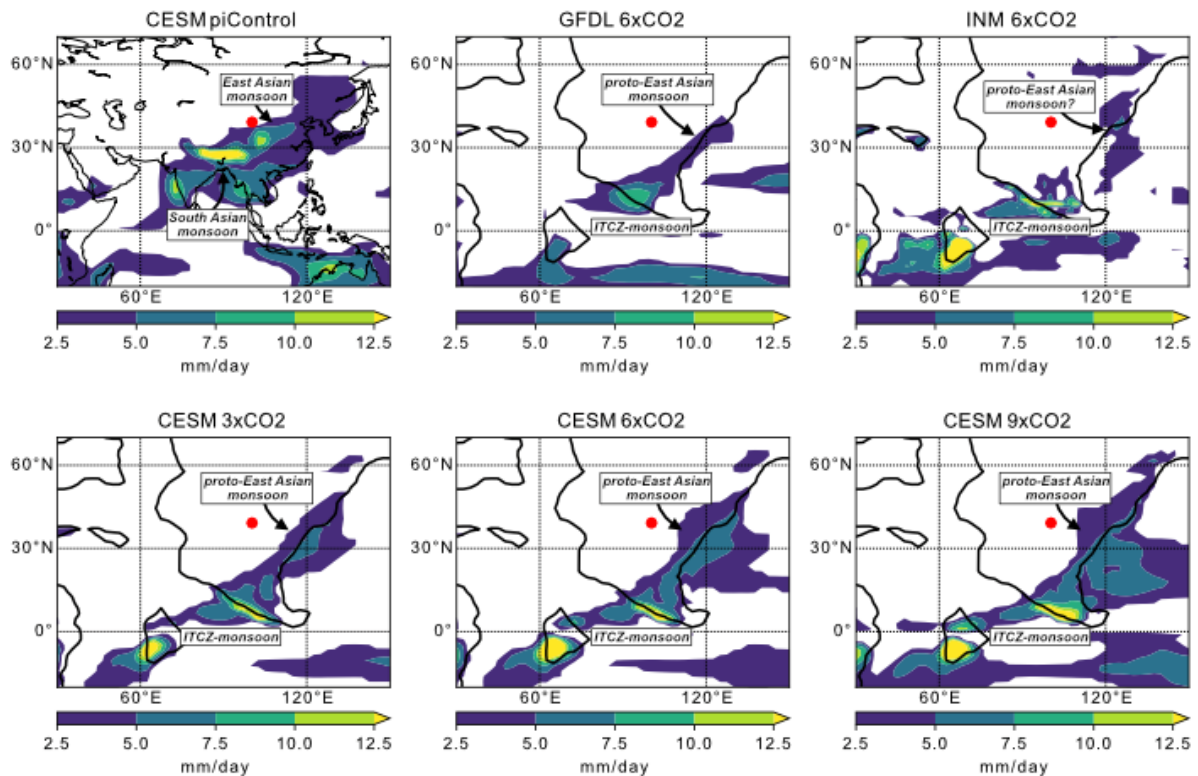
322 *Fig. 2. Paleo-environmental reconstruction showing temperate forests replacing*
323 *steppe shrubs during a negative carbon isotope excursion (CIE). The CIE is indicated*
324 *by the brown shaded vertical bar. Data include: A) palynological data classified in Plant*
325 *Functional Types, B) reconstructed Mean Annual Precipitation (MAP) using late*
326 *Paleocene, PETM and early Eocene $\delta^{13}\text{C}_{\text{atm}}$ and $\delta^{13}\text{C}_{\text{SOM}}$ values (root mean square*
327 *error of ± 422 mm/year of the MAP proxy is shown)²⁴, C) $\delta^{13}\text{C}$ values of both soil*
328 *organic matter and n-alkanes, D) Δ_{47} temperatures of vadose-grown carbonates (with*
329 *solid and dashed lines indicating 68% and 95% confidence intervals respectively*
330 *based on at least 13 replicates), one sample produced an unrealistically low*
331 *temperature of -1.3°C and is considered an outlier, E) X-Ray Diffraction (XRD) derived*
332 *mineral content, F) age constraints^{23,30} based on radiometric ages (stars) and*
333 *magnetic polarities (black ticks), lithostratigraphy showing stratigraphic positions of*
334 *palynological samples with red marks.*



335

336 *Fig. 3. Proxy comparison with simulated surface air temperature (A) and precipitation*
 337 *(B) in the Eocene Xining Basin. Averaged values were derived from a box of 1.875°*
 338 *latitude by 2.5° longitude centred at the paleolocation of the Xining Basin (102.5°N,*
 339 *39.5°E). Simulations are derived from the DeepMIP climate model ensemble (colors*
 340 *indicate each model) under an early Eocene paleogeography with no Tibetan Plateau*
 341 *and a Paratethys Sea lowstand, preindustrial orbital configuration and 3x, 6x or 9x*
 342 *preindustrial pCO₂ when available (solid, dashed and dotted lines respectively)⁴⁰. The*
 343 *histogram in A) shows the number of measured Δ_{47} temperatures in 5°C bins. CIE*
 344 *temperatures of ~20°C are best explained by soil dewatering and carbonate formation*

345 during the colder winter months. Simulated precipitation in B) varies from dry
 346 throughout the year to summer wet similar to the present-day monsoons.



347
 348 *Fig. 4. Distribution of monsoonal domains in climate model simulations. Upper panel*
 349 *shows a pre-industrial control simulation (CESM) and two early Eocene simulations*
 350 *with 6x pCO₂ (GFDL and INM). Lower panel shows different pCO₂ scenarios (3x, 6x*
 351 *and 9x) of the CESM model. Additional simulations are shown in Extended Data Fig.*
 352 *7. Summer (MJJAS) minus winter precipitation (NDJFM) is used as a metric and*
 353 *regions where this difference is >2.5 mm/day are considered monsoonal⁴⁵. Red dot*
 354 *indicates the location of the Xining Basin. All Eocene simulations show an ITCZ*
 355 *(Intertropical Convergence Zone) monsoon near the equator. CESM and GFDL show*
 356 *a proto-monsoonal domain along the East Asian coast, but none of the models*
 357 *simulate monsoons extending far into the Asian interior.*
 358

359 **References**

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581

582 **Methods**

583

584 *Sampling*

585

586 Rock samples were collected after digging trenches of at least ~0.5 meter deep in the
587 Bingling Shan (BS), Caijia (CJ) and Ledu (LD) sections covering the early to middle
588 Eocene^{23,30}. In addition, three horizons with pedogenic carbonate nodules were
589 identified and sampled at the Ledu section. Samples for plant wax analyses were
590 collected using gloves and wrapped in aluminum foil to prevent contamination.
591 Palynological samples were collected from the Bingling Shan and Caijia sections as
592 well as the East Xining (EX), Xiejia (XJ) and Dazhai (DZ) sections^{23,30}.

593

594 *Composite section and age model*

595

596 The individual sections were combined into one composite section using litho- and
597 magnetostratigraphic correlations³⁰. First, we assembled the core of the composite
598 section by stitching the Honggou Fm of the Caijia section (CJ) and the Qijiachuan Fm
599 of the Ledu section (LD) together at the liver-brown to red mudrock transition observed
600 in both locations (transition from Qijiachuan to Honggou Fm; blue dashed line Fig. 2
601 in ref³⁰). We assume this to be a chronostratigraphic marker and define it as the 0
602 meter-level in the composite log. The stratigraphic positions of these sections can be
603 converted to the composite stratigraphic position as follows:

604

605 Comp m level = CJ [44.5m to 280m] - 44.5

606 Comp m level = LD [entire section] - 15

607

608 Next, we project the Qijiachuan Fm of the Caijia (CJ) and Bingling Shan (BS) sections
609 on the composite section making sure that the lithostratigraphic boundaries between
610 the evaporites of the middle Qijiachuan Fm and the liver-colored mudrocks of the
611 upper Qijiachuan Fm correlate. This is necessary because the thicknesses of the
612 formations vary between the sections and the transition between these formations is
613 assumed to represent a chronostratigraphic marker. The stratigraphic positions of
614 these sections can be converted to the composite stratigraphic position as follows:

615

616 Caijia:

$$\begin{aligned} 617 \text{ Comp m level} &= (\text{CJ}[2\text{m to } 44.5\text{m}] - 44.5) * (1 - 5.9 / 42.5) \\ 618 &= \text{CJ}[0 \text{ to } 2\text{m}] - 38.6 \end{aligned}$$

619 Bingling Shan:

$$\begin{aligned} 620 \text{ Comp m level} &= ((\text{BS}[5.2\text{m to } 43\text{m}] - 79.6) + ((\text{BS}[5.2\text{m to } 43\text{m}] - 43) / -37.8) * 8.05) \\ 621 &= \text{BS}[43\text{m and higher}] - 79.6 \end{aligned}$$

622

623 Based on our observation of the large negative CIE, we propose to revise the age
624 model of ref³⁰. The previous age model is reliable down to chron C22n and constrained
625 by the radiometric age of a tuff²³. However, chron C23n below is very short and does
626 not fit the pattern of the geomagnetic polarity timescale³⁰. Hence, we infer an
627 unconformity at the base of C23n which also corresponds to the lithological shift from
628 calcareous to dolomitic mudrocks (Fig. 2). The negative CIE occurs within a long
629 reversed polarity zone on top of a short normal zone. This polarity pattern fits well with
630 C25n and C24r and is within the radiometric age constraints of 53.7±3.3 Ma recorded
631 in a carbonate bed just below the CIE (Fig. 2)³⁰. The PETM and ETM2 occur in C24r

632 and could correlate with the CIE in the Xining Basin as discussed in the main text.
633 Throughout the manuscript we use the geomagnetic polarity timescale of ref⁴ for
634 correlating our paleo-environmental data with the Cenozoic Global Reference benthic
635 foraminifera carbon and oxygen Isotope Dataset (CENOGRID).

636

637 *XRD*

638

639 X-ray diffractometry (XRD) analyses were performed on 27 mudrock samples and 5
640 carbonate beds throughout the stratigraphy to assess the mineralogy and the
641 carbonate type. Powdered rock samples were measured using a PANALYTICAL
642 X'Pert Bragg-Brentano diffractometer using a copper beam powered by 30 mA and 40
643 KV generator current. The characteristic diffraction maxima of the mineral phases
644 were determined using the MacDiff software ([http://mill2.chem.ucl.ac.uk/ccp/web-](http://mill2.chem.ucl.ac.uk/ccp/web-mirrors/krumm/html/software/macdiff.html)
645 [mirrors/krumm/html/software/macdiff.html](http://mill2.chem.ucl.ac.uk/ccp/web-mirrors/krumm/html/software/macdiff.html)). Intensities were converted to 1°
646 divergence characteristics and weighted by reference intensity ratios (RIR) to
647 semiquantitatively calculate the relative abundance of each mineral phase. The
648 identified mineral phases are (Extended Data Fig. 3): mixed-layer illite-smectite (main
649 diffraction maxima position: 10.4 Å, RIR value: 0.52), illite/muscovite (10.0 Å, 0.43),
650 chlorite (14.0 Å, 1.00), analcime (3.43 Å, 1.41), quartz (4.26 Å and 3.34 Å, 3.03) K-
651 feldspar (3.25 Å, 0.60), albite (3.20 Å, 0.64), calcite (3.04 Å, 2.20) and calcian dolomite
652 (2.91 Å, 2.51). The mol% of MgCO₃ in the calcian dolomite was estimated using the
653 relative position of the 104 peak (d in Å) compared to calcite (3.035 Å) and
654 stoichiometric dolomite (2.885 Å):

$$655 \quad \text{Mol\% MgCO}_3 = \frac{3.035 - d}{3.035 - 2.885} \times 50\%$$

656

657 *Compound-specific lipid biomarkers and stable isotopes*

658 Plant waxes were extracted from 25 mudrock and gypsiferous mudrock samples at
659 the Organic Surface Geochemistry Lab (OSGLab) at the German Research Centre for
660 Geosciences (GFZ) in Potsdam. Samples were crushed and then powdered using a
661 shatterbox with an agate grinding chamber. Soluble organic matter was extracted
662 using a Dionex accelerated solvent extractor (ASE 350) with a
663 dichloromethane/methanol mixture of 9:1 at 100°C and 1500 psi. Total extracts of
664 three 18-minute cycles were captured in 250 ml bottles, later concentrated to 4 ml in
665 a Turbovap, and then separated on silica gel using a solid phase extraction (SPE) as
666 described in ref⁵¹. The C₂₇, C₂₈, C₂₉, C₃₀, C₃₁, C₃₂, and C₃₃ *n*-alkanes were identified
667 and quantified using a gas chromatograph with a coupled flame ionization and mass-
668 selective detector (GC-FID/MSD Agilent 7890A GC, 5975C MSD) flushed with helium
669 carrier gas. Temperatures in the GC oven were programmed to increase at a rate of
670 12°C/min starting from 70°C to 320°C at which temperatures were held constant for
671 21 min. The PTV injector had a split ratio of 5:1 at an initial temperature of 70°C. The
672 *n*-alkane FID-peak areas were compared with a previously added 5 α -androstane
673 standard from which *n*-alkane concentrations were calculated. The *n*-alkane
674 concentrations are reported as $\mu\text{g}/\text{gram}$ dry sediment. In addition the average chain-
675 length (ACL) and the carbon preference index (CPI) were calculated.

676

677 For all samples compound specific stable isotope values were measured using a GC-
678 Isolink coupled to a Thermo Scientific MAT 253 isotope ratio mass spectrometer
679 (IRMS) at the Stable Isotope and Organic Molecular Biogeochemistry Laboratory at
680 the University of Connecticut. The *n*-alkane fractions were dried and concentrated to
681 200 $\mu\text{g}/\mu\text{l}$ per compound in hexane for $\delta^2\text{H}_{\text{wax}}$ measurements. Each sample was

682 measured in at least duplicate and in most samples triplicate analysis. The injector of
683 the GC was operated in splitless mode at 300°C and the oven was held at 70°C for 2
684 minutes. The oven was heated at 15°C/min until 150°C, then heated with 5°C/min to
685 320°C. The final temperature was held for 10 minutes. The column effluent was
686 transferred into an isotope ratio mass spectrometer after conversion to H₂ in a high-
687 temperature oven at 1420°C. Stable isotopes are reported in delta notation in permil
688 (‰). The standard deviation for each sample was better than 3‰ and the $\delta^2\text{H}_{\text{wax}}$ are
689 reported in Vienna Standard Mean Ocean Water scale (VSMOW). $\delta^{13}\text{C}_{\text{wax}}$ values were
690 measured using the same fractions used for $\delta^2\text{H}_{\text{wax}}$. The *n*-alkane fractions were
691 concentrated to 60 µg/µl in hexane for $\delta^{13}\text{C}_{\text{wax}}$ measurements. The same instrumental
692 setup (GC-IRMS) and temperature programming was used as for the $\delta^2\text{H}_{\text{wax}}$
693 measurements, only the oxidation oven was run at a lower temperature of 960°C.
694 Duplicates were measured for each sample and a CO₂ gas with known isotopic
695 composition was used as reference gas and $\delta^{13}\text{C}_{\text{wax}}$ values are reported in Vienna
696 Pee Dee Belemnite scale (VPDB). The analytical precision of each sample had a
697 typical standard deviation of 0.5‰. Isotopic compositions were standardized using a
698 suite of *n*-alkanes in a standard mixture.

699

700 *Stable and clumped isotopes*

701

702 Bulk organic stable isotopes were measured on 113 mudrock and gypsiferous
703 mudrock samples. These samples were decarbonated twice with 6N HCl and washed
704 with 18 M-Ohm water several times for acid removal; carbon content and organic
705 matter $\delta^{13}\text{C}$ values were then measured. Samples from the Ledu section were
706 measured using a Costech ECS 4010 Elemental Analyzer coupled to a Thermo

707 Scientific MAT 253 gas source IRMS at the University of Washington. Samples from
708 the Caijia section were measured using a Flash Elemental Analyzer 1112 connected
709 to a Thermo Scientific MAT 253 IRMS at the Goethe University-Senckenberg BiK-F
710 Stable Isotope Facility in Frankfurt. Samples from the Bingling Shan section were
711 measured using a Flash Elemental Analyzer 2000 coupled to a Thermo Scientific
712 Delta-V IRMS at the Alfred-Wegener Institute Isotope Laboratory (AWI) in Potsdam.
713 Measured $\delta^{13}\text{C}$ values (all relative to VPDB) were corrected using in-house reference
714 materials and long-term repeatability was below 0.2‰. International reference
715 materials USGS 24 and IAEA-CH-7 were analyzed along with the samples to check
716 for accuracy.

717

718 Carbonate stable isotopes were measured on 51 calcite-bearing mudrocks and 3
719 pedogenic carbonate nodules. The calcite-bearing mudrocks were reacted with 105
720 % orthophosphoric acid and carbonate $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values were then measured.
721 Samples from the Ledu section were measured using a Thermo Scientific Kiel III
722 Carbonate Device coupled to a Thermo Scientific Delta Plus IRMS at University of
723 Washington. Samples from the Caijia section were measured with a Thermo Scientific
724 GasBench II interfaced with a Thermo Scientific MAT 253 IRMS at the Goethe
725 University-Senckenberg BiK-F Stable Isotope Facility. Measured carbon and oxygen
726 isotopic compositions were corrected using in-house reference materials as well as
727 NBS18 and NBS19 carbonate reference materials. Long-term repeatability was below
728 0.03‰ for $\delta^{13}\text{C}$ and below 0.06‰ for $\delta^{18}\text{O}$. All $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values are expressed
729 relative to VPDB.

730

731 Clumped isotopes were analyzed on 16 of these calcite-bearing mudrocks and on the
732 3 pedogenic nodules at the Goethe University-Senckenberg BiK-F Stable Isotope
733 Facility, using a Thermo Scientific Kiel IV Carbonate Device paired with a Thermo
734 Scientific MAT 253 Plus IRMS. The Kiel IV was modified from its original configuration
735 with the addition of two water traps and a different heating-cooling system on the
736 purification trap. Depending on the carbonate content, 110 to 2500 μg of sample was
737 reacted with 105% orthophosphoric acid at 70°C for 5 minutes starting from the last
738 acid drop. In each sequence, half of the analyses were reserved for measurement of
739 the ETH carbonate standards⁵², with ETH 1 and 2 measured in equal numbers as ETH
740 3, and with ETH 4 less frequently. Data were corrected for negative background using
741 intensities measured on the 47.5 Faraday cup during acquisition, scaled by a factor
742 that minimizes the slope between ETH 1 and 2. Baseline corrected data were
743 projected into the Intercarb Carbon-Dioxide Equilibrium Scale (I-CDES)⁵² using the
744 “independent” approach in the python package D47crunch⁵³. The resulting Δ_{47} values
745 were converted into temperature using the calibration of ref⁵⁴. Each sample was
746 measured in 13 to 18 replicates and the errors are reported as both 68% and 95%
747 confidence intervals (1SE and 2SE respectively). Long-term repeatability of both
748 samples and ETH standards was 0.035‰.

749

750 The $\delta^{18}\text{O}$ values of the soil water ($\delta^{18}\text{O}_{\text{water}}$) are reconstructed using the Δ_{47} -derived
751 temperatures and the equation of ref⁵⁵. These values are compared with modern-day
752 $\delta^{18}\text{O}_{\text{water}}$ values of the summer monsoon and wintertime westerlies⁵⁶ (Extended Data
753 Fig. 5).

754

755 *Palynology*

756

757 The palynological samples were prepared by Palynological Laboratory Services
758 (Anglesey, UK) using standard palynological techniques (HCL, HF, sieving at 10 μ m),
759 the University of Amsterdam (Amsterdam, the Netherlands; HCL, sieving at 212 μ m,
760 20% pyrophosphate, acetolysis, bromoform treatment), and at the Sediment
761 Laboratory of the Vrije Universiteit (Amsterdam, the Netherlands; following methods
762 of ref⁵⁷ and described in ref⁵⁸). Plant Functional Types (PFTs) were assigned using
763 the Nearest Living Relatives (NLR) approach applied by ref⁵⁹.

764

765 *MAP reconstruction*

766

767 We use an empirical relationship which can be used to reconstruct Mean Annual
768 Precipitation (MAP) using the $\delta^{13}\text{C}_{\text{SOM}}$ values of top soils (O- and A-horizons)²⁴. The
769 organic matter in deeper horizons might be affected by decomposition resulting in
770 fractionation to higher $\delta^{13}\text{C}_{\text{SOM}}$ values⁶⁰, which would lead to an underestimation of the
771 MAP. The alluvial mudflat deposits of the Xining Basin show clear evidence of
772 pedogenesis such as vertic slickensides, mudcracks, mottling, blocky peds, and clay
773 coatings in thin sections, but no obvious horizonation due to overprinting of stacked
774 paleosols. It is therefore impossible to preselect only O- and A-horizons. Nevertheless
775 we argue that the bulk $\delta^{13}\text{C}_{\text{SOM}}$ values and the CIE are not strongly affected by
776 degradation because we observe no clear relationship between $\delta^{13}\text{C}_{\text{SOM}}$ and total
777 organic content (Extended Data Fig. 2) and the $\delta^{13}\text{C}_{\text{SOM}}$ values of alluvial mudflats are
778 indistinguishable from laterally equivalent gypsum beds which have a high quality of
779 organic preservation⁵⁹. In addition, an even larger CIE is observed in the *n*-alkanes

780 ($\delta^{13}\text{C}_{\text{wax}}$) which are less affected by degradation⁶¹. If this difference in CIE is caused
781 by the decomposition of bulk organic matter, the actual CIE and precipitation changes
782 could be even larger and our MAP reconstructions should be considered as a minimum
783 estimate.

784

785 Before using the paleoprecipitation proxy, we correct the $\delta^{13}\text{C}_{\text{SOM}}$ values to account
786 for the different CO_2 $\delta^{13}\text{C}$ values ($\delta^{13}\text{C}_{\text{atm}}$) of the Paleocene/Eocene atmosphere
787 compared to the pre-industrial value of -6.5‰ that is implicit in the empirical relation²⁴.
788 We apply a -1.5‰ correction to the $\delta^{13}\text{C}_{\text{SOM}}$ values to correct for the baseline
789 Paleocene ($\delta^{13}\text{C}_{\text{atm}} = -5.0\text{‰}$)⁶², a $+3.1\text{‰}$ correction for the PETM assuming a lowering
790 of 4.6‰ ($\delta^{13}\text{C}_{\text{atm}} = -9.6\text{‰}$) through the injection of ^{13}C -depleted carbon^{35,62} and a -
791 1.0‰ correction for the early Eocene baseline ($\delta^{13}\text{C}_{\text{atm}} = -5.5\text{‰}$)⁶². Other early Eocene
792 hyperthermals corresponded to smaller isotope excursions than the PETM and we use
793 a lowering of 3.0‰ ($\delta^{13}\text{C}_{\text{atm}} = -8.5\text{‰}$) to illustrate the effect on the MAP reconstruction.
794 The MAP is then calculated using the following equation²⁴:

795

$$796 \text{MAP (mm/year)} = \exp((\delta^{13}\text{C}_{\text{SOM}} + 11.94) / -2.04)$$

797

798 *Carbonate characterization*

799 We have characterized the carbonates of the Xining strata following the methods of
800 ref²⁷. Thin sections were prepared for six representative mudrock samples and three
801 carbonate nodules. These were examined using cathodoluminescence (CL) and
802 plane- and cross-polarized microscopy to examine the origin and possible diagenesis
803 of the carbonates. The samples were examined at the University of Potsdam using an
804 Olympus BXFM-F polarizing microscope equipped with a lumic hot-cathode electron

805 source and an Olympus XC10 digital color camera. In addition, seven powdered
806 samples were mounted on a stub and carbon coated. These were further examined at
807 the University of Potsdam using a JEOL JSM-6510 Scanning Electron Microscope
808 (SEM) coupled to an Oxford Instruments INCAx-act Energy Dispersive X-ray
809 Spectrometer (EDS) to assess their mineralogy.

810 The calcite is a dull luminescent micrite and is observed throughout the clayey matrix
811 (Extended Data Fig. 4 A, B). Some brighter specks of microspar are observed as well
812 but no detrital carbonate grains nor blocky sparite in cracks, veins or vugs which would
813 have been typical for phreatic cements^{37,63}. These observations are very similar to
814 ref²⁷ and we follow their interpretation of an early diagenetic cement forming in the
815 vadose zone from either shallow groundwater or percolating soil water^{37,63}. The
816 $\delta^{18}\text{O}_{\text{carb}}$ values are consistent with coeval pedogenic carbonates in the region¹⁰ and
817 the $\delta^{13}\text{C}_{\text{carb}}$ are very low ($\sim -9\text{‰}$ in Caijia and $\sim -11\text{‰}$ in Ledu) as is expected during the
818 negative CIE (Extended Data Fig. 5). The stable isotope values do show some positive
819 linear relation (Extended Data Fig. 6) indicating the effects of evaporation⁶⁴. This
820 suggests that the carbonates formed relatively close to the surface of the paleosol.
821 The calcite occurs in a specific stratigraphic interval in multiple locations of the basin
822 (Fig. 2), which suggests a primary origin rather than precipitation/precipitation during
823 modern-day weathering which would have affected all mudrocks. In addition, Eocene
824 climatic events such as the EOT have been recorded in these calcite cements with no
825 indication for overprinting²⁷.

826 In addition to the calcareous mudrocks, three horizons with pedogenic nodules were
827 observed in the Ledu section. Most of these nodules are entirely composed of micrite
828 and floating detrital grains (Extended Data Fig. 4 C), but some contain vugs filled with

829 sparite indicating a phreatic overprint. These were carefully avoided when powdering
830 the samples for isotopic analyses. The nodules occurring in the coarse sandstone (at
831 53.3 Ma) show features of a mixed origin. The floating grains show an isopachous rim
832 of microspar / sparite while the rest of the matrix is filled with micrite (Extended Data
833 Fig. S4 D). This indicates an initial vadose origin for the micrite which is overprinted
834 by phreatic cements filling circum-granular cracks³⁷.

835 Dolomite occurs as rhombic crystals within the mudrock matrix (Extended Data Fig. 4
836 E and F), but also as massive cm-scale beds interbedding lacustrine mudrocks
837 (Extended Data Fig. 4 G) and as cm-scale burrows (Extended Data Fig. 4 H). The
838 SEM-EDS analysis revealed that the dolomites are non-ferroan but calcian dolomite
839 with an XRD peak at $\sim 2.91 \text{ \AA}$ (Extended Data Fig. 3) and MgCO_3 contents ranging
840 from 44 to 46%. No dolomite has been reported from the mountain ranges surrounding
841 the Xining Basin⁶⁵ and a detrital origin is therefore unlikely. Instead, the dolomite could
842 have formed either by diagenetic replacement of calcite or syn-sedimentary / early-
843 diagenetic precipitation from saline brines on a mudflat^{38,66-68}. We note that the
844 dolomite co-occurs in a specific stratigraphic interval with analcime, which typically
845 forms in saline brines⁶⁹, and with increased concentrations of hematite⁷⁰ indicative of
846 dry conditions⁷¹. This suggests a paleo-environmental control on the formation of
847 calcian dolomite, but we cannot rule out the role of diagenetic replacement.

848

849 *DeepMIP climate model simulations*

850

851 We use the DeepMIP climate model ensemble⁴⁰ to compare our proxy
852 reconstructions with simulated early Eocene precipitation and temperature

853 estimates. The paleolocation of the Xining Basin was reconstructed using the
854 hotspot reference frame ref⁷² at 55 Ma
855 (<https://www.odsn.de/odsn/services/paleomap/paleomap.html>). This resulted in a
856 paleolocation (102.5°N, 39.5°E) which fits well in relation to paleotopographic
857 features such as the Hangay Dome and Qinling belt that are observed on the
858 paleogeographic map used in the DeepMIP simulations⁷³.

859

860 **Data availability**

861 Palynology counts, TOC and bulk organic $\delta^{13}\text{C}$, *n*-alkane $\delta^{13}\text{C}$ and $\delta^2\text{H}$, XRD bulk
862 mineralogy and carbonate $\delta^{13}\text{C}$, $\delta^{18}\text{O}$ and Δ_{47} are available at
863 <https://doi.org/10.1594/PANGAEA.962894>. The DeepMIP preindustrial and Eocene
864 climate model simulations are available by following the instructions at
865 <https://www.deepmip.org/data-eocene/>. The weather station data from Xining and
866 Tashkent is available at <https://www.ncei.noaa.gov/products/wmo-climate-normals>.

867

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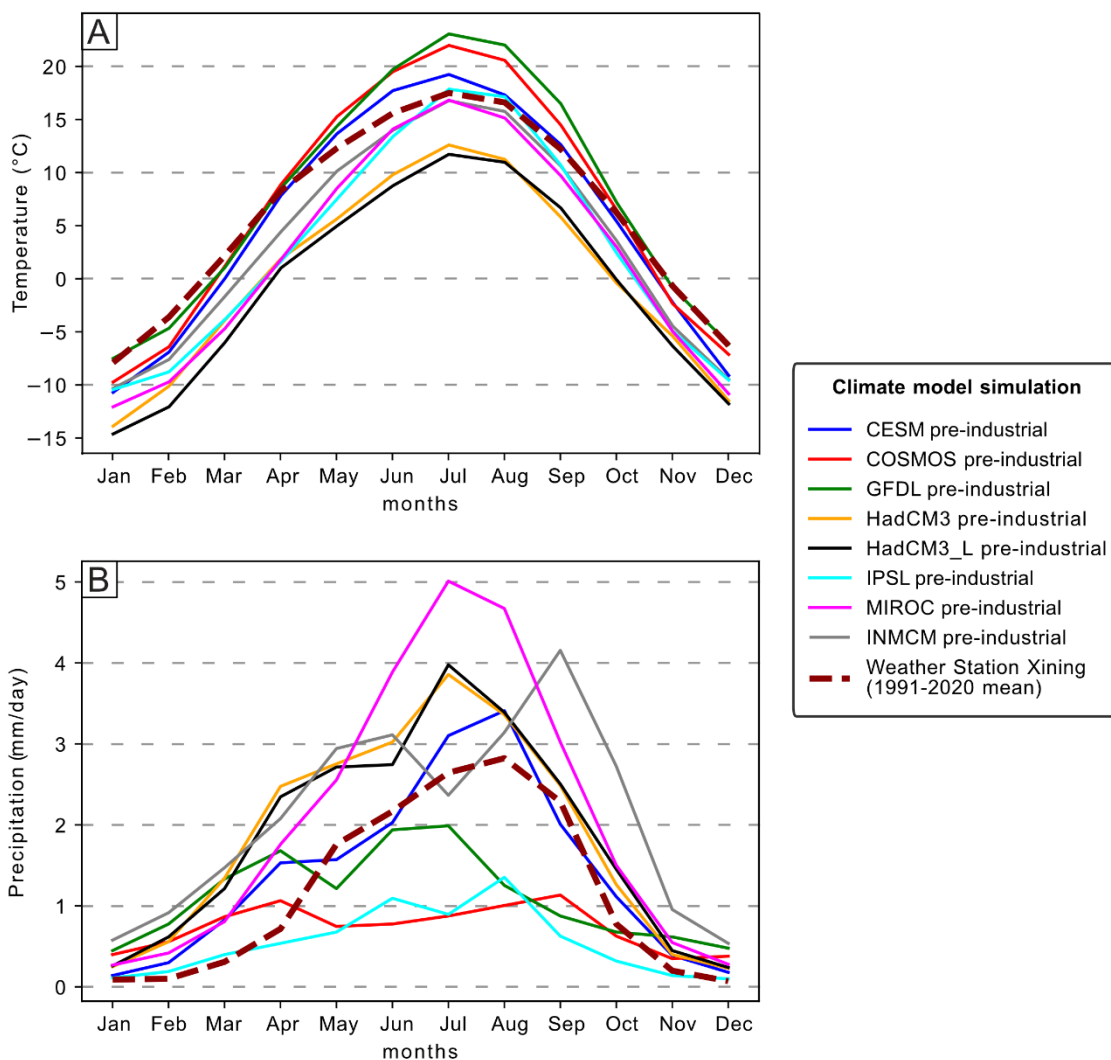
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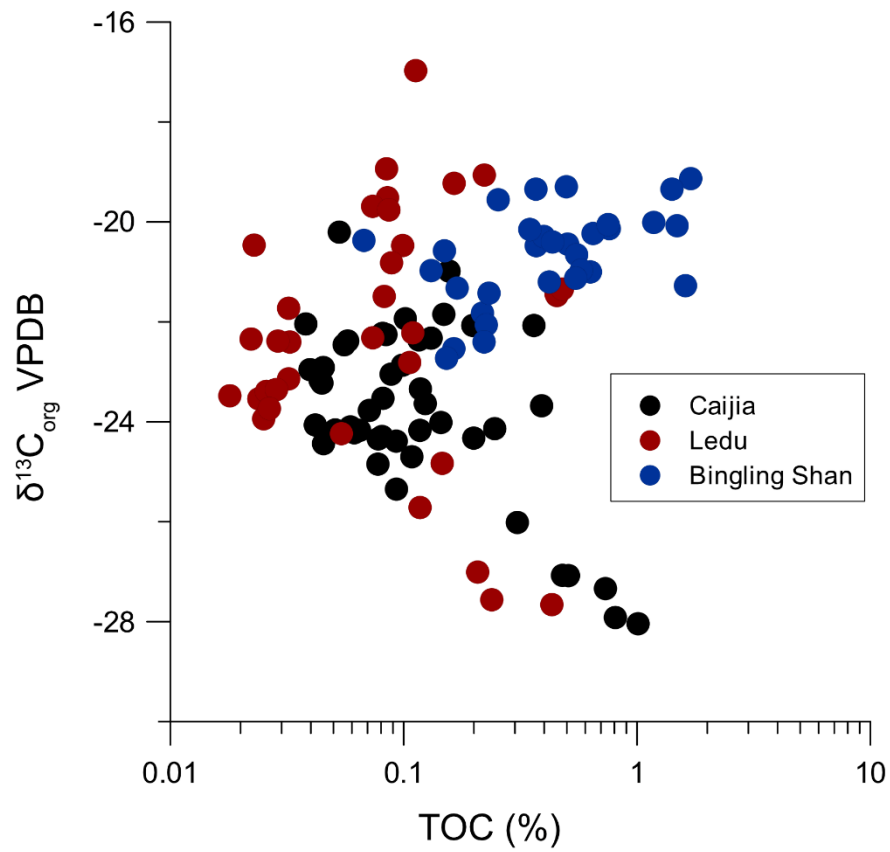
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956
 957 **Extended Data**

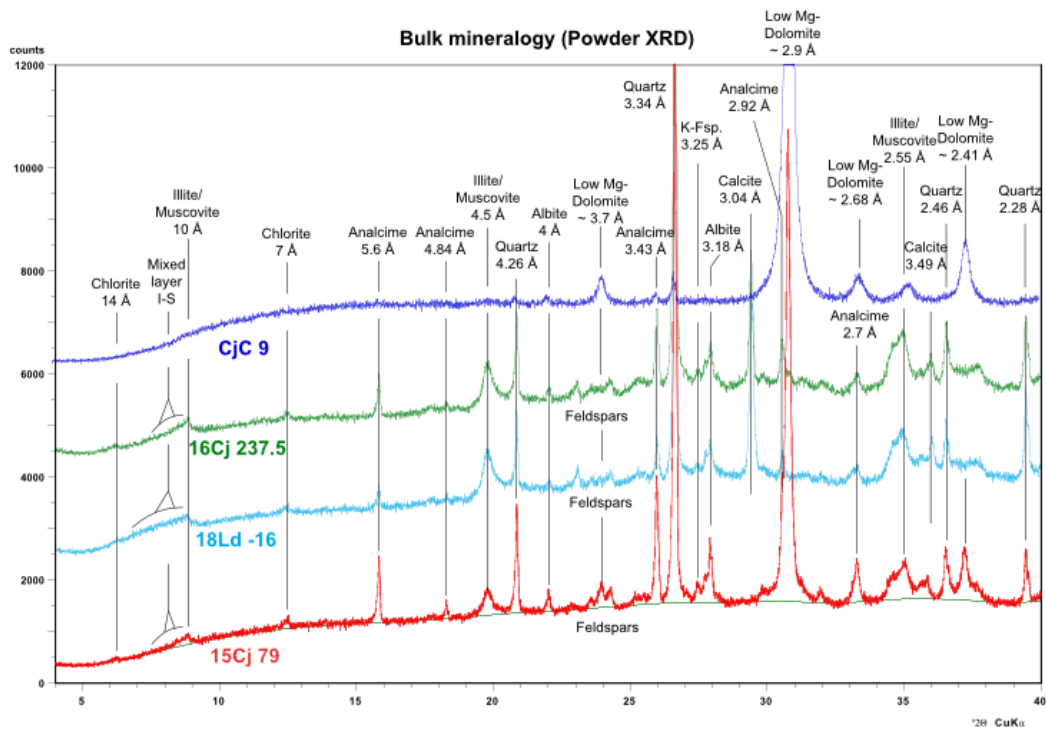


958
 959 *Extended Data Fig. 1. Pre-industrial control simulations and weather station data from*
 960 *the Xining Basin. Weather station precipitation data*
 961 *(<https://www.ncei.noaa.gov/products/wmo-climate-normals>) and pre-industrial control*

962 simulations of A) temperature and B) precipitation in the Xining Basin using the
963 DeepMIP climate model ensemble.

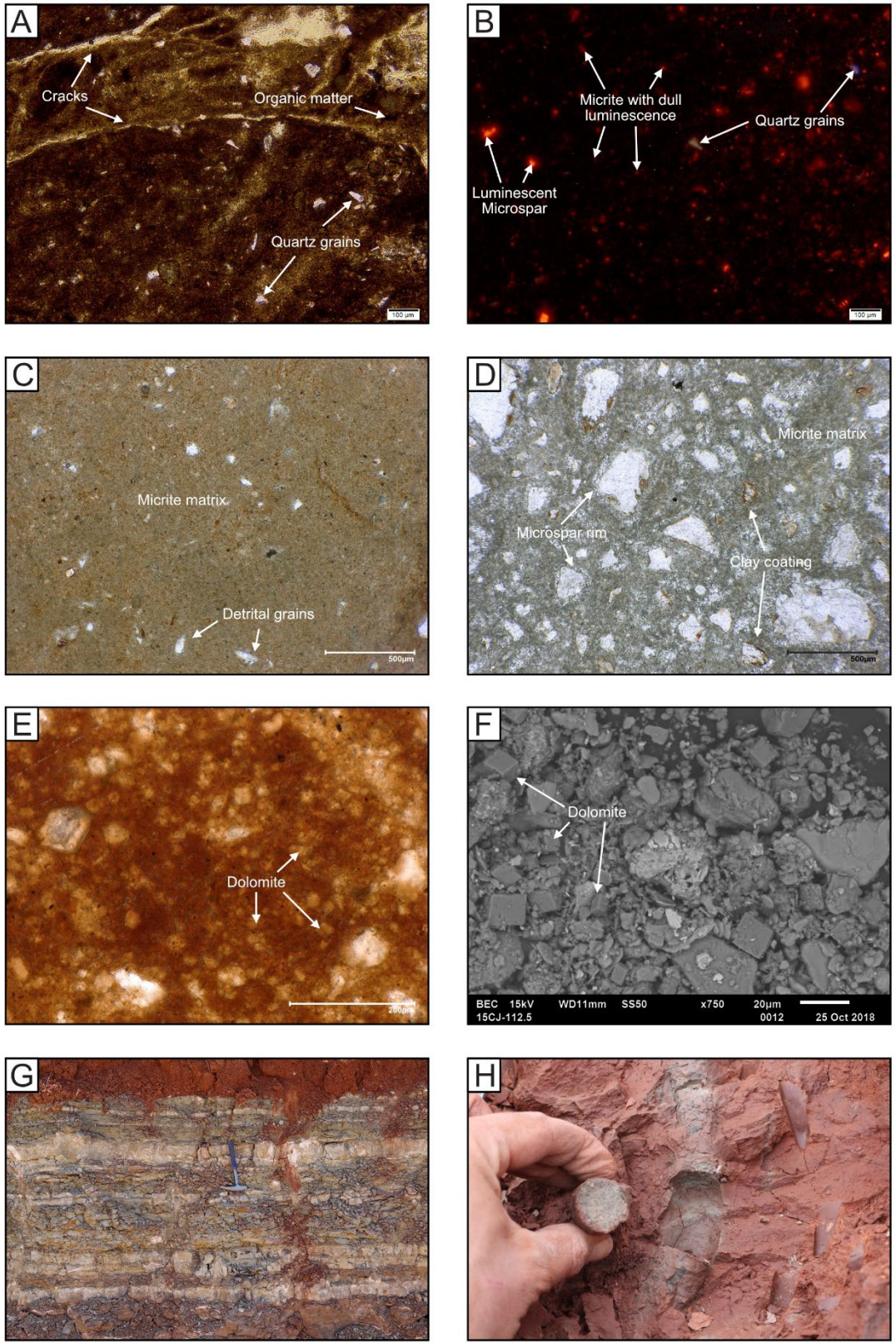


965 Extended Data Fig. 2. Soil organic matter $\delta^{13}\text{C}_{\text{org}}$ versus Total Organic Content (TOC).



966

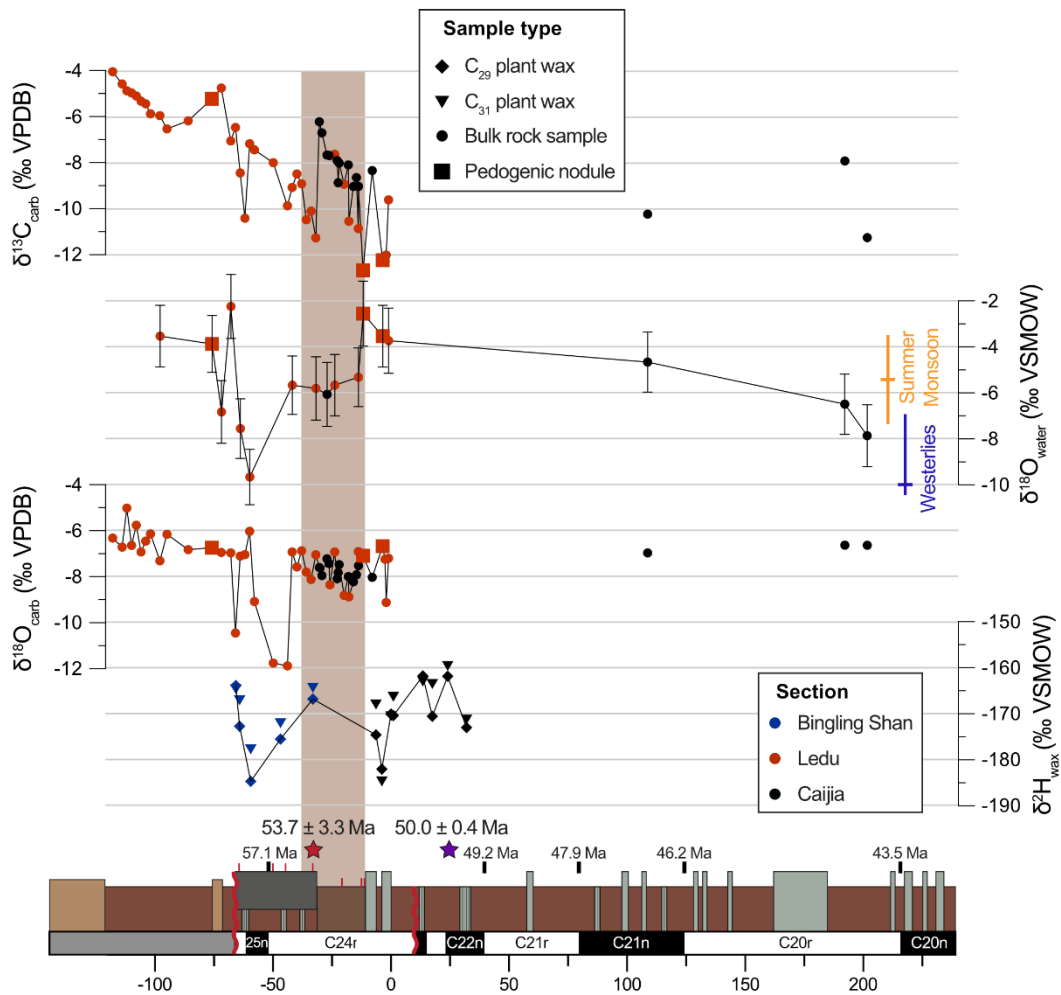
967 *Extended Data Fig. 3. XRD diffractograms showing the dominant peaks and bulk*
 968 *mineralogy of selected samples including: a dolomite bed (CjC9, as shown on*
 969 *Extended Data Fig. 4G; composite position: -37.1 m), a calcareous mudrock*
 970 *(16Cj237.5; composite position: 193 m), a calcareous mudrock from the CIE interval*
 971 *(18Ld-16; composite position: -31m) and a dolomitic mudrock (15Cj79; composite*
 972 *position: 54 m). The samples 16Cj237.5 and 18Ld-16 were used for stable and*
 973 *clumped isotope analyses.*



974

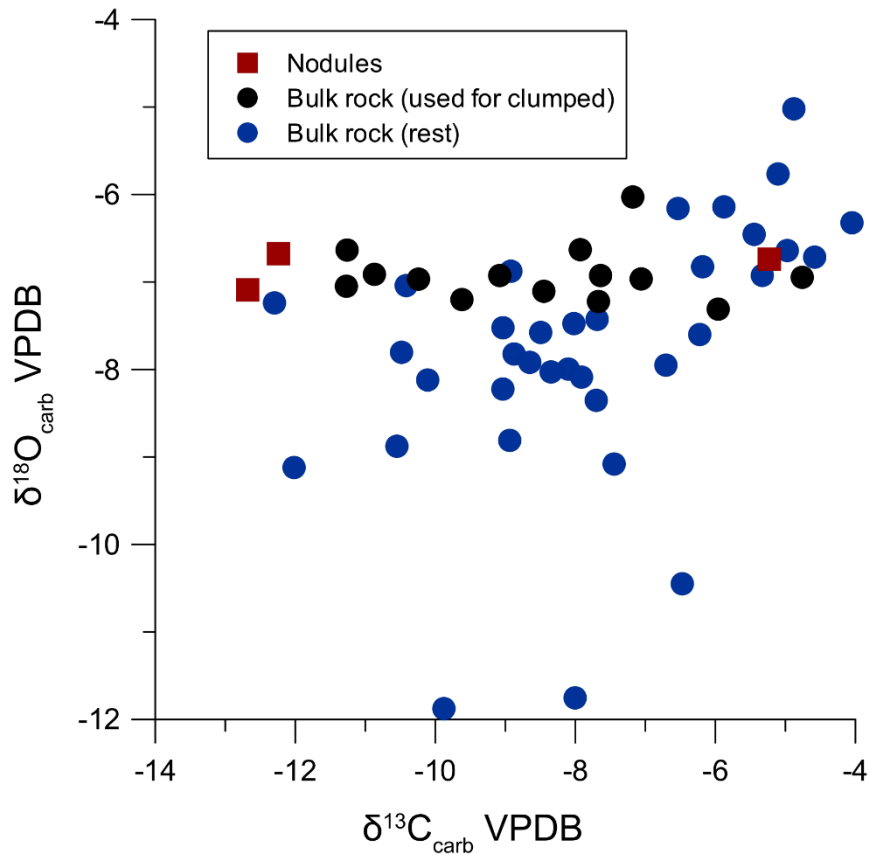
975 *Extended Data Fig. 4. Microscope and field photographs. A) and B) Cross-polarized*
 976 *and cathodoluminescence image of a mudrock sample with vadose-grown calcite*
 977 *cement (18LD-15; Ledu section; -15 meter-level) showing silt-sized matrix with*

978 dispersed organic matter, dull luminescent micrite and some luminescent microsparite
 979 and sparite. Plane-polarized images of pedogenic nodules showing: C) Micrite matrix
 980 with floating detrital grains in 18LD+4nodB (Ledu section; 12.3 meter-level). D)
 981 Floating detrital grains with isopachous microsparite to sparite rim and micritic matrix
 982 in 18LDconglonod (Ledu section; -60 meter-level). Massive red dolomitic mudrock
 983 samples showing rhombohedral dolomite crystals in E) plane-polarized light (15CJ-
 984 111_75; Caijia section 131.25 meter-level) and F) SEM (15CJ-112_5; Caijia section;
 985 132 meter-level). G) Greenish lacustrine mudrocks interbedded with massive dolomite
 986 beds (Caijia section; 75 meter-level), hammer for scale. H) Dolomite-filled burrow
 987 (Caijia section; 93 meter-level).

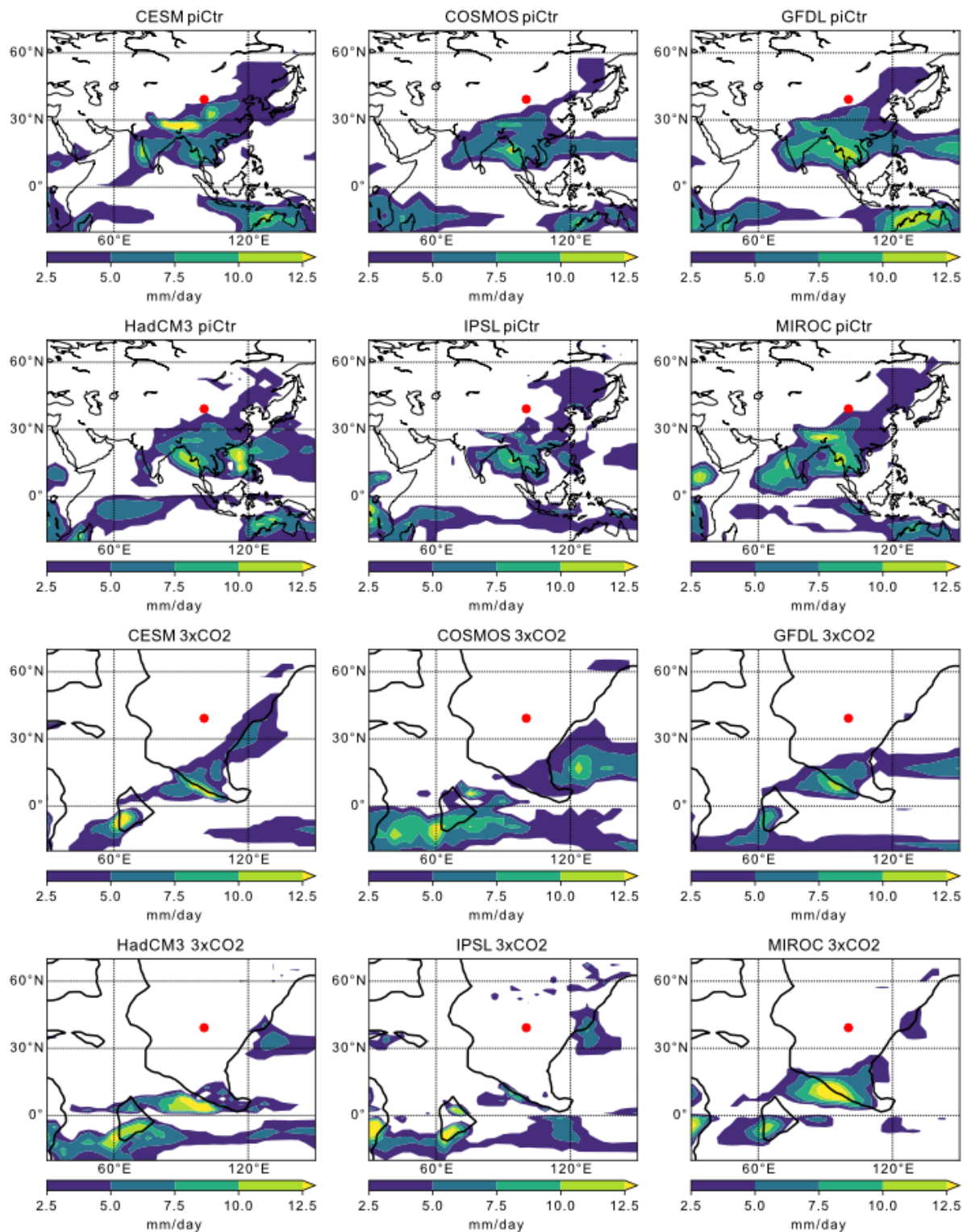


988

989 *Extended Data Fig. 5. Carbonate and plant wax stable isotopes. Stable isotope record*
990 *($\delta^{18}\text{O}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{carb}}$) of the vadose-grown and pedogenic carbonates and $\delta^2\text{H}_{\text{wax}}$*
991 *values of plant waxes. $\delta^{18}\text{O}$ values of the soil water ($\delta^{18}\text{O}_{\text{water}}$) are reconstructed using*
992 *the Δ_{47} -derived temperatures. Modern-day $\delta^{18}\text{O}_{\text{water}}$ values of the summer monsoon*
993 *and wintertime westerlies are shown⁵⁷.*



994
995 *Extended Data Fig. 6. Carbonate stable isotope biplot. Stable isotope biplot of the*
996 *studied calcites.*



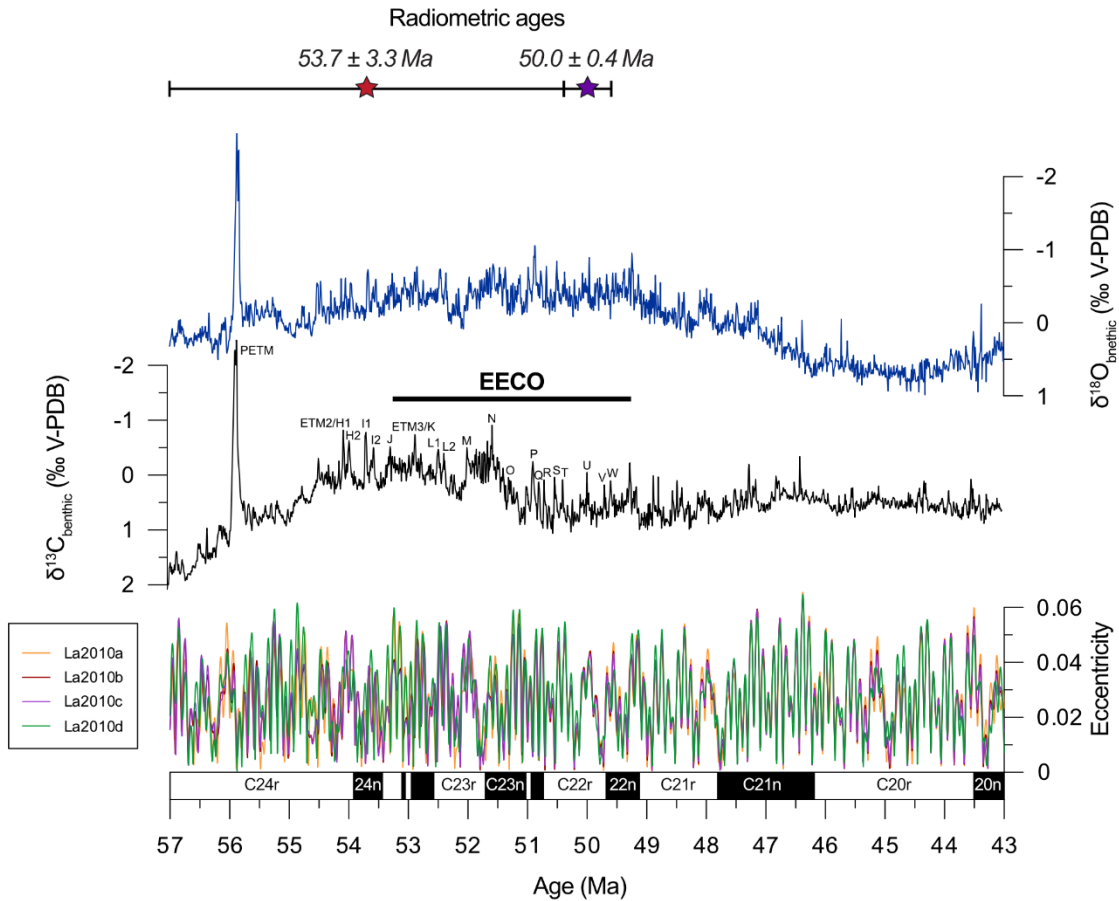
997

998 *Extended Data Fig. 7. Monsoonal domains in DeepMIP climate model simulations.*

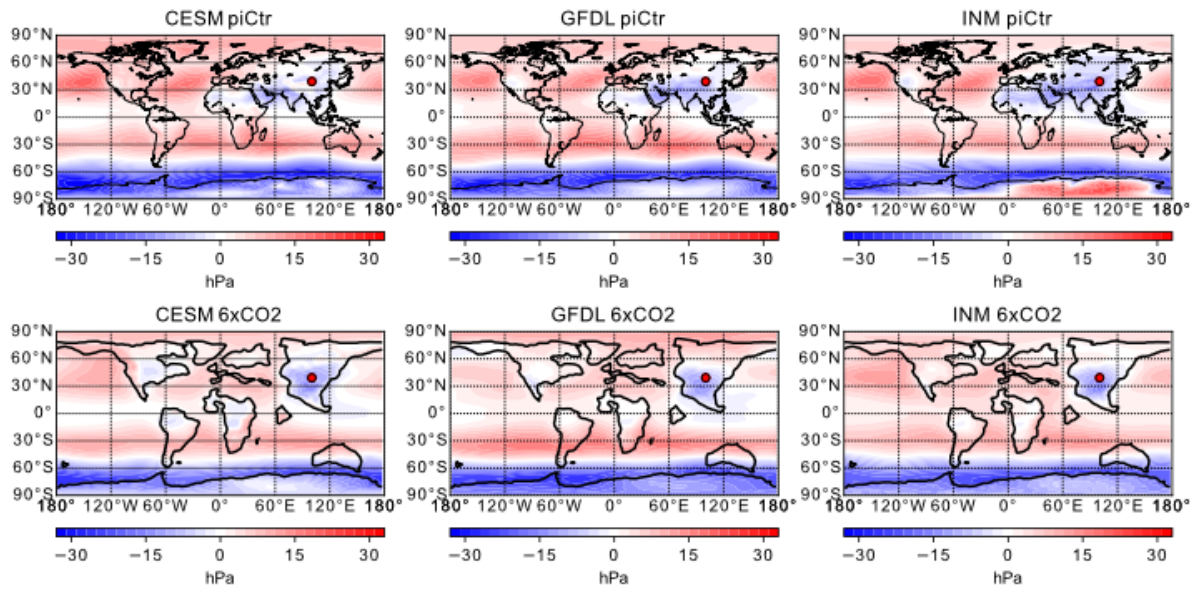
999 *Distribution of monsoonal domains in the pre-industrial control and 3x pCO₂ Eocene*

1000 *DeepMIP simulations. Summer (MJJAS) minus winter precipitation (NDJFM) is used*

1001 as a metric and regions where this difference is > 2.5 mm/day are considered
 1002 monsoonal⁴⁴. Red dot indicates the location of the Xining Basin.



1003
 1004 *Extended Data Fig. 8. Timing of Xining CIE compared with CENOGRID and*
 1005 *astronomical solutions. CENOGRID showing the evolution of benthic foraminifera*
 1006 *stable isotope values⁴. The $\delta^{18}\text{O}$ values show peak warmth of the EECO and the $\delta^{13}\text{C}$*
 1007 *values show several hyperthermal events characterized by negative CIE's. The*
 1008 *astronomical solutions⁷⁶ show the evolution of the eccentricity of Earth's orbit. The*
 1009 *radiometric ages bracketing the CIE in the Xining Basin are indicated.*



1010

1011 *Extended Data Fig. 9. Summer sea-level atmospheric pressure anomalies in DeepMIP*
 1012 *climate model simulations. Sea-level atmospheric pressure anomalies during summer*
 1013 *(JJA) in the 6x pCO₂ Eocene and the pre-industrial control simulations of the DeepMIP*
 1014 *ensemble. Red dot indicates the location of the Xining Basin.*

1015