

Supplementary Information *for*

Climate–carbon-cycle interactions and spatial heterogeneity of the Late Triassic Carnian Pluvial Episode

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1 **Text S1. Names and key references for the numbered locations of the CPE**
2 **records in Fig. 1**

3 1: Kapp Toscana Group, central Spitsbergen, Norway¹; 2: Gråklint Beds, Jameson
4 Land Basin, Greenland²; 3: Stuttgart Formation, central European Basin, Germany
5 and Poland (ref.³, but see ref.⁴); 4: Mercia Mudstone Group, Wessex Basin, United
6 Kingdom⁵; 5: Stockton Formation, Fundy Basin, Canada⁶; 6: Doswell Formation,
7 Richmond and Taylorsville basins, United States⁷; 7: Timezgadouine Formation,
8 Essaouira Basin, northwest Africa⁸; 8: Manuel Formation, Iberia, Spain and Portugal⁶;
9 9: Raibl Group, Southern Alps, Italy⁹; 10: Lunz Formation, Northern Calcareous Alps,
10 Austria¹⁰; 11: Veszprém Marl and Sándorhegy formations, Transdanubian Range,
11 Hungary¹¹; 12: Kartoz and Kasımlar Formations, Anatolian terrane, Turkey¹²; 13:
12 Mohilla Formation, Levant Basin, Israel¹³; 14: Hamrat Duru Group, Oman
13 Mountains, Oman¹⁴; 15: Santa Maria Formation, Chaco–Paraná Basin, Brasil¹⁵; 16:
14 Los Rastros Formation, Ishigualasto and Cujo basins, Argentina¹⁶; 17: Molteno
15 Formation, Karoo Basin, South Africa¹⁷; 18: Tiki Formation, Rewa Basin, India¹⁸; 19:
16 Chomule and Rama formations, Spiti, India¹⁹; 20: Mungaroo Formation, Carnarvon
17 Basin, Australia²⁰; 21: Bagong Formation, Tibet Plateau, China²¹; 22: Tanzhuang and
18 Anyao formations, Jiyuan Basin, China²²; 23: Zhuganpo and Wayao formations,
19 Nanpanjiang Basin, China²³; 24: Ma'antang Formation, Sichuan Basin, China²⁴; 25:
20 TR5A zone, Inuyama area, central Japan²⁵; 26: Miankuhi Formation, Turan Plate,
21 Iran^{26,27}.

22

Text S2. U-Pb dating

The samples were crushed and separated for zircon selection using standard sieving and magnetic and heavy-liquid separation techniques. A total of 150 inclusion-free zircon grains (80–200 μm) from each sample were then hand-selected under a binocular microscope and mounted in epoxy resin. Hardened mounts were polished to expose zircon grain mid-sections at approximately 2/3rds to 1/2 of their width. Cathodoluminescent (CL) imaging was used to document grain morphologies and internal structure for *in situ* analysis. U-Pb isotopic data from zircons were obtained at the Department of Earth Sciences, The University of Hong Kong, using a Nu Instruments Multiple Collector (MC) inductively coupled plasma mass spectrometer (ICP-MS) with a Resonetics RESolution M-50-HR excimer laser ablation system. The analyses used a beam diameter of 30 μm , repetition rate of 4 Hz, and energy density of 5 J/cm² on the sample surface. The average ablation time was approximately 40 s, and pit depths reached approximately 30 μm . Harvard reference zircon 91500 ($1065.4 \pm 0.3 \text{ Ma}$)²⁸ and zircon Plešovice ($337.1 \pm 0.4 \text{ Ma}$)²⁹ were used for preliminary calibration and as a second reference, respectively. The zircon 91500 was used as an external calibration sample to evaluate the magnitude of mass bias and inter-elemental fractionation. The zircon Plešovice was used to evaluate the accuracy and precision of the laser-ablation results. ICPMSDataCal³⁰ software was used for off-line signal selection, quantitative calibration and time-drift correction. We used a function given in Anderson (2002)³¹ to correct for common Pb in Microsoft Excel. Isoplot v. 3.0³² was used to construct concordia diagrams and probability density plots. Within the overall detrital age distribution, we cite ²⁰⁶Pb/²³⁸U ages for zircon grains younger than 1000 Ma and ²⁰⁷Pb/²⁰⁶Pb ages for older grains. In this study, zircon grains were randomly selected from each sample, so the results are expected to reflect the characteristics of the age populations. Ages with a concordance degree of $\geq 95\%$ (sandstone) were selected for the analysis.



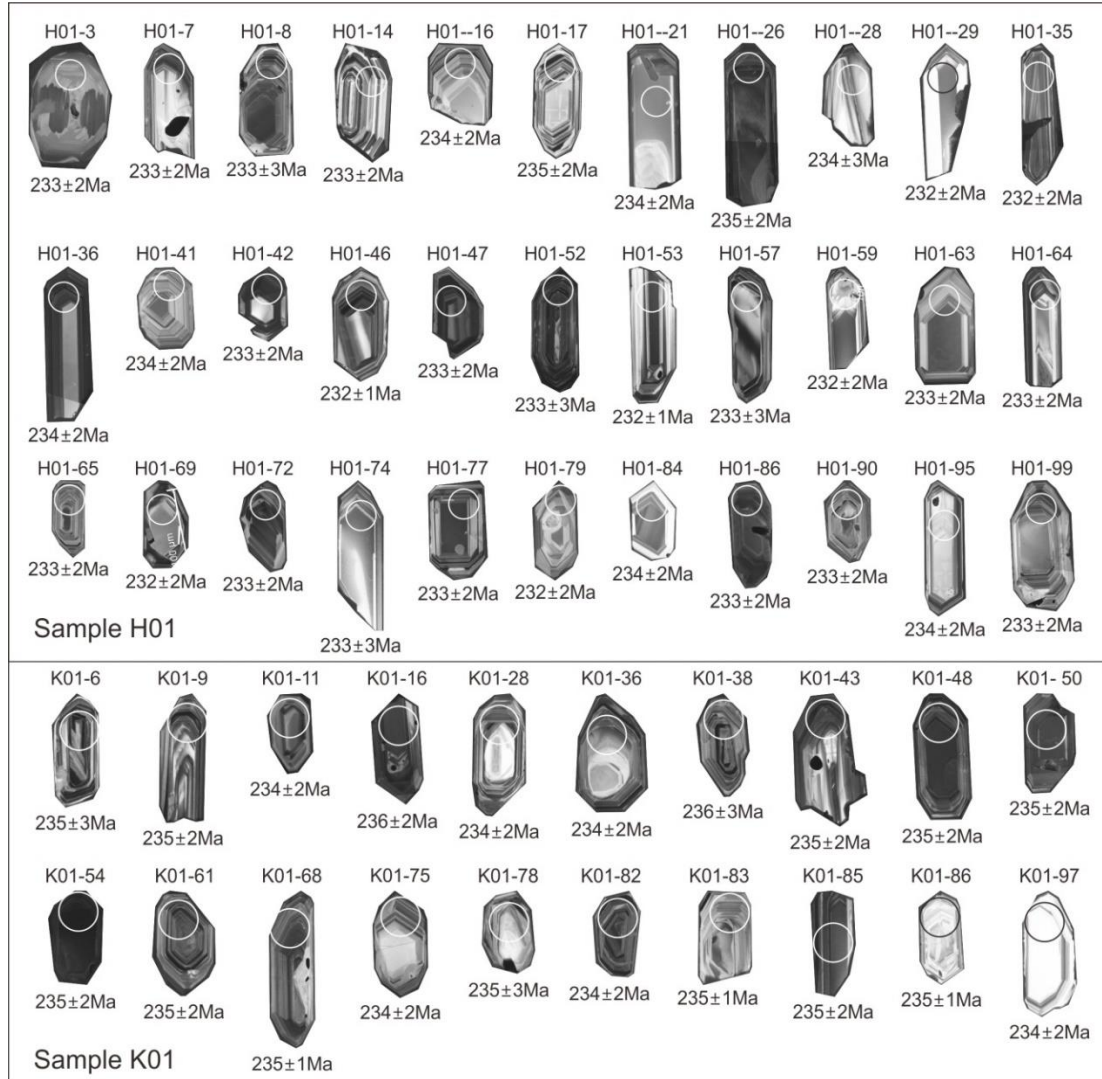
Fig. S1. Photographs showing representative lithologies and collected samples for U-Pb dating in the Dalongkou section. **a**, Photograph of the Dalongkou section shows the locations of the two U-Pb dating samples (K01 and H01). **b**, Grey-green sandstone (K01) from the Karamay Formation. **c**, Grey-green sandstone (H01) from the

Huangshanjie Formation. **d**, Interbedded mudstone and marl in the middle part of the Huangshanjie Formation. **e**, Mudstone and black shale in the CPE interval. **f**. Petrographic photograph of the dating sample K01. **g**. Petrographic photograph of the dating sample H01. Abbreviations: Qz, Quartz; Fs, Feldspar; Bt, Biotite.

Two grey-green sandstone samples were collected for LA-MC-ICP-MS U-Pb dating, separately from the top Karamay Formation and lower Huangshanjie Formation (Fig. S1). The sizes of zircon grains separated from two samples range from 80 to 120 μm (Fig. S2). All grains exhibited euhedral morphologies and oscillatory zoning patterns, indicating an igneous origin. The angular grain facets indicated minimal abrasion during sedimentary transport and support the interpretation that the source material was extruded near the depocentre of the tuff. For sample H01, eighty-four of 100 analyses (concordance $\geq 95\%$) provided concordant ages ranging from 871 to 232 Ma (Fig. S3a). The major subpopulation of ages generally fell between 235 and 232 Ma (thirty-three ages) as the youngest age population, which was analysed to provide a weighted mean of 232.9 ± 0.7 Ma (MSWD = 0.2) representing the maximum depositional age of H01 (Fig. S3b). For sample K01, eighty-eight of 100 analyses (concordance $\geq 95\%$) provided concordant ages ranging from 859 to 234 Ma (Fig. S3c). The major subpopulation of ages generally fell between 236 and 334 Ma (twenty ages) as the youngest age population, which was analysed to provide a weighted mean of 234.8 ± 0.8 Ma (MSWD = 0.1) for the maximum depositional age for K01 (Fig. S3d).

According to the “law of detrital zircon”, the deposition of the strata in question cannot not be older than the age calculated from the youngest zircon grains from the youngest magmatic events^{33–36}. Therefore, the maximum depositional age (MDE) can be close to the real depositional age, especially when the former is consistent with the biostratigraphical age^{37–40}. In this study, the youngest zircons are morphologically uniform with no visible signs of rounding due to even minor sedimentary abrasion during transportation, a conclusion also supported by the angular shapes of quartz grains (Fig. S1f, g). These observations indicate that the depocenter of two sandstone

84 samples from the Dalongkou section was situated near the magmatic source of the
 85 zircons. Considering that the maximum depositional ages of two samples are
 86 consistent with the age analyzed from the sporo-pollens, we use the maximum
 87 depositional ages to tightly constrain the age of the CPE in this study.



88
 89 Fig. S2. Cathodoluminescent (CL) images of analysed zircon grains.

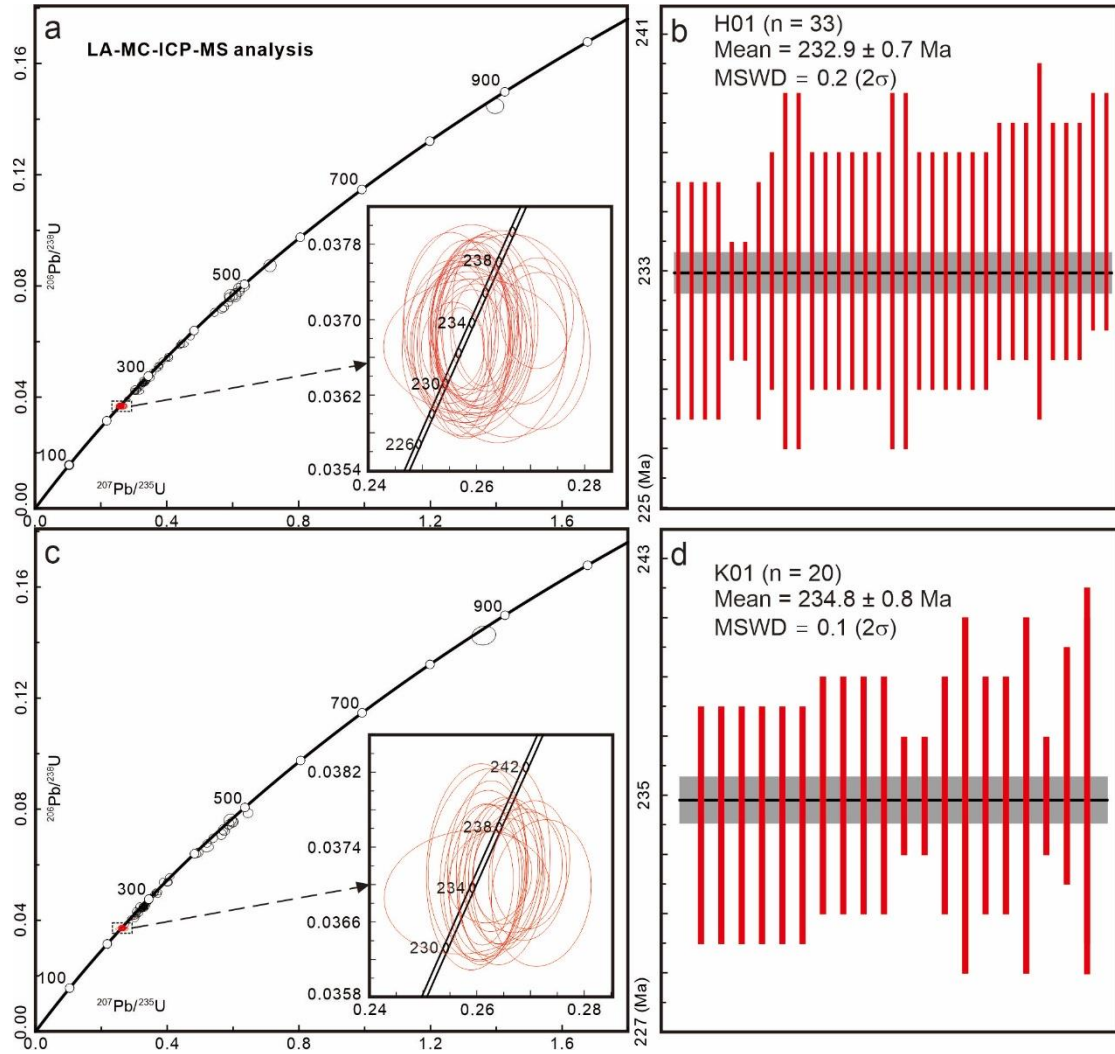


Fig. S3. U-Pb geochronology of samples (H01 and K01) from the Dalongkou section. Left: U-Pb concordia plots for the youngest zircons. Right: rank order plots for the youngest zircons, showing the weighted means of $^{206}\text{Pb}/^{238}\text{U}$ dates (uncertainties are given at the 2σ level). MSWD, mean square of weighted deviates. All data are provided in Supplementary Data 1.

Text S3. Astronomical cycle

(1) Spectral analysis

The magnetic susceptibility (MS) series (1095 samples, interpolated with a main sampling interval of 0.25 m, Supplementary Data 2) spans 292.7 m (estimated to represent ~3.922 Myr) from the bottom of the Huangshanjie Formation. Spectral analysis of the MS series indicates a series of obvious sedimentary wavelengths (Fig. S4, Table S1), which are approximately in accord with the ratios of the 405-kyr-long eccentricity, 173-kyr obliquity, 131–95-kyr short-eccentricity, ~33-kyr obliquity and ~21–17-kyr precession cycles around 234 Ma, based on the La2004 astronomical solution⁴¹.

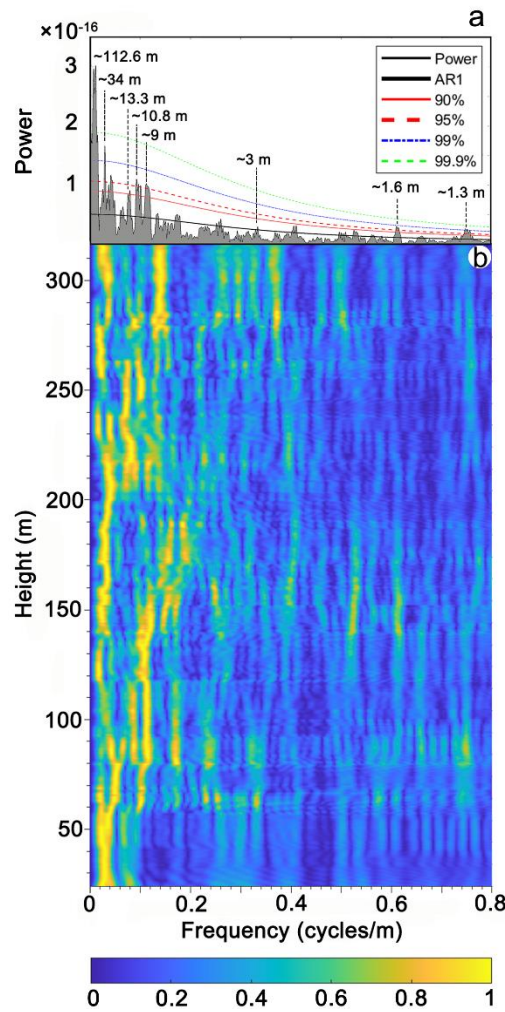


Fig. S4. 2π -MTM and evolutionary FFT of the detrended MS series. The sliding

window and step of evolutionary FFT are set to 60 m and 0.5 m, respectively.

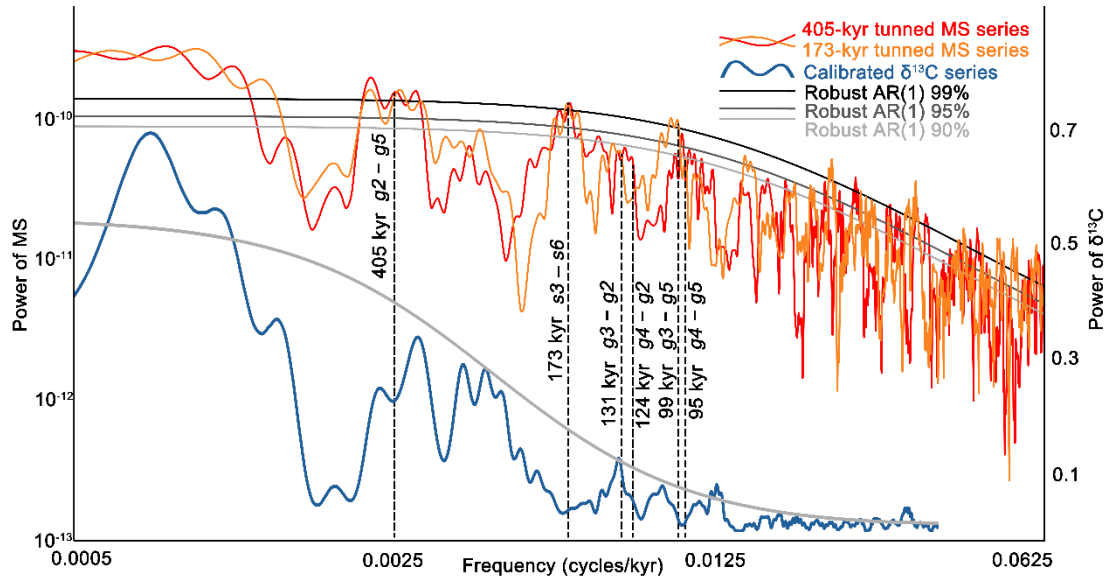


Fig. S5. 2π -MTM of the tuned MS series (405-kyr and 173-kyr tuning schemes) and calibrated $\delta^{13}\text{C}_{\text{org}}$ series.

Table S1: Significant sedimentary wavelengths in Spectral analysis of the MS series.

Sedimentary wavelengths	Ratios of the sedimentary wavelengths						Significance levels
~34 m	26.2	21.3	11.3	3.8	3.1	2.6	99%
~13.3 m	10.2	8.3	4.4	1.5	1.2	1	90%
~10.8 m	8.3	6.8	3.6	1.2	1		95%
~9 m	6.9	5.6	3	1			95%
~3 m	2.3	1.9	1				AR1
~1.6 m	1.2	1					95%
~1.3 m	1						99%

(2) Sedimentation rate analysis

The correlation coefficient analysis (COCO/eCOCO) is a quantitative technique for estimating the potential variations in sedimentation rate through the time series by evaluating the correlation coefficient between a power spectrum of a theoretical orbital forcing target and palaeoclimate proxies, the number of contributing orbital

parameters also being considered⁴². The null hypothesis of no significant astronomical frequencies is tested by Monte Carlo simulation⁴². We adopted 3000 times Monte Carlo simulations in this analysis. The optimal sedimentation rate suggested by COCO/eCOCO and null hypothesis analysis varies between 5 and 10 cm/kyr, which is highly consistent with the sedimentation rate based on the 405-kyr tuning scheme (Fig. S6).

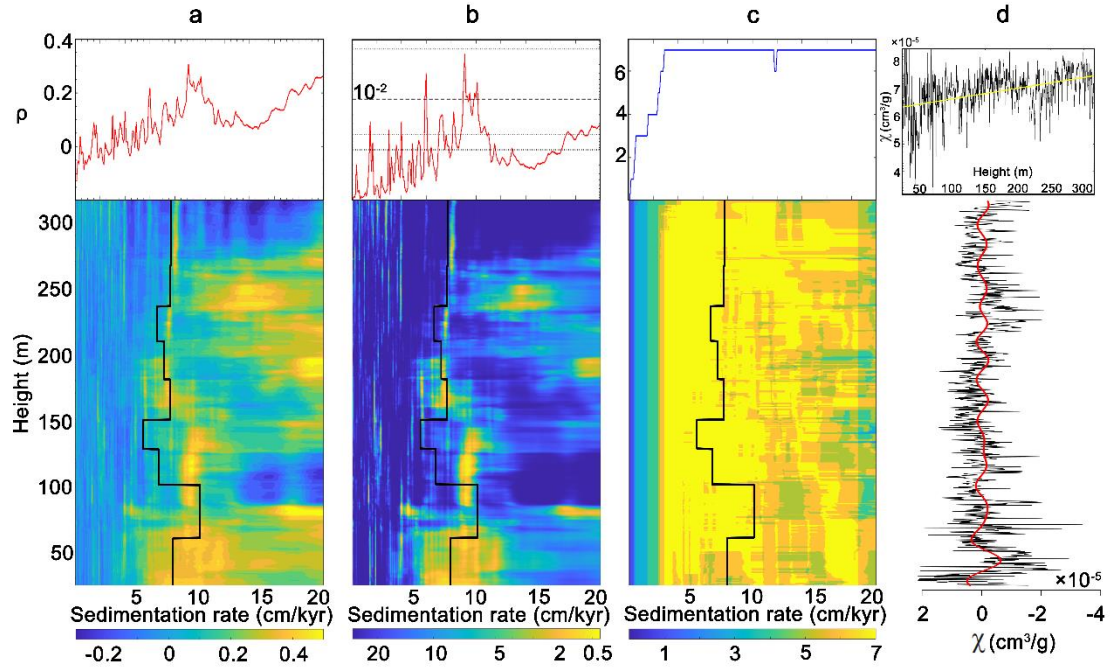


Fig. S6. Sedimentation rate analysis of the detrended MS series. **a**, Correlation coefficient. **b**, Null hypothesis (H0) significance level (c) No. of contributing orbital parameters. The sliding window and step of the evolution plots in (a), (b), and (c) are set as 75 m and 0.5 m. **d**, MS series and its linear trend (yellow line), detrended MS series and Gaussian filtered wavelengths of 50–25 m (0.02–0.04 cycles/m) (red curve). The black line in (a), (b), and (c) is the sedimentation rate based on the 405-kyr tuning scheme.

(3) Phase relation between the magnetic susceptibility and $\delta^{13}\text{C}_{\text{org}}$

On the scale of 405-kyr long eccentricity, the detrended MS series (tuned to the 405-kyr pure sinusoid, interpolated to a uniform sampling interval of 3.2661 kyr, removed 35% LOESS trend) and $\delta^{13}\text{C}_{\text{org}}$ series (interpolated to the uniform sampling interval

refer to the MS series, clipped from -25.5% , removed 35% LOESS trend) show high coherence (>0.6) and in-phase relationship ($\sim 0^\circ$ lag) (Fig. S7).

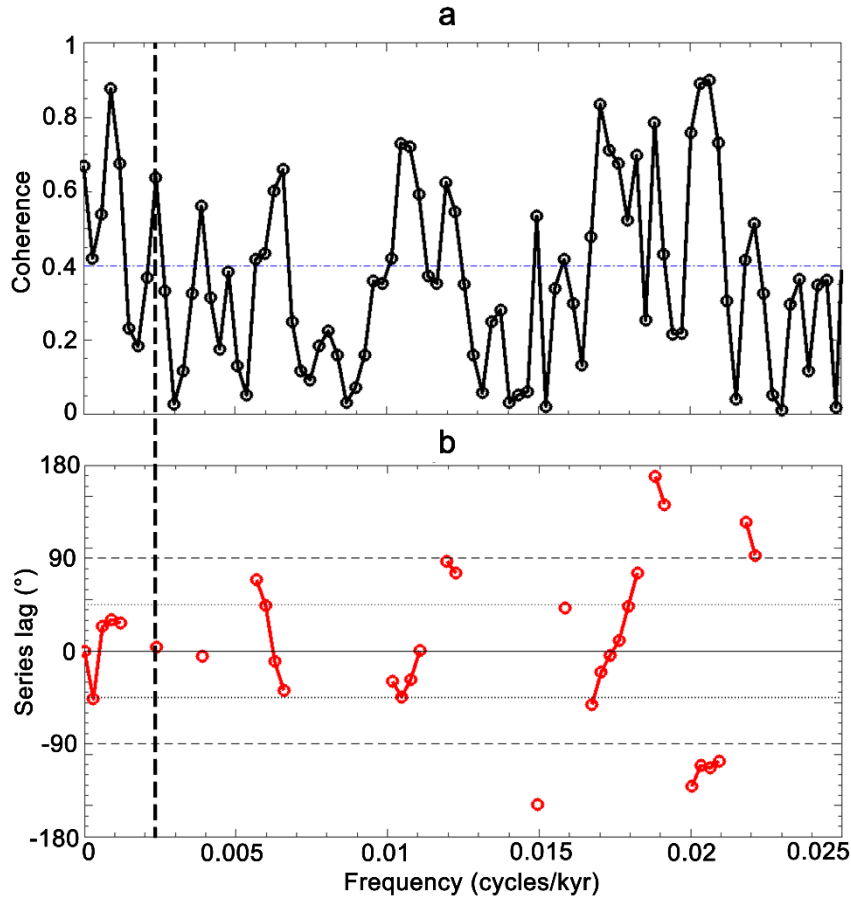


Fig. S7. **a**, Magnitude-Squared Coherence. **b**, Cross-Spectrum phase of the detrended MS series tuned to 405-kyr pure sinusoid and calibrated $\delta^{13}\text{C}_{\text{org}}$ series. Window size is 1900 kyr, overlap is 950 kyr. The black dotted line in (a) and (b) is the frequency of the 405-kyr long eccentricity period.

(4) Threshold response model of orbital forcing climate changes

The response of climate changes driven by orbital cycles exhibits non-linear and complex characteristics throughout geological history, which is attributable to the chaotic nature of Earth's climate system^{43,44}. This behaviour might lead to a threshold within the Earth's surface response system, limiting the influence of orbital forcing signals and further amplifying amplitude modulation of shorter cycles in proxy datasets^{42,45}. For example, the original precession cycles based on the La2004

astronomical solution of 0–3 Ma and the clipped precession series exhibit similar frequency components of precession, but the long and short eccentricity components (the amplitude envelope of precession) are enhanced in the clipped series (Fig. S8).

In our study, the 405-kyr long-eccentricity signal is enhanced in the lacustrine $\delta^{13}\text{C}_{\text{org}}$ during the CPE, which can be explained by crossing of a geochemical threshold in response to the eccentricity forcing. During hyperthermal intervals, the fluctuations of lacustrine $\delta^{13}\text{C}_{\text{org}}$ at the scale of $\pm 1\%$ are mainly influenced by the variation in the terrestrial carbon pool⁴⁴. The geochemical response of intense seasonal differences during high- CO_2 intervals is capable of clipping the eccentricity signal. When the eccentricity varies below the threshold, the fluctuations induced by this parameter are insufficient to surpass the geochemical threshold, promoting relatively stable climate conditions in a relatively warm background, which facilitated the expansion of continental carbon reservoirs, thereby enhancing the land storage of isotopically light carbon^{46,47}. Conversely, as eccentricity exceeds the critical threshold, more light carbon was released from organic matter into the atmosphere. Additionally, intervals of eccentricity maxima may impose greater stress on terrestrial ecosystems, resulting in diminished isotopic fractionation in terrestrial plants^{46,47}. Due to the complexity of the processes involved, the threshold within the Earth's surface response system might not be stable. Our study indicates that the concentration of CO_2 may be a crucial factor in determining the threshold response model.

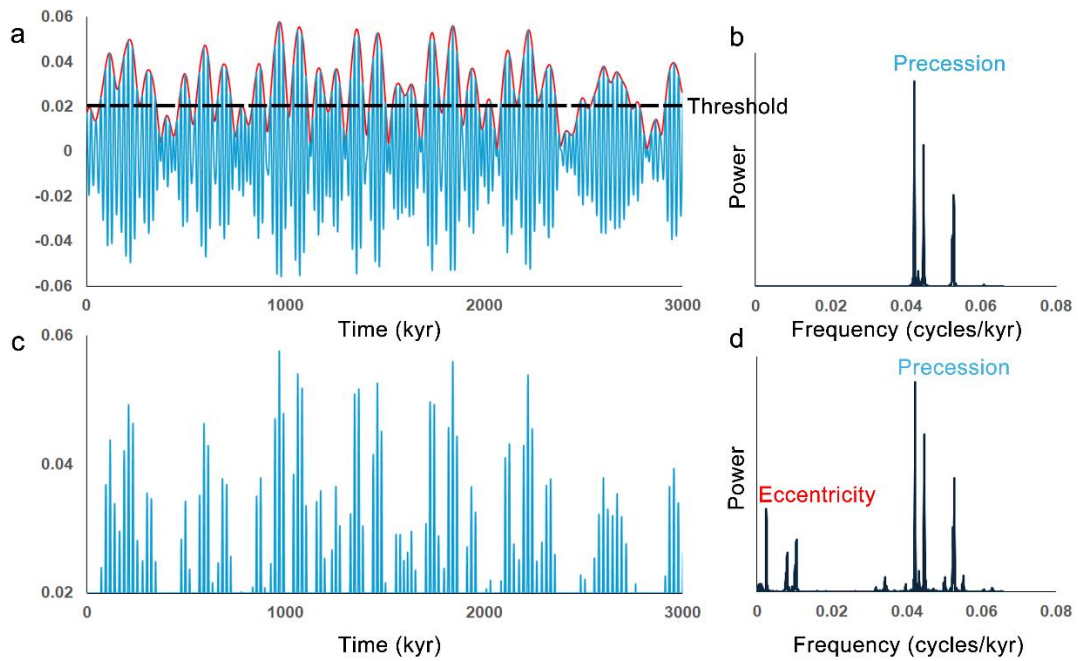


Fig. S8. A simplified threshold response model of eccentricity forcing. **a**, Theoretical La2004 astronomical solution of eccentricity (red) and precession (blue) from 0 to 3 Ma. **b**, Periodicity spectrum of theoretical precession signal. **c**, Cartoon model for seasonal variations in low-CO₂ background. **d**, Precession signal clipped from the threshold.

Text S4. Carbon sources of the DLK section

(1) Carbon-isotope stratigraphy

The high-resolution bulk organic carbon-isotope ($\delta^{13}\text{C}_{\text{org}}$) records span 316.20 m, covering the uppermost part of the Karamay Formation and the lower–middle part of the Huangshanjie Formation (approximately 234–230 Ma, Fig. S9).

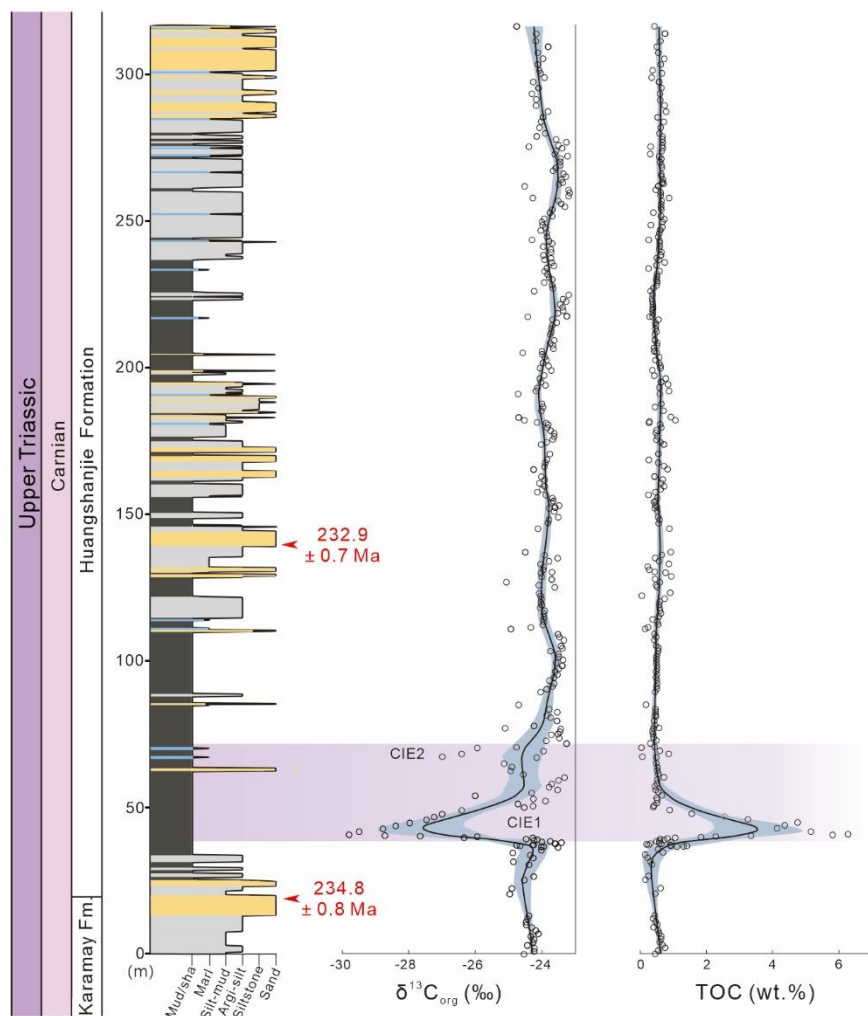


Fig. S9. Bulk organic carbon-isotope ($\delta^{13}\text{C}_{\text{org}}$) and total organic-carbon (TOC) contents of the Dalongkou section. Abbreviations: Mud, Mudstone; Sha, Shale; Silt-mud, Silty mudstone; Argi-silt, Argillaceous siltstone; Sand, Sandstone.

(2) Rock Eval

Rock Eval pyrolysis generates hydrogen index (HI, mg hydrocarbons/g TOC), oxygen index (OI, mg CO_2 /g TOC), and T_{max} ($^{\circ}\text{C}$) data, which can be useful for understanding

the type and maturity of organic matter⁴⁸. Sixteen bulk-rock samples spread through the DLK section were analysed at the Guangzhou Institute of Geochemistry, Chinese Academy of Sciences, using a Rock Eval 6 instrument, following the laboratory procedures from Espitalié et al. (1985)⁴⁸. The IFP 160000 standard was used to calibrate results.

The $T_{\max}$ values of the DLK section range from 404 to 519 °C, with an average value of 438 °C, indicating moderately low maturity. Low HI values (14 to 88 mg, mean 36 HC/g TOC) and moderately high OI values (55 to 276 mg, mean 22 mg CO₂/g TOC) characterize the background of the DLK section, likely indicating a mixture of terrestrially derived Type III and Type IV organic matter (Fig. S10). However, HI values are high and OI values are low through the CIE1 interval, representing Type II organic matter in the deposits, typically sourced from algae.

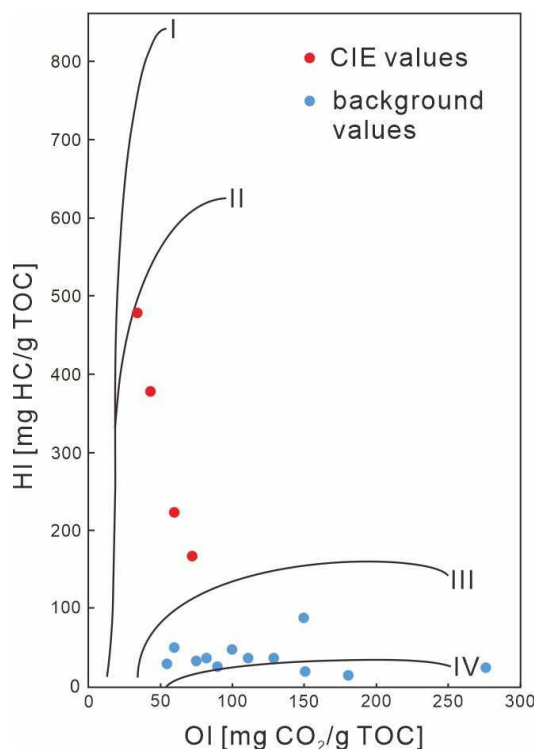


Fig. S10. Cross-plot (van Krevelen diagram) of HI versus OI from Rock Eval measurements of the Dalongkou samples.

(3) Kerogen macerals

Kerogen macerals include four groups (vitrinite, exinite, sapropelinite and inertinite) based on different organic-matter sources. Generally, the vitrinite, inertinite and exinite groups mainly originate from terrestrial higher plants, whereas the sapropelinite group is mainly formed from lacustrine plankton, predominantly algae⁴⁹. According to the China national standard (SY/T5125–2014), kerogen enrichment and identification of 10 samples through the DLK section were undertaken at the Yangtze University. At least 300 effective points per sample were analyzed by the point-counting method under a microscope.

The macerals of the DLK section are characterized by abundant components from the exinite group ($\bar{x} = 59.1\%$), followed by vitrinite ($\bar{x} = 18.8\%$), sapropelinite ($\bar{x} = 13.7\%$), and inertinite ($\bar{x} = 5.0\%$) groups, indicating that the organic matter was mainly derived from terrestrial higher plants (Supplementary Data 3; Fig. S11 and S12). However, an abrupt increase in abundance of the sapropelinite group (up to 29%, Fig. S12) at the CIEs interval suggests increased aquatic primary productivity during the CPE.

Dalongkou Section

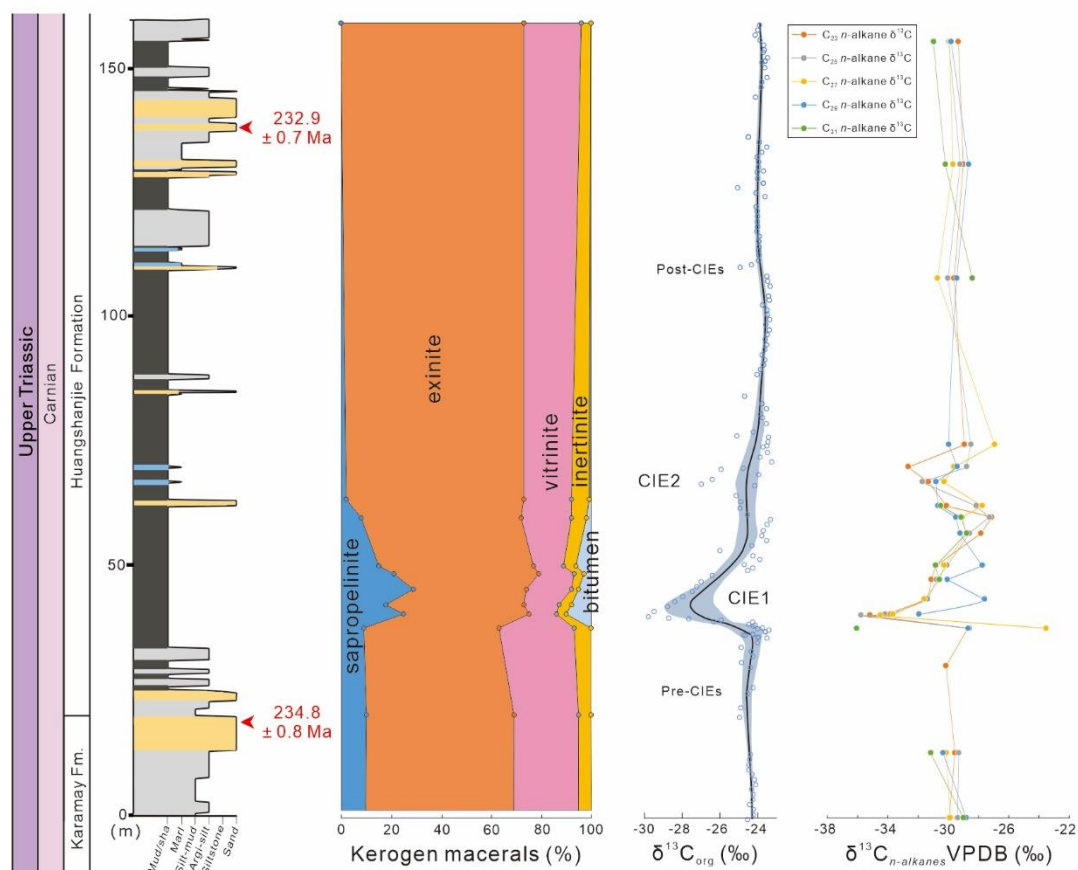


Fig. S11. Results of kerogen macerals and comparison of $\delta^{13}\text{C}$ signatures of terrestrial n -alkanes of the Dalongkou section. Abbreviations: Mud, Mudstone; Sha, Shale; Silt-mud, Silty mudstone; Argi-silt, Argillaceous siltstone; Sand, Sandstone.

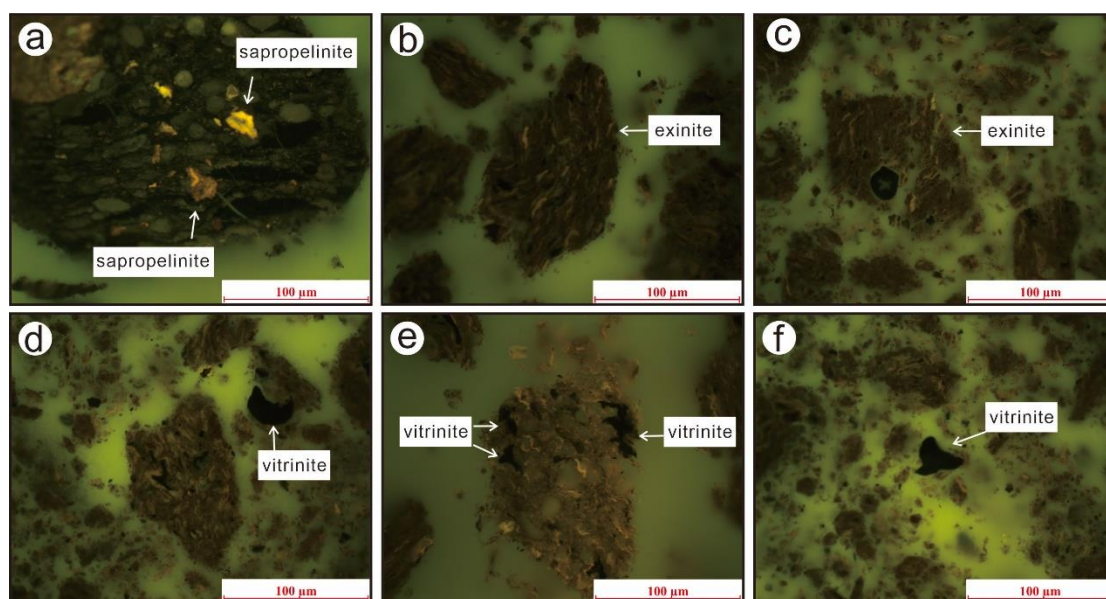


Fig. S12. Photomicrographs showing kerogen macerals from the DLK section.

(4) Compound-specific (long-chain *n*-alkane) $\delta^{13}\text{C}$ analysis

The long-chain *n*-alkanes derived from higher-plant leaf waxes display moderate odd-over-even carbon number predominance and are widely used as a proxy for palaeoenvironmental reconstruction^{50,51}. Here, we measured $\delta^{13}\text{C}_{n\text{-alkane}}$ and made comparisons with the bulk $\delta^{13}\text{C}_{\text{org}}$ values in order to evaluate the extent of diagenetic alteration and mixing effects⁵². The compound-specific carbon-isotope analysis was performed on eighteen samples taken from the DLK section. Approximately 40 g of powdered sample was extracted on a Dionex 350 Accelerated Solvent Extraction system using dichloromethane/methanol (93:9 v/v) with six extraction cycles to produce a total lipid extract (TLE). The TLE were deasphalted using a 40-fold excess *n*-hexane solution. After filtration, the soluble fraction was subjected to column chromatography on silica- Al_2O_3 and the saturated fraction was eluted with *n*-hexane. The saturated fractions were further separated into the linear and branched/cyclic fractions by urea adduction. The *n*-alkanes in the linear fractions were subjected to compound-specific isotopic analysis using a GV Isoprime Isotope Ratio Mass Spectrometer (IRMS) instrument interfaced to an Agilent 6890 GC via a combustion interface at GC5. Individual compounds separated by GC were quantitatively converted to CO_2 at 860 °C using CuO as the catalyst. A mixture of standards of *n*-alkanes (C_{12} , C_{14} , C_{16} , C_{18} , C_{20} , C_{22} , C_{25} , C_{28} , C_{30} , C_{32} and C_{35} from A. Schimmelmann of Indiana University) was measured once or twice daily to monitor the accuracy of the GC-IRMS system. The precision for carbon-isotope analysis was better than 0.3‰ (working standard analysis, SD, 1σ), and the deviation of $\delta^{13}\text{C}$ values for the working standards from the calibrated values was generally <0.4‰. Each sample was analyzed in duplicate, and the average for the two runs was accepted as the final sample isotopic composition.

The major constituents of the longer chain *n*-alkanes in most of the DLK samples are the C_{21} , C_{23} , C_{25} , C_{27} , C_{29} *n*-alkanes; in addition, the samples from the CPE interval contain C_{17} and C_{19} *n*-alkanes (Fig. S13). This breakdown is in agreement with the high HI values and low OI values and increased percentage of the

sapropelinite group from the samples of the CPE interval, representing an increase of aquatic sources for organic matter^{53,54}. Compound-specific long-chain *n*-alkane (C₂₃–C₃₁) $\delta^{13}\text{C}$ values in the DLK section show both a distinct $\sim 6.5\text{‰}$ negative isotopic excursion and a $\sim 4.0\text{‰}$ negative isotopic excursion, corresponding to the bulk $\delta^{13}\text{C}_{\text{org}}$ CIE1 and CIE2, respectively (Fig. S12). This result indicates that carbon-cycle perturbations in the atmosphere are consistent with the bulk $\delta^{13}\text{C}_{\text{org}}$ in the lacustrine deposits, suggesting that the mix of organic matter from algae, pollen and spores, waxy leaf and wood could have amplified the magnitude by a few permil compared to the average $\sim 3\text{--}4\text{‰}$ CIE1 in marine deposits⁵⁵. Source changes are therefore clearly not the main cause of the negative CIE in bulk $\delta^{13}\text{C}_{\text{org}}$. Major ^{13}C -depleted carbon released into the ocean–atmosphere system, and associated enhanced hydrological cycling and continental weathering, and increased nutrient supply to oceans and lakes, led to a major global shift in $\delta^{13}\text{C}$ and increased aquatic primary organic productivity (e.g., algae) with high HI values⁵⁶.

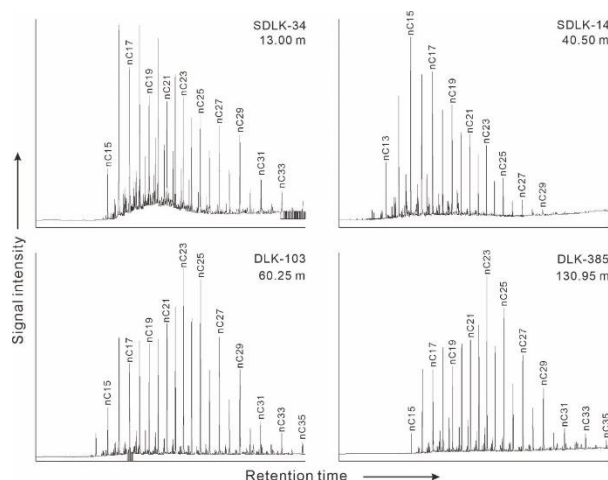


Fig. S13. Characteristic chromatogram (Total Ion Current, TIC) of extracted apolar fraction from four samples in the DLK section. Identified *n*-alkanes peaks are marked.

Text S5. Mercury as a volcanic proxy

Mercury (Hg) concentrations have emerged as a promising proxy for volcanic activity and particularly subaerial emplacement of large igneous provinces, as volcanic eruptions are the dominant natural source of Hg into the Earth's surface system^{57–59}. Mercury in the atmosphere has a residence time of 0.5–2 yr⁵⁸, allowing hemispheric and potentially global dispersal. After deposition from the atmosphere, changes in mercury loading can be reconstructed using marine and terrestrial sediments, where enhanced Hg loading may be expressed as higher concentrations, accumulation rates or spikes of ratios of Hg to common sedimentary host phases such as total organic carbon, Hg/TOC⁵⁹. In addition, Hg isotopes, particularly mass-independent fractionation (MIF) of odd Hg isotopes (e.g., $\Delta^{199}\text{Hg}$), have the potential to track Hg sources and deposition pathways^{59–61}. Hg-MIF ($\Delta^{199}\text{Hg}$) is primarily controlled by kinetic photo-chemical reduction of mercuric Hg and is not easily affected by post-depositional sedimentary processes⁶⁰. The processes of photo-chemical reduction generate positive $\Delta^{199}\text{Hg}$ and negative $\Delta^{199}\text{Hg}$ values in marine and terrestrial systems, respectively; in addition, near-zero values are found in volcanic gases^{59–62}.

Mercury (Hg) concentrations in the DLK section show considerable variation, ranging from 0.01 to 299.26 ppb, with the highest values occurring in strata stratigraphically below the interval of CIE1 at 34.50 m to 50.50 m (Fig. S14, S15). The Hg concentrations show a reasonable covariation with TOC ($r = +0.70$), whereas detrital material (Al, $r = -0.06$) and total sulfide ($r = +0.30$) (Fig. S14) present much weaker relationships. We normalized Hg concentrations to TOC, excluding all samples with TOC values below 0.2 wt.% (except for one sample with TOC of 0.19 wt.%). Previous studies have shown that Hg behaviour is non-linear with host-phase abundance, resulting in elevated normalized Hg values when host-phase abundance is low (e.g., low TOC or low TS)^{63–65}. Moreover, changes in source material have been postulated to impact normalized Hg⁶⁶.

From background values of <100 ppb/wt.% (at 0 – 36.50 m, pre-ME interval), the ratio of mercury to total organic carbon (Hg/TOC) rises to the peak value of 512

ppb/wt.% at 37.25 m, a level that is also lower than the position of the CIE1 interval, then returns to background levels at 40.50 m (Fig. S15). Note that the normalized values for CIE1 are likely suppressing normalized Hg values due to enhanced TOC preservation⁶³ and potentially a system change from receptor to supply limitation⁶⁵.

Mass independent fractionation (MIF) of odd Hg isotopes ($\Delta^{199}\text{Hg}$) shows stable values (-0.07 to $+0.01\text{‰}$) within the background intervals and a larger range of values (-0.12 to $+0.05\text{‰}$) within the CPE interval (Fig. S15).

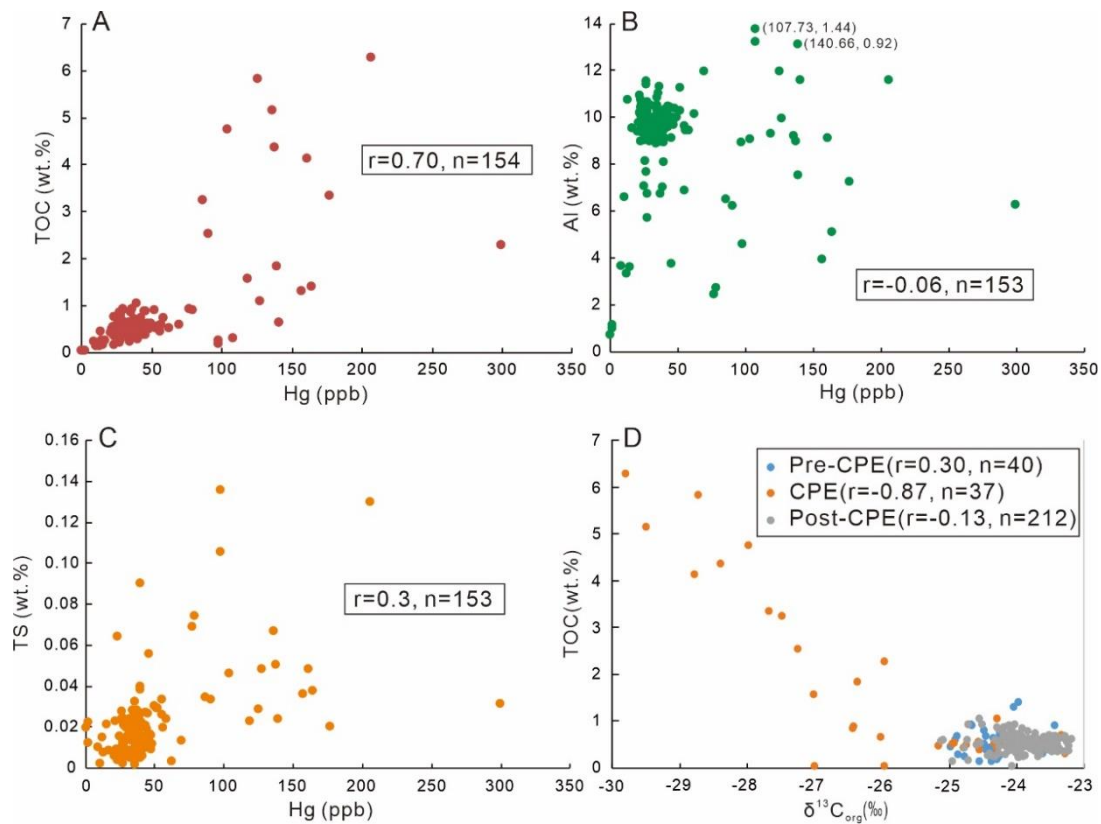


Fig. S14. Cross-plots of Hg versus total organic carbon (TOC) (a), total sulfur (TS) (b), aluminum (Al) (c), and TOC versus $\delta^{13}\text{C}_{\text{org}}$ values (d) at the DLK section.

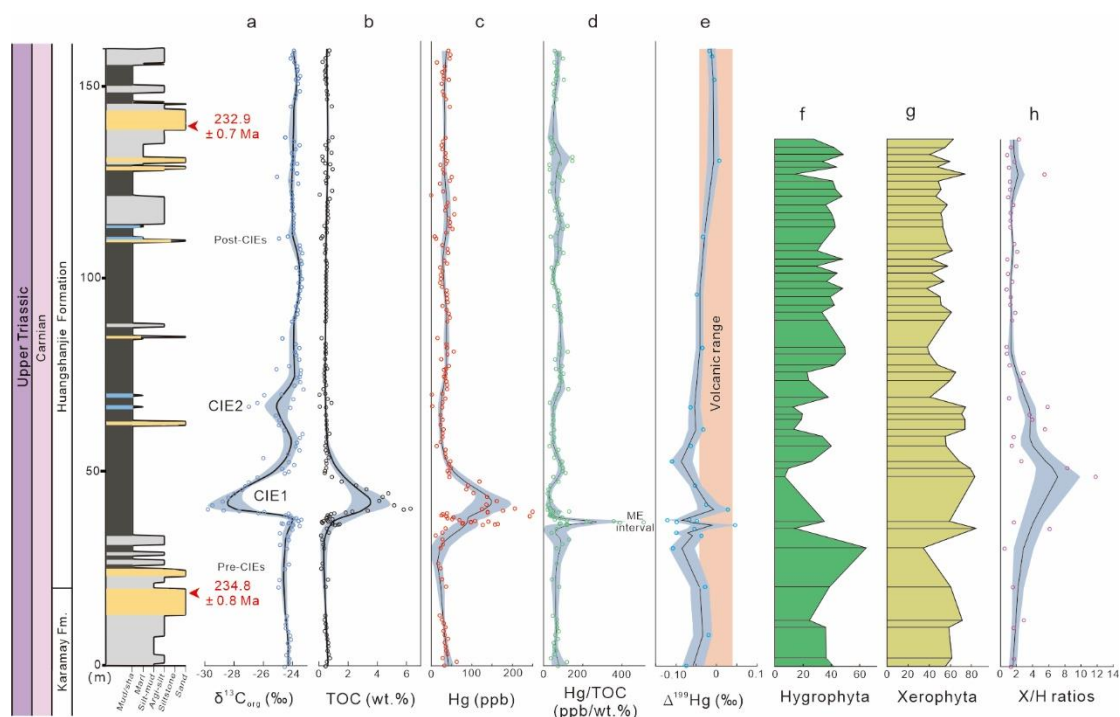


Fig. S15. Geochemical and Palynological profiles of the lacustrine Dalongkou section. **a**, Organic carbon isotopes ($\delta^{13}\text{C}_{\text{org}}$, ‰). **b**, Total organic carbon (TOC, wt.%). **c**, Mercury concentrations (Hg, ppb); **d**, Ratios of mercury to total organic carbon (Hg/TOC, ppb/wt.%), ME—mercury enrichments. **e**, Mass independent fractionation of odd-Hg isotope ($\Delta^{199}\text{Hg}$, ‰). **f**, hygrophytic abundances. **g**, xerophytic abundances. **h**, xerophytic/hygrophytic (X/H) ratios. Thick lines in (**a–e**) indicate the LOESS trendline (10% smoothing in **a–d** and **f**, 25% smoothing in **e**, 20% smoothing in **h**). The shaded envelopes indicate the range of 2σ uncertainty in **a–e** and σ uncertainty in **h**. Abbreviations: Mud, Mudstone; Sha, Shale; Silt-mud, Silty mudstone; Argi-silt, Argillaceous siltstone; Sand, Sandstone.

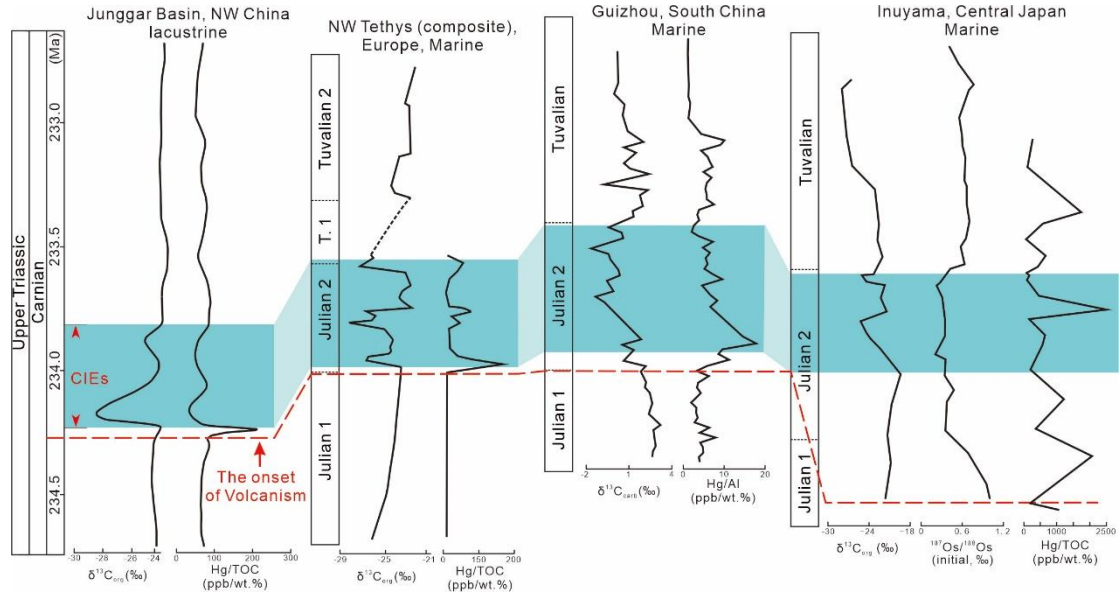


Fig. S16. Correlation of $\delta^{13}\text{C}$, Hg/TOC and $^{187}\text{Os}/^{188}\text{Os}$ values from the Junggar Basin (this study), NW Tethys (Europe)⁶⁷, South China⁶⁸ and Central Japan^{69,70}.

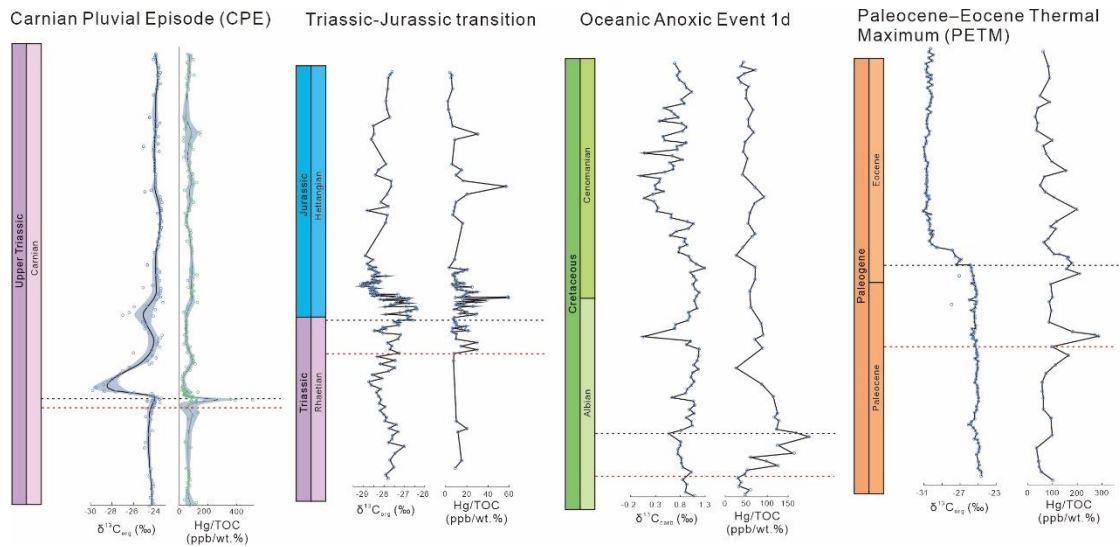


Fig. S17. Comparison of Hg/TOC characteristics between the Carnian Pluvial Episode (this study), Triassic–Jurassic transition⁷¹, Oceanic Anoxic Event 1d⁷², and Paleocene–Eocene Thermal Maximum⁷³. Red dashed lines represent the start of volcanic eruptions, and black dashed lines represent the onset of hyperthermal events.

Text S6. Simple mass-balance calculation of CPE carbon-cycle evolution

We estimated the amount of isotopically light carbon released using a simple mass-balance calculation:

$$M_r = M_t (\text{CIE}) / (\delta^{13}\text{C}_r - \delta^{13}\text{C}_t - \text{CIE})$$

M_t and $\delta^{13}\text{C}_t$ represent the amount and isotopic composition of the whole exogenic carbon reservoir (with $M_t = 42400$ Gt and $\delta^{13}\text{C}_t = -2.27\text{‰}$; these are assumed modern values based on ref.⁷⁴); CIE represents the magnitude of the global carbon-isotope excursion ($\approx -3\text{‰}$ or -4‰)⁵⁵; $\delta^{13}\text{C}_r$ represents the isotopic composition of the carbon released from methane ($\approx -60\text{‰}$)⁷⁴. Following the above equation, we can estimate the amount the carbon released from dissociation of methane hydrate required to produce such a CIE to be about 2.282×10^{18} to 3.043×10^{18} g (2,282–3,043 GtC).

In our simulation with 568 ppmv CO_2 , the Carnian global permafrost surface area was about $3.05 \times 10^6 \text{ km}^2$ (Fig. 4). We assume an organic-carbon content of near-surface (0–3 m soil depth) is 100 kg/m^2 (ref.⁷⁵). With these figures, the total amount of carbon buried in the Carnian permafrost was estimated to be 3.05×10^{17} g (305 GtC). Therefore, carbon from permafrost alone is unlikely to have caused a measurable carbon-isotope shift in the whole exogenic carbon reservoir.

Text S7. Seasonal spatio-temporal climatic changes during the CPE

Geological reconstructions indicate that the Earth's climate around 235 million years ago was much warmer than today, with a global mean surface air temperature of around $25^{\circ}\text{C}^{23,76,77}$. We performed two model simulations to reconstruct the spatio-temporal climatic changes during the CPE. The simulations indicate that the subtropical western continent was hot and dry prior to the CPE (Fig. S20a,c). By contrast, the eastern continent, including the Junggar Basin (this study), was relatively wet due to the presence of an oceanic warm pool and an Inter-Tropical Convergence Zone over the Tethys Ocean. Under CO_2 forcing, the subtropical dry climate zone expands poleward, causing significant drying in most of the subtropical continent, with the exception of the southeast subtropical continent, where enhanced monsoons are predicted to have contributed to increased summer precipitation (Fig. S20b,d).

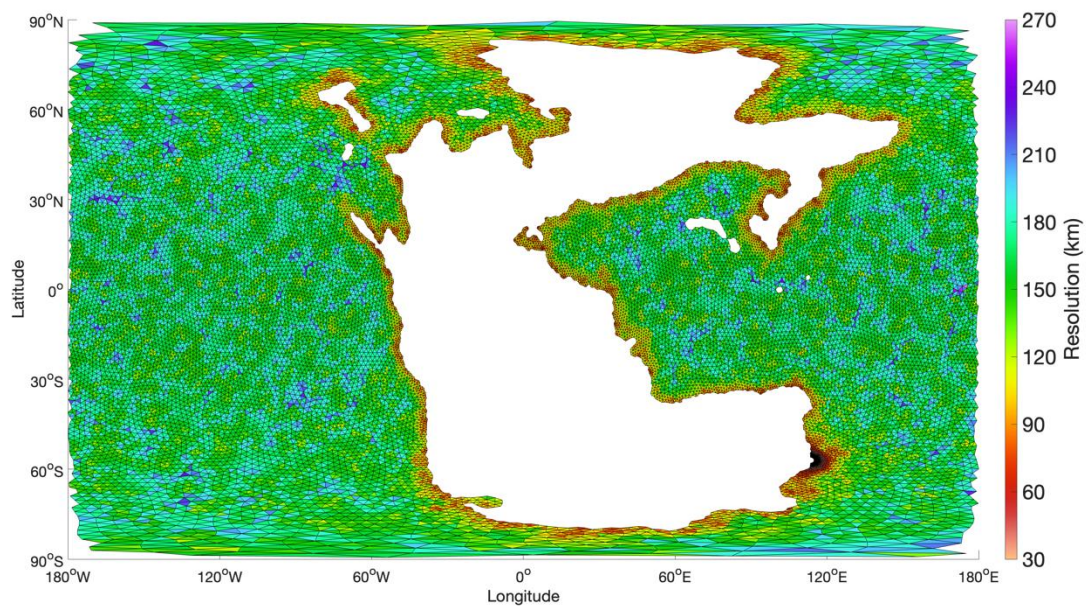


Fig. S18. The applied horizontal resolution of ocean component in the climate simulations reconstructing the Carnian Pluvial Episode climate.

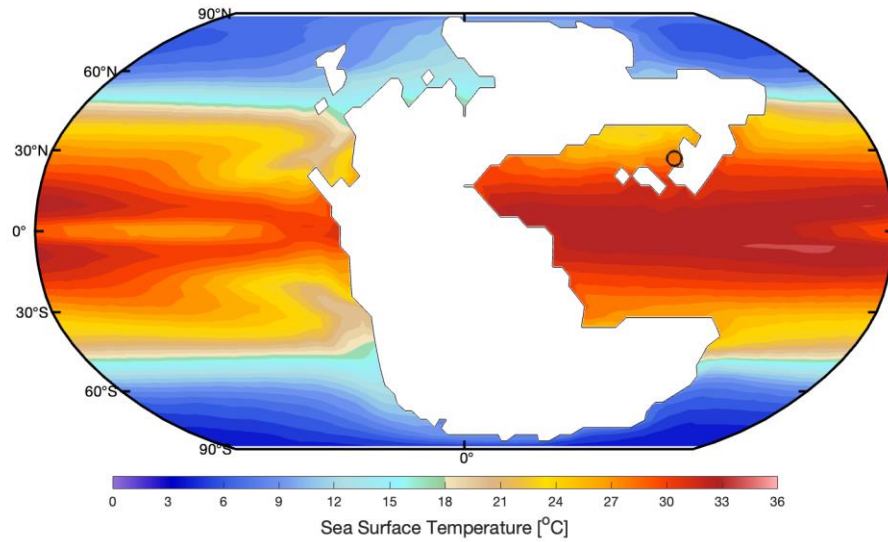


Fig. S19. The simulated sea-surface temperature before the CPE. The colour inside the black circle represents reconstructed sea-surface temperature around the subtropical eastern Tethys, and is in agreement with the oxygen-isotope records from conodont apatite from South China²³.

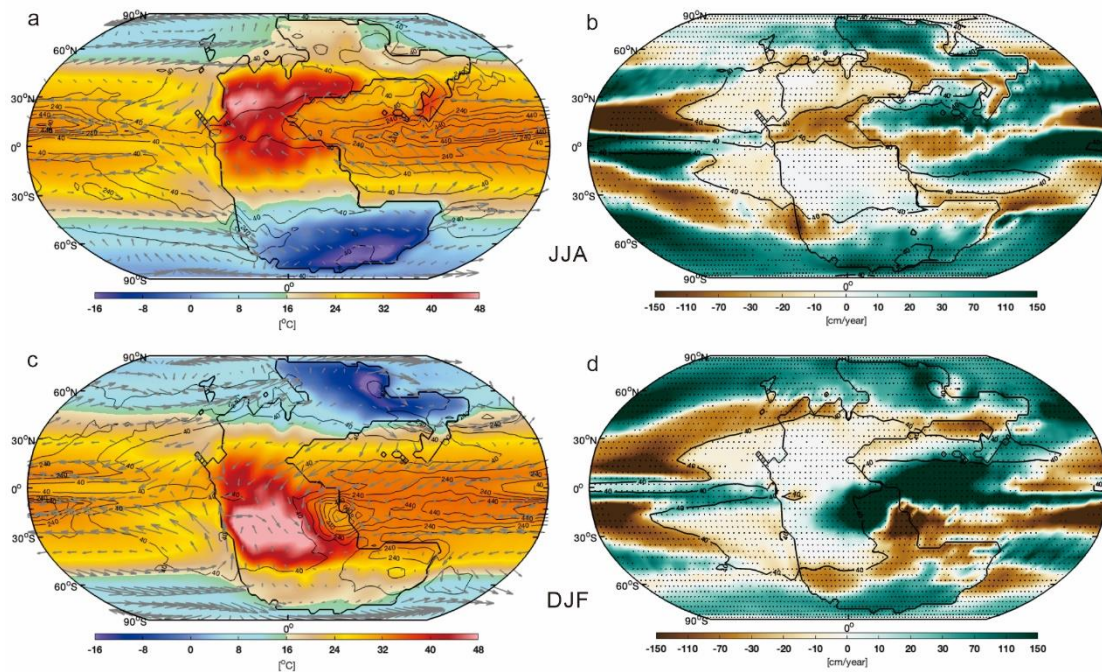


Fig. S20. **a**, Boreal summer (June, July, August) 2-m air temperature (shading), precipitation (black contours) and near-surface wind (vectors) before the CPE (with $p\text{CO}_2$ set as 568 ppmv). **b**, Boreal summer precipitation anomalies during the CPE

380 ($p\text{CO}_2$ changes from 568 ppmv to 1758 ppmv). The black contours locate the arid
381 climate zone with precipitation lower than 40 cm/year. Similar to subpanels (a) and
382 (b), subpanel (c) and (d) illustrate results from the boreal winter (December, January
383 and February). Results based on AWI-ESM simulations under land–sea configuration
384 of around 235 million years ago.

Text S8. Palynology

Forty-seven samples of mudstones and fine siltstones were collected for palynological analysis. Rock samples were cleaned, crushed, weighed (*c.* 50 g per sample), and processed using standard palynological processing techniques⁷⁸. Residues were sieved using 10 and 200 μm meshes. All forty-seven samples were productive. A minimum of 200 spores and pollen were counted for each sample (Supplementary Data 4). Palynological processing was performed at the Nanjing Institute of Geology and Palaeontology, Chinese Academy of Sciences (NIGPAS). Data collection and photography were carried out using an AXIO LAB A1 light microscope at NIGPAS. All rock samples and slides are stored at NIGPAS.

Pollen and spores were linked to their parent plants or morphological groups to analyse compositional changes of local vegetation (Supplementary Data 4), including Bryophyta (mosses), Pteridophyta (ferns), Sphenophyta (horsetails), Lycophyta (club-mosses), Non-striate bisaccates ('seed-ferns' or conifers), Striate bisaccates ('seed-ferns' or conifers) and Non-saccate pollen (gymnosperms). Pollen and spores were grouped into hygrophytes (wet-adapted) and xerophytes (dry-adapted) according to the Palaeoclimatological Model, which was instituted by Visscher and van der Zwan (1981)⁷⁹ and updated in subsequent studies^{1,4,5,9,10,22,80–82} (Supplementary Data 4). As suggested by Traverse (2007)⁷⁸, most pollen and spores are transported by water, with airborne pollen being a minor component and primarily derived from nearby regions rather than distant areas. Furthermore, since the Junggar Basin was a highly restricted basin, aeolian activity is likely to have had a minimal impact on the palynological compositions. Therefore, our palynological data are well-suited for reconstructing the palaeoclimate of the Junggar Basin.

In total, twenty-six spore genera, four monosulcate pollen genera, and two bisaccate pollen categories (striate and non-striate bisaccate pollen) were recorded in the DLK section, and their parent plants were grouped into Bryophyta, Pteridophyta (ferns), Sphenophyta, Lycophyta, monosulcate-producers (gymnosperms), Pteridosperms and conifers, based on botanical affinities (Fig. S21). Bryophyta, Sphenophyta, Lycophyta, and striate bisaccate pollen-producers show no distinct

trends throughout the section. Ferns exhibit a peak at approximately 30 m (represented by the peak of *Apiculatisporis*) and an abundance decline over approximately the 35–75 m interval (mostly represented by the decline of *Leiotrilete* and *Todisporites*), succeeded by a modest rebound in abundance upwards to the top of the section; a relatively drier condition, recorded in sediments over the interval ~ 35–75 m, is indicated, given that Mesozoic ferns are suggestive of wet-adapted plants (e.g. ref.^{83,84}). Monosulcate pollen-producers show an increase in abundance from approximately 35 m upwards; however, their debated ecological preferences^{1,4,10,85} hampered further understanding of their palaeoclimatic implications.

Hygrophytic elements constitute ca. 30% (averaging 34%) of the palynofloras throughout the studied section, and their relatively lower abundances (averaging 22%; down to approximately 10%) between 35–75 m indicate a comparatively drier condition, although three hygrophytic upheavals within this drier interval (at approximately 40, 55 and 70 m) possibly represent minor climatic oscillations (Fig. S15). Hygrophytic plant recovery starts at approximately 75 m, suggesting that a climatic shift to wetter conditions is recorded upwards to the top of the studied section.

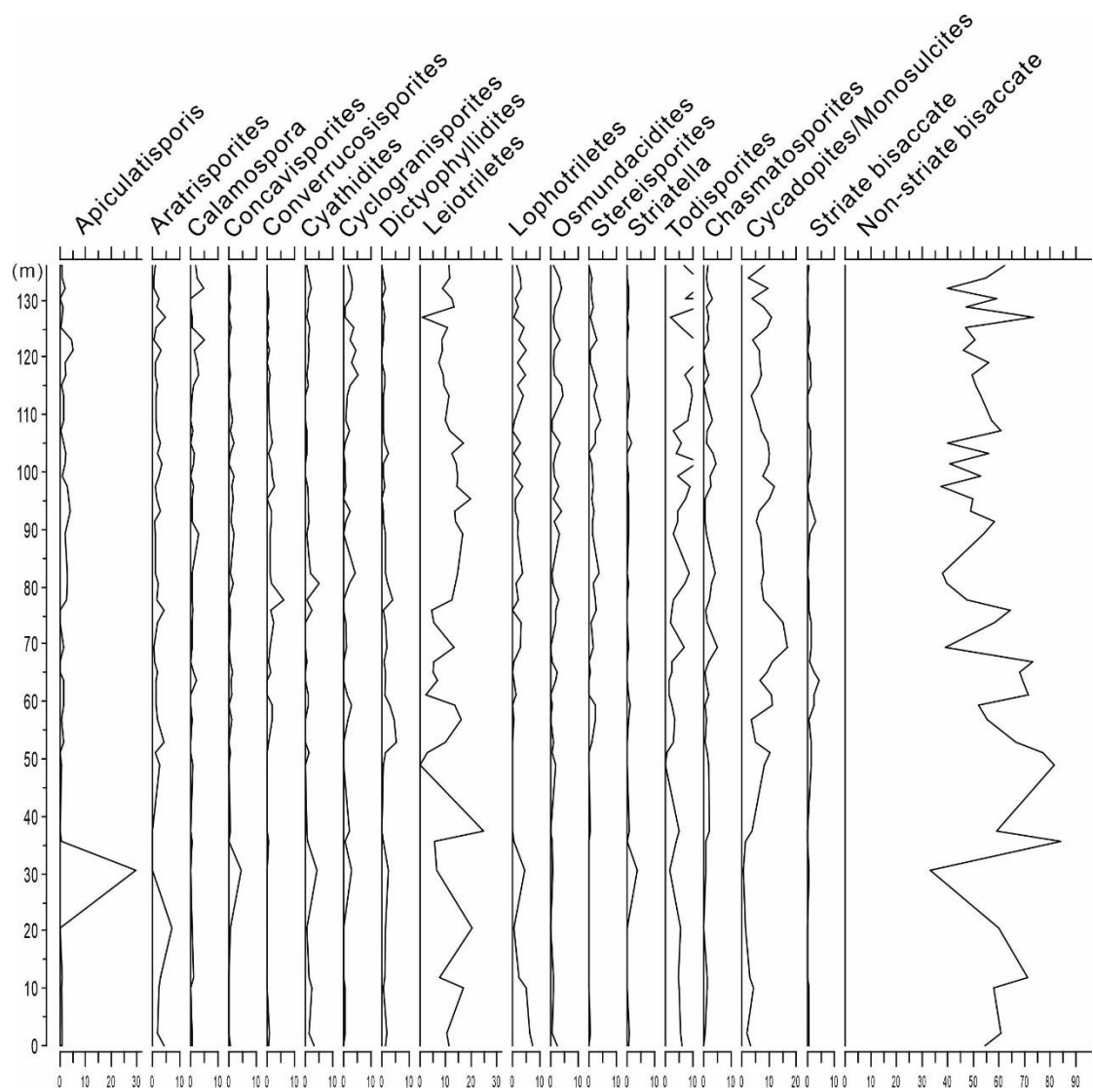


Fig. S21. Results of spore-pollen genera from the DLK section in the Junger Basin. More detailed data are shown in Supplementary Data 4.

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