Supplementary Information for

Marine anoxia linked to abrupt global warming during Earth’s penultimate icehouse

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S1. Paleogeography of the South China sections
During the late Carboniferous, the South China Block was a nearly isolated terrain, located across the paleo-equator at the confluence of the Paleo-Tethys Ocean (west) and Panthalassic Ocean (east) based on various paleogeographic reconstructions (Fig. S1A and Fig. 5A). The Carboniferous-Permian carbonate successions in South China were widely developed on carbonate platform and intra- and inter-platform slopes and basins that defined an open-water seaway (1, 2). The Naqing and Narao successions (~22 km apart) formed on carbonate slopes of the Qian-Gui Basin, now located in Luodian region, Guizhou Province (Fig. S1B).

These successions consist mainly of thin-bedded lime mudstone with episodic bioclastic wackestone to packstone that were deposited by relatively proximal sediment gravity flows (3). The carbonate successions contain abundant conodonts that are used to define a high-resolution bio- and chronostratigraphy, for which the well-studied Naqing section has been chosen as the GSSP (the Global Stratotype Sections and Points) candidate for the four unratified Carboniferous stages (i.e., Serpukhovian, Moscovian, Kasimovian, and Gzhelian stages) (4).

S2. Age model of δ\(^{13}\)C and δ\(^{238}\)U
Ages for geochemical data of the South China Naqing succession were calibrated using the Carboniferous timescale (5) and our updated conodont biostratigraphy (4, 6), assuming constant sedimentation rate within stages. The ages of samples and geochemical data from the Naqing section are relatively reliable given that the stage boundaries (between Moscovian, Kasimovian, Gzhelian, and Asselian) are well constrained based on bed-by-bed conodont biostratigraphic studies (4, 6). The stratigraphic depth and age of the KGB were assigned to 255.65 m (paint marks) and 303.7 Ma, respectively. The Carboniferous astronomical time scale was also established in the Naqing section (7); the ages of samples were, however, calibrated assuming that the stratigraphic depth and age of the KGB are 220.33 m (aluminum spike marks) and 303.4 Ma, respectively. We recalibrated the astronomical ages of our samples using the numerical numbers (255.65 m (paint marks) and 303.7 Ma) that we used for age calibration assuming a linear sedimentation rate. The difference between the two sets of age calibration for the interval between 306–301 Ma is ~0.09 myr on average (Fig. S2).

The base of the Gzhelian at the Narao section is also well constrained by the detailed conodont biostratigraphy (6), but the bases of the Kasimovian and Asselian are not studied in detail, and therefore, the ages of the geochemical data at the Narao section are not as reliable as those of the Naqing section. We thus use only the Naqing geochemical data (carbonate δ\(^{13}\)C and δ\(^{238}\)U) and associated ages for the LOSCAR and \(^{238}\)U modeling.

S3. Evaluation of Diagenetic Alteration
The relationship between δ\(^{13}\)C\(_{\text{carb}}\) and δ\(^{18}\)O\(_{\text{carb}}\) is commonly used to access the degree of diagenetic alteration, which may be represented as a linear form or inverted ‘J’ pattern (8). The δ\(^{13}\)C\(_{\text{carb}}\) and δ\(^{18}\)O\(_{\text{carb}}\) data from the Naqing and Narao successions do not exhibit co-variation (with \(r^2<0.06\)), or the “meteoric calcite line” or inverted ‘J’ trends indicative of diagenetic alteration. The δ\(^{18}\)O\(_{\text{carb}}\) of the two studied successions are primarily between -8‰ and -2‰, with an average value of ~4.0‰ for the Naqing (n=141) and -3.4‰ for the Narao (n=103) successions (Figs. S5–S7). The δ\(^{13}\)C\(_{\text{carb}}\) values of the Naqing and Narao successions largely overlap with those of diagenetically screened, well-preserved, calcitic brachiopods (9), further supporting that the South China carbonates record depositional seawater δ\(^{13}\)C (1, 10).

It has long been suggested that δ\(^{13}\)C\(_{\text{carb}}\) and δ\(^{13}\)C\(_{\text{org}}\) are not altered simultaneously, and thus coupled changes in δ\(^{13}\)C\(_{\text{carb}}\) and δ\(^{13}\)C\(_{\text{org}}\) most likely suggest a perturbation of global C cycling rather than diagenetic alteration (11). However, a recent study on the drill core of the Great Bahama Bank suggests coupled δ\(^{13}\)C\(_{\text{carb}}\) and δ\(^{13}\)C\(_{\text{org}}\) time series can record alteration of the δ\(^{13}\)C\(_{\text{carb}}\) by \(^{12}\)C-enriched meteoric waters while incorporation of \(^{12}\)C-enriched terrestrial organic C is added to the sediment during subaerial exposure (12). Both the Naqing and Narao successions were deposited in carbonate slope settings and detailed facies
analysis suggests that these successions were unlikely to have been altered by meteoric fluids (3, 13). Therefore, the paired δ\(^{13}\)C\(_{\text{carb}}\) and δ\(^{13}\)C\(_{\text{org}}\) negative excursion in the Naqing and Narao successions is most likely a result of perturbation to C cycling.

Trace elements (e.g., Sr and Mn) can be used to access the fidelity of primary geochemical compositions of carbonate minerals given that they substitute into the lattice of carbonate minerals. Recent studies of Bahaman sediment (14, 15) indicate no obvious correlation between δ\(^{238}\)U and common sedimentary diagenetic indicators including Mn/Sr ratios in bulk carbonates, although early diagenesis systematically shifts carbonate δ\(^{238}\)U by up to 0.27 ± 0.14‰ heavier than contemporaneous seawater (14, 15). Nevertheless, it was suggested that both limestones and dolostones with Mn/Sr ratios <10 can be expected to retain their primary marine isotopic signatures, such as δ\(^{13}\)C values (16, 17), although other studies suggested Mn/Sr ≤2 as a cutoff criteria (18). Studies of carbonate δ\(^{238}\)U use relatively more conservative cutoffs of Mn/Sr (e.g., <1 (19, 20); <2 (21, 22), or <2.5 (23)) for diagenetic evaluation, whereas other studies do not use Mn/Sr as a tracer for diagenetic alternation (24-26).

For the Naqing and Narao successions, a total 72 U samples show Mn/Sr ratio <2.5, and only 9 samples have Mn/Sr ratios varying between 2.5 and 3.5. Nonetheless, 8 out of the 9 samples are generally consistent with the overall δ\(^{238}\)U trend (Figs. S5–S7). The one outlier sample (NQ255.8) contains relatively high Mn/Sr (=2.9) and anomalously high Mo/Al (784), suggesting the high δ\(^{238}\)U value (0.02‰) might have resulted from secondary incorporation of U from reducing, and most likely sulfidic, pore water enriched in Mn and Mo and \(^{238}\)U (24, 27). The other obvious “outlier” (NQ254.2), although with very low Mn/Sr (=0.9), has a very negative δ\(^{238}\)U value of -0.69‰ and lies totally off the overall δ\(^{238}\)U trend (Fig. S5), the cause of which is not clear. Negative δ\(^{238}\)U values might record the influence of either dissolution of Fe or Mn oxides (24, 28) or dolomitization under meteoric and hypersaline waters with more negative δ\(^{238}\)U (27, 29). However, the sample has a low Mn/Sr (=0.9) and Mg/Ca (=0.03) ratio (Figs. S6 and S7), suggesting it was not altered by the aforementioned processes. We examined cross-plots of δ\(^{238}\)U against δ\(^{13}\)O, [Sr], [Mn], and Mn/Sr ratios for all the samples (Figs. S6 and S7), and no correlations are observed. Furthermore, existing studies suggest that bulk carbonates faithfully record the general trend of seawater δ\(^{238}\)U, regardless of palaeogeographic location (20), palaeo-bathymetric variations (26), or carbonate lithology (30).

Whereby, we interpret our δ\(^{238}\)U records to most likely reflect original seawater isotopic variability, although absolute isotopic values might be overall overprinted and shifted towards more positive values during early diagenesis.

**S4. Evaluation of Detrital Contamination**

Incorporation of δ\(^{238}\)U from detrital material would artificially shift the measured carbonate δ\(^{238}\)U to slightly more positive values given that detrital material have relatively higher δ\(^{238}\)U (approximately -0.3‰) (31, 32), inherited from continental crust, than seawater values (-0.39‰) (14). The Naqing and Narao carbonates mostly have low Rb/Sr ratios of < 0.03, suggesting minimal incorporation of clay minerals in the carbonates (Figs. S6 and S7). Exceptions are the NR228.8 with ~0.11 and NR229 with ~0.13 from the Narao section (Figs. S6 and S7), which are from the slightly calcareous black mudstone immediately below the KGB. The high TOC and high Rb/Sr ratios of these two samples (NR228.8 and NR229) suggest that their relatively high δ\(^{238}\)U values may record secondary overprint.

Mo, Zn and Fe are redox-sensitive elements that are often enriched in sediments with dissolved H: S present in pore fluids (33). Enrichment of these elements may also result from elevated terrestrial input (34). The enrichment of Mo, Zn and Fe in the samples immediately below the KGB is most likely due to high clay contents with high [Al] (Figs. S5–S7). Thus, the δ\(^{238}\)U values across the KGB that partly come from organic-rich, carbonate-poor samples suggest that the carbonate δ\(^{238}\)U could have only been more negative if the analyzed U were partially leached from the clay portion of the bulk sediment powder or were overprinted by authigenic enrichment of isotopically heavy U(IV) in carbonates during early diagenesis.

**S5. Biodiversity reconstruction**

A total of 1,215 foraminiferal species, including 1,068 fusulinid species and 147 non-fusulinid species, were compiled for the Kasimovian through Gzhelian, based on the global foraminiferal database (provided
by J.R. Groves). Eight time bins with one-million-year interval were assigned for the Kasimovian–Gzhelian interval. The foraminiferal diversity was obtained according to the species counting of each time slice during this interval. In this study, the time framework was recalibrated based on the Carboniferous timescale (5) and the foraminiferal ages in the database of J.R. Groves (35). The diversity curve of all marine species comes from a new Cambrian-Triassic biodiversity curve with an imputed temporal resolution of 26 ± 14.9 kyr, which was calculated using a parallel computing program of constrained optimization method – CONOP.SAGA, and data from 11,000 marine fossil species, collected from over 3,000 stratigraphic sections (36). The biodiversity curve suggests a ~24.9% loss of total marine species, and notably, this biodiversity crisis represents a pronounced interruption of the recently recognized, long-term late Carboniferous–early Permian Biodiversification Event (36).

S6. Potential impact of sea level on proxy-derived temperature change

Quantifying the influence of sea-level change on δ¹⁸O-derived temperature changes requires sufficient knowledge of changes in global land ice volume. Although our climate model simulations (LowCO₂ and HighCO₂) incorporate glacial-interglacial changes in Gondwanan land ice surface extent based on the most up to date knowledge of glacial evidence for the LPIA (see Experimental design portion of the Coupled Climate Model Simulations section), we cannot use ice volume estimates from LowCO₂ and HighCO₂ simulations to constrain the influence of ice volume and sea-level change on measured δ¹⁸O values because poorly constrained paleo-elevation and thickness of Gondwanan ice sheets. In the version of CESM1.2 used for LowCO₂ and HighCO₂ simulations, land ice is prescribed by lowering the surface albedo over specific regions of the land surface and the surface extent of the ice sheet remains fixed throughout the simulation. Because the elevation and thickness of Gondwanan ice sheets is unknown, we set the elevation of prescribed ice sheets to values that compare well with those of the modern Greenland and Antarctic ice sheets (ranging between ~1,500–3,000 m above sea level). As a result, calculations of total ice volume based on prescribed ice sheets in LowCO₂ and HighCO₂ are highly dependent on the assumptions made about the proportion of land:ice within each ice center.

For example, if we assume that all of the grid cells under the ice sheet surface (above sea level) are ice, then the total land ice volumes in LowCO₂ and HighCO₂ are 170E6 km³ and 76E6 km³, respectively, with a change in total ice volume of 94E6 km³ (~160 m sea level change) (37). We recognize that in reality there would be some proportion of land underneath each ice sheet, so these total ice volume values are high compared to the most recent ice volume estimates (37), even though the areal extent of the prescribed ice sheets are based on the most up to date knowledge of glacial evidence. If we make an assumption of the proportion of land to ice within each ice center, i.e., conservatively that 25% of the volume within each ice sheet is the land surface underneath the ice sheet, then the total ice volume in LowCO₂ and HighCO₂ would decrease to 127.5E6 km³ and 57E6 km³, respectively, and the change in total ice volume would be 70.5E6 km³ between the simulations (~120 m sea-level change) (38). These absolute estimates now lie within the current range of ice volume estimates for the LPIA (37), but there is no way to validate this assumption based on glacial proxy evidence. Thus, the ice volumes we infer from the prescribed land ice in each climate simulation are not sufficiently robust to be used to back-out the influence of sea-level change on the proxy δ¹⁸O values as a means of further constraining the paeo SSTs. Although the representation of continental ice in CESM1.2 and the uncertainty in Gondwanan ice thickness precludes using such prescribed ice volume estimates to disentangle the influences of ice volume, sea level, and temperature signals archived in the proxy δ¹⁸O values, proxy-model comparison gives us confidence that the δ¹⁸O-derived temperature changes are reasonable. In Main Text Fig. 1C, sea surface temperatures calculated based on the δ¹⁸O of calcite brachiopods correspond to mean values of ~25.1 and 29.4°C for the pre-warming state and amplitude of warming, respectively. The simulated mean sea surface temperatures (upper 100 m) in the South China region for the LowCO₂ and HighCO₂ simulations are 25.5 and 29.8°C, which corresponds to a regional warming of 4.3°C. Thus, the simulated sea surface temperatures compare very well with the amplitude of warming of sea surface temperatures inferred from calcite brachiopod δ¹⁸O values (~4.3°C), providing confidence that the elevations and areal extents of the prescribed ice sheets in the low and high CO₂ experiments are reasonable.
S7. Regression analysis

We smoothed the geochemical records with a locally weighted polynomial regression (LOWESS) method using the Bootstrap function in freeware ACYCLE (39) and PAST (40), which yield essentially similar trends. We also used a cross-validation approach in MATLAB script to select robust smoothing factors for various datasets.
Fig. S1. Paleogeography of the Pennsylvanian. (A) Global paleogeography of the Pennsylvanian (adapted from Ron Blakey), showing the location of South China. (B) Pennsylvanian paleogeography of South China adapted from Jiao et al. (2003) (5), showing locations of the Naqing (N 25° 14′ 40″, E 106° 29′ 26″) and Narao (N 25° 24′ 39″, E 106° 36′ 25″) sections.
Fig. S2. Age model of $\delta^{13}$C and $\delta^{238}$U for the Naqing section. (A) Comparison of our interpolated ages based on stratigraphic height and conodont biozones assuming constant sedimentation rate within stages, and astronomically calibrated ages using stratigraphic height (220.33 m of aluminum spikes) and age (303.4 Ma) of the Kasimovian–Gzhelian boundary (7). (B) Comparison of our interpolated ages based on stratigraphic height and conodont biozones assuming constant sedimentation rate within stages, and astronomically recalibrated ages by correcting the precise stratigraphic height (255.65 m of paint marks) and age (303.7 Ma) of the Kasimovian–Gzhelian boundary.
Fig. S3. Sensitivity tests of the evolution of average surface ocean $\delta^{13}$C reconstructed using LOSCAR during the carbon emission event. By varying climate sensitivity, $\delta^{13}$C of the carbon source, carbon emission duration and the amount of carbon emission, we have generated 432 scenarios in total.
Fig. S4. Sensitivity tests of the evolution of atmospheric $\rho$CO$_2$ reconstructed using LOSCAR during the carbon emission event. By varying climate sensitivity, carbon emission duration and the amount of carbon emission, we have generated 108 scenarios in total.
Fig. S5. Correlation between $\delta^{238}\text{U}$ and trace elements and total organic carbon (TOC) of the Naqing (A) and Narao (B) successions for evaluation of diagenetic alteration and detrital contamination. The light red and gray shaded area indicates intervals with relatively high contents of TOC and negative $\delta^{238}\text{U}$ values across the KGB, respectively. Note the difference in thickness of individual conodont biozones and lithofacies between the two sections.
Fig. S6. $\delta^{238}$U and stable isotopes and elements cross-plots of the Naqing succession for diagenetic evaluation. Lack of co-variations suggests minimum alteration on carbonate $\delta^{238}$U by later diagenesis.
Fig. S7. δ²³⁸U and stable isotopes and elements cross-plots of the Narao succession for diagenetic evaluation. Lack of co-variations suggests minimum alteration on carbonate δ²³⁸U by later diagenesis.
Fig. S8. Linked LOSCAR and U mass balance model results of greenhouse hyperthermal events, including PTB (A), TJB (B), OAE2 (C), and PETM (D) events. Carbonate $\delta^{238}$U data with a regression line (blue); red line indicates diagenetic correction (subtracting 0.27‰). PTB data from ref. (19, 20, 41), TJB data from ref. (42), OAE2 data from ref. (26), and PETM data from ref. (43). LOSCAR modeling (black solid line) of atmospheric $p^{\text{CO}_2}$ compared to available $p^{\text{CO}_2}$ data (red cycle) for TJB (44). U riverine flux generated from LOSCAR and used in U cycle modeling. Model estimate of anoxic seafloor area ($f_{\text{anox}}$) of PTB and TJB based on diagenetically corrected $\delta^{238}$U data, whereas for the OAE2 and PETM based on original values, as suggested by original papers that the data were derived from. For $f_{\text{anox}}$ panels, the upper and lower bounds of the shaded area represent the 97.5th and 2.5th percentile, respectively, while the red solid line represents the mean.
Fig. S9. Comparison between the KGB warming event and greenhouse C perturbation events over the last 300 Myr, using various $k_{\text{anox}}$ values for sensitive test on $f_{\text{anox}}$ (see Materials and Methods in the main text). (A) and (B) Increase in anoxic seafloor area ($f_{\text{anox}}$) modelled using $k_{\text{anox}}=5.10E-05$ yr$^{-1}$. (C) and (D) Increase in anoxic seafloor area ($f_{\text{anox}}$) modelled using a full range of $k_{\text{anox}}$ between 5.10E-05 yr$^{-1}$ and 1.74E-04 yr$^{-1}$ integrated in Monte Carlo simulation (SI Appendix, Table S3). The dashed trend lines are plotted based on all events excluding the KGB and PETM. Data points indicate simulations from this study only (linked LOSCAR and U modeling); error bars are the 97.5$^{\text{th}}$ and 2.5$^{\text{th}}$ percentile.
Fig. S10. Coupled climate model simulations. (A) Annual mean meridional overturning circulation by Eulerian mean flow (Sv) in the late Pennsylvanian Panthalassic Ocean in HighCO2 (560 ppm, interglacial state) and LowCO2 (280 ppm, glacial state) simulations. Red indicates clockwise circulation, and blue indicates counterclockwise circulation. (B) Depth-latitude cross-section of ideal age (years) in the upper 1000 m of the Panthalassic Ocean, suggesting ventilation in the Southern Hemisphere for both simulations and a lack of ventilation in the Northern Hemisphere in HighCO2. Ideal Age is a tracer in the ocean model (POP2) of CESM used to estimate ventilation timescales in the ocean. Water masses with a low ideal age have been in recent contact with the surface and/or mixed with young water masses, while water masses with a high ideal age have been removed from the surface for that duration in years. (C) Vertical profiles of Panthalassic seawater density at 30˚N, the equator, and 30˚S show enhanced surface stratification in HighCO2 (solid lines) compared to LowCO2 (dashed lines). (D) Near-equilibrium conditions over the last 300 years of the LowCO2 and HighCO2 simulations. Global mean upper (0 m) and deep (3000 m) ocean temperature (top). Variability in the maximum strength of the Northern Panthalassic Overturning Circulation (positive values) and Southern Panthalassic Overturning Circulation (negative values) (bottom). The global mean salinity remains stable throughout each of the simulations.
Table S1. U mass balance model parameters.

| Parameter       | Value                                      | Unit  | Reference |
|-----------------|--------------------------------------------|-------|-----------|
| $N_{sw}$ (initial)* | $1.96 \times 10^{13} (\pm 20\%)$          | mol   | (45)      |
| $\delta^{238}U_{riv}$ | -0.24 (± 0.1)                             | ‰     | (46)      |
| $k_{anox}$      | $5.10 \times 10^{-5} (\pm 20\%)$ to $1.74 \times 10^{-4}$ | yr$^{-1}$ | (23, 47) |
| $k_{other}$     | $1.74 \times 10^{-6} (\pm 20\%)$          | yr$^{-1}$ | (23)      |
| $\Delta_{anox}$ | 0.55 to 0.85                               | ‰     | (23, 46, 48) |
| $\Delta_{other}$ | -0.1 to 0.1                               | ‰     | (23)      |
| $\delta^{238}U_{sw}$ | $\delta^{238}U_{carb} - 0.27$ (where indicated by authors) | ‰     | (32)      |
| $f_{anox}$ (initial) | determined by spin up*                      | -     | -         |
| $A_{seafloor}$  | $3.62 \times 10^{14}$                      | m$^2$ | (49)      |
| $J_{riv}$       | $(2.35 \times 10^{-6}) \times J_{weath \_LOSCAR} - (7.98 \times 10^{-9})$ | mol/yr | (50)      |

*a new steady state is achieved during the spin up for each simulation based on the selected input parameters*
| Event | Total C injection (Gt) | Duration of C injection (kyr) | Injection rate (Gt/kyr) | Duration of SST warming (AT) | Rate of warming (°C/kyr) | $f_{anox}$ (%) | $f_{anox}$ (%) peak | $M_{anox}$ (%) | Note |
|-------|------------------------|-------------------------------|-------------------------|------------------------------|--------------------------|----------------|---------------------|-------------|------|
| KGB   | 5000                   | 300                           | 0.0167                  | 4                            | 0.005                    | 4.19           | 22.41               | 18.22       | 1    |
|       | 10000                  | 300                           | 0.0333                  | 4                            | 0.013                    | 4.29           | 22.46               | 18.17       |      |
|       | 7000                   | 17                            | 0.420                   | 10                           | 0.270                    | 0.2            | 17                  | 16.8        | 2    |
|       | 22400                  | 15                            | 1.520                   | 10                           | 0.500                    | 0.2            | 60                  | 59.8        |      |
|       | 22400                  | 15                            | 0.420                   | 9                            | 0.231                    | 3.0            | 48.1                | 45.1        | 3    |
|       | 12000                  | 20                            | 0.600                   | 3                            | 0.035                    | 0.21           | 8.4                 | 8.19        | 4    |
|       | 12000                  | 10                            | 1.200                   | 4                            | 0.047                    | 0.21           | 21.0                | 20.79       |      |
|       | 12000                  | 85                            | 0.141                   | --                           | --                       | 2.9            | 48.0                | 45.1        |      |
| PTB   | 6000                   | 220                           | 0.027                   | 2                            | 150                      | 0.013          | --                  | 2.5         | 6    |
|       | 18750                  | 220                           | 0.113                   | 5                            | 150                      | 0.033          | --                  | 7.5         |      |
|       | 7200                   | 150                           | 0.048                   | 1.5                           | 100                      | 0.015          | 3                   | 2.7         | 7    |
|       | 18900                  | 150                           | 0.126                   | 2                            | 100                      | 0.020          | 2.8                 | 5.2         |      |
|       | 18900                  | 150                           | 0.126                   | --                           | --                       | 3.02           | 4.49                | 1.47        | 8    |
|       | 7200                   | 150                           | 0.048                   | --                           | --                       | 3.15           | 5.01                | 1.86        |      |
|       | 18900                  | 150                           | 0.126                   | --                           | --                       | 1.00           | 2.03                | 1.83        |      |
|       | 28000                  | 150                           | 0.072                   | 1.1                           | 100                      | 0.011          | 3                   | 7.7         | 9    |
|       | 27000                  | 150                           | 0.180                   | 2.3                           | 100                      | 0.023          | 2.8                 | 12.2        |      |
|       | 27000                  | 150                           | 0.180                   | --                           | --                       | 4.50           | 9.32                | 4.82        | 10   |
|       | 10200                  | 21                            | 0.486                   | --                           | --                       | 0.50           | 2.33                | 1.83        | 11   |
|       | 9600                   | 3                             | 3.220                   | --                           | --                       | 0.50           | 2.35                | 1.85        |      |
|       | 9600                   | 75                            | 0.128                   | --                           | --                       | 0.2            | 2.03                | 1.83        |      |
|       | 9600                   | 5                             | 1.920                   | --                           | --                       | 0.2            | 2.23                | 2.03        |      |
|       | 0.300                  | 3                             | 3                        | 1.000                         | --                       | --             | 0.2                 | 1.10        | 13   |
|       | 1.700                  | 5                             | 21                       | 0.238                         | 0.900                    | 2.275          | 0.580               |      |      |

Note:
1: All estimates from this study.
2: C injection data from ref. (51); $f_{anox}$ data from ref. (23); T change data from ref. (52-54)
3: C injection data from ref. (51); $f_{anox}$ from this study
4: C injection data from ref. (55); $f_{anox}$ from ref. (42).
5: C injection data from ref. (55, 56); $f_{anox}$ from this study.
6: C injection data from ref. (57); $f_{anox}$ from ref. (58) base on Mo isotopes, and we add 50% error.
7: All from ref. (26).
8: C injection data from ref. (26); $f_{anox}$ from this study.
9: All from ref. (26).
10: C injection data from ref. (26); $f_{anox}$ from this study.
11: C injection data from ref. (59, 60) for max. and min. rate of C injection; $f_{anox}$ from this study.
12: All from ref. (43).
13: C injection data from ref. (60-64); Duration of T change was purposely chosen two extremes from ref. (59)
Table S3. Comparison of anoxic seafloor area ($f_{anox}$) during greenhouse OAE/hyperthermal events in various modeling efforts.

| Event | Peak $f_{anox}$ (%) of ref. (47) using $k_{anox}=1.74E-04$ yr$^{-1}$ | Peak $f_{anox}$ (%) of this study using $k_{anox}=5.10E-05$ yr$^{-1}$ | Peak $f_{anox}$ (%) of this study using a full range between $k_{anox}$ of ref. (23, 47) | Peak $f_{anox}$ (%) of previous studies |
|-------|---------------------------------------------------------------|---------------------------------------------------------------|---------------------------------------------------------------------------------|--------------------------------------|
| PTB  | $10.0^{±5.6}_{±14.0}$                                      | $48.10$                                                        | $35.01$                                                                           | $17–60$ (ref. (23, 47))             |
| TJB  | $3.7^{±1.3}_{±1.2}$                                          | $10.70$                                                        | $6.41$                                                                            | $8.4–21$ (ref. (42))               |
| OAE2 | $0.6^{±0.5}_{±1.1}$                                          | $9.32$                                                         | $4.56$                                                                            | $8–15$ (ref. (26))                |
| PETM | $0.15^{±0.0}_{±0.4}$                                         | $2.33$                                                         | $1.77$                                                                            | $2.03–2.23$ (ref. (43))           |

Dataset S1 (separate file). Isotope and element data.

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