Six-fold increase of atmospheric $p$CO$_2$ during the Permian–Triassic mass extinction

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The Permian–Triassic mass extinction was marked by a massive release of carbon into the ocean-atmosphere system, evidenced by a sharp negative carbon isotope excursion. Large carbon emissions would have increased atmospheric $p$CO$_2$ and caused global warming. However, the magnitude of $p$CO$_2$ changes during the PTME has not yet been estimated. Here, we present a continuous $p$CO$_2$ record across the PTME reconstructed from high-resolution $\delta^{13}$C of C$_3$ plants from southwestern China. We show that $p$CO$_2$ increased from 426 $^{+133/-96}$ ppmv in the latest Permian to $2507^{+4764/-1193}$ ppmv at the PTME within about 75 kyr, and that the reconstructed $p$CO$_2$ significantly correlates with sea surface temperatures. Mass balance modelling suggests that volcanic CO$_2$ is probably not the only trigger of the carbon cycle perturbation, and that large quantities of $^{13}$C-depleted carbon emission from organic matter and methane were likely required during complex interactions with the Siberian Traps volcanism.
The Permian–Triassic mass extinction (PTME; ca. 252 Ma) coincided with rapid global warming that produced one of the highest intervals of the Phanerozoic1–5, which was likely triggered by a massive release of greenhouse gases6–7. The emplacement of the Siberian Traps large igneous province has been widely suggested as the ultimate trigger for the extinction of ~90% of marine species and ~70% of terrestrial vertebrate species at the Permian–Triassic boundary8, with major losses amongst plants (e.g. refs. 9,10). Alongside volcanic degassing, CO₂, SO₂, and halogen volatiles were likely released due to thermal metamorphism by Siberian Traps’ intrusions into organic-rich sediments9–11. The global negative carbon isotope excursion (CIE) found in both marine and terrestrial settings at the PTME (for a review, ref. 13) indicates a major carbon cycle perturbation in the ocean-atmosphere system, which implies a rise in the atmospheric CO₂ levels (pCO₂). However, pCO₂ changes during the PTME still remain poorly constrained.

On the one hand, records of pCO₂ from proxies (stomata index, palaeosol carbonates, and biomarkers) are mainly focused on the late Permian and/or Phanerozoic long-term trends without detailed pCO₂ data for the earliest Triassic (refs. 13–17). On the other hand, various models show large variability of peak CO₂ changes during past hyperthermals is indeed crucial to understand the possible imminent environmental effects of today’s CO₂ increase: pCO₂ has risen from 280 to more than 400 ppmv in the last 150 years and is projected to go higher21. Experiments on living C₃ plants (in the field and in growth chambers) suggest that carbon isotope fractionation (Δ¹³C) during photosynthesis increases with increasing CO₂ levels, lowering the carbon isotope signature of C₃ plants (Δ¹³C,22. Based on this relationship, Δ¹³C calculated from δ¹³C, measured in fossil C₃ plants remains can be used as a proxy for past pCO₂. This proxy successfully reproduced ice-core records of pCO₂ for the Last Glacial Maximum23, and has been applied to reconstruct pCO₂ during Early Eocene hyperthermals24, the Cretaceous Period25, and the Toarcian Oceanic Anoxic Event26.

Here, we present high-resolution δ¹³C records of fossil C₃ plant remains from sedimentary successions of southwestern China. Using the δ¹³C data of C₃ plants, we calculated a six-fold increase of atmospheric pCO₂ during the PTME, from 426 ±133/−96 ppmv to 2507 ±4764/−1193 ppmv. Furthermore, the pCO₂ estimates are compared with carbon isotope mass balance calculations showing that in addition to volcanic CO₂, large quantities of ¹³C-depleted carbon emission from organic matter and methane were likely required to trigger the observed global negative CIE in the exogenic carbon pool.

**Results and discussion**

High-resolution terrestrial carbon isotope records. We present high-resolution terrestrial organic carbon isotope records (δ¹³Corg) from plant cuticles, wood and bulk organic matter (OM) together with our previous work10 from four terrestrial Permian–Triassic boundary sections (Chahe, Jiucaichong, core ZK4703 and Chinahe) in southwestern China (Supplementary Fig. 1; Supplementary Fig. 2). The δ¹³C of bulk OM and C₃ plant remains from the four study sections exhibit nearly identical secular trends (Fig. 1). Each profile can be divided into four stages: (1) a pre-CIE stage, (2) an onset of the negative CIE (onset of CIE) stage, (3) a prolonged CIE body stage and (4) a post-CIE stage. In the pre-CIE stage, δ¹³Corg records from the Xuanwei Formation are characterized by steady values around −25.0% (Fig. 1). The synchronous prominent onset of CIEs with peak values of −32% occurs at the bottom of the Kayitou Formation. Subsequently, the onset of the CIE stage is followed by a prolonged interval with sustained low values (ca. −30%) through the whole Kayitou Formation, interrupted by a slight positive shift immediately after the onset of CIE. A recovery to slightly higher δ¹³Corg values (−28% to −26%) starts in the uppermost part of the Kayitou Formation and the Dongchuan Formation. Previously published terrestrial δ¹³Corg profiles in southwestern China (e.g. refs. 27,28) all belong to mixed organic carbon source.

**Fig. 1** Carbon isotope excursion trend recorded in global terrestrial C₃ plants and marine carbonates. The secular carbon isotope excursion (CIE) trend can be divided into four stages (i.e. pre-CIE, onset of CIE, CIE body and post-CIE) in terrestrial bulk organic matter, C₃ plants and marine carbonates, and are shown as different color fields. The last appearance datum (LAD) of coal beds and Gigantopteris flora distributions represent the coal gap and collapse of tropical peatlands respectively10,45. Carbon isotope (δ¹³C) data source: Chahe (δ¹³C of bulk organic matter from ref. 27; δ¹³C of plants data from this study), Jiucaichong (this study), ZK4703 core and Chinahe (δ¹³C data in this study together with our previous work10), Amb (Pakistan)32 and global marine carbonate δ¹³C (Methods). The locations of marine and terrestrial carbon isotope profiles are shown in the late Permian palaeogeographic map.
in bulk OM. Few unusually negative values (< -34%) observed in the upper Kayitou Formation, e.g., in a published record from Chahe27, are statistical outliers and local signals, as such negative values are not observed in our high-resolution study. These outliers may be caused by local 13C-depleted samples possibly containing an algal and/or bacterial component29.

The four-stage terrestrial δ13Ccarb trend is also seen in the marine carbonate carbon isotope (δ13Ccarb) records (Fig. 1). A total of 10 global-distributed marine Permian–Triassic boundary sections with both high-resolution δ13Ccarb and conodont biostratigraphy were integrated as a global marine δ13Ccarb profile, using the age model from the Meishan Global Stratotype Section and Point (GSSP)30 (Supplementary Fig. 3; Supplementary Fig. 4). These newly compiled global δ13Ccarb records are nearly identical to those published previously (e.g. ref. 12).

$\rho$CO2 estimates based on $\Delta^{13}C$ of fossil plants. Constraining the magnitude of the CIE is crucial to estimate accurate mass, rate, and source of the 13C-depleted carbon released during the crisis22. δ13C profiles from southwestern China (low latitude) and Pakistan32 (middle latitude) exhibit CIE magnitudes of ca. -7% and ca. -5.5% respectively, which are significantly larger than the ca. -3.5% marine CIE magnitude estimated from global marine δ13Ccarb records (Fig. 1). Data compilation confirms this discrepancy: terrestrial CIE magnitudes range from -3.6% to -6.1% (bulk OM, 25th percentile to 75th percentile, n = 29), and from -5.2% to -7.1% (C3 plants, n = 9), whereas marine CIE magnitudes range from -3.0% to -4.7% (n = 69) (Fig. 2; Supplementary Table 1). As shown both in modern and fossil plants, elevated atmospheric $\rho$CO2 was likely responsible for amplifying the magnitude of the CIE in the terrestrial δ13C record due to an increase of $\Delta^{13}C^{22,33}$. Therefore, following the relationship between $\Delta^{13}C$ and $\rho$CO2 developed by Cui and Schubert24 (Methods), we could calculate the $\rho$CO2 across the PTME. The $\Delta^{13}C$ was calculated using the δ13Cp records of four study sections from southwestern China, and the δ13CCO2 (the $\Delta^{13}C$ of atmospheric CO2; Supplementary Fig. 5) calculated from the global marine δ13Ccarb compiled in this study. As explained above, this is possible because the marine and terrestrial δ13C records are closely comparable and can be readily correlated (Fig. 1), the correlation being supported also by biostatigraphy (flora and conchostracans), and radioisotope dating (Supplementary Fig. 6; Supplementary Information). The initial, background late Permian $\rho$CO2(0) = $q_0$ is set in our calculations at 425 ± 68 ppmv based on the late Changhsingian $\rho$CO2 estimates calculated by Li et al.16 using stomatal ratio method and mechanistic gas exchange model for fossil conifers from the Dalong Formation in South China, with good age control and reliable taxonomy.

Our estimates (Fig. 3) show that $\rho$CO2 was moderately low (252.1 Ma within the pre-CIE stage (upper part of conodont Clarkina changxingensis zone). Subsequently, the $\rho$CO2 began to increase rapidly in the Clarkina yini zone, reaching a maximum level (2507 +4764/-1193 ppmv), immediately after the Permian–Triassic boundary (Hindeodus parvus zone). This near-six-fold increase of atmospheric $\rho$CO2 occurred within ~75 kyr and coincided with the onset of the global CIE. The $\rho$CO2 remained high (ca. 1500 to 2500 ppmv) immediately after the onset of the CIE, with only one transient drop (down to ca. 1300 ppmv). Coupled to the recovery of $\Delta^{13}C$, $\rho$CO2 drops to ca. 700 ppmv at the top of Isarcicella isarica zone. Atmospheric CO2 levels show a close correlation with estimated sea surface temperatures ($r = +0.60$, $p < 0.001$, n = 173; Supplementary Fig. 7), implying that CO2 was the dominant greenhouse gas across the PTME, although the contribution of other greenhouse gases such as methane and water vapor cannot be excluded here. The six-fold increase of atmospheric $\rho$CO2, together with a 10°C increase in sea surface temperatures estimated from low latitude conodont oxygen isotope (Fig. 3) implies Earth system sensitivity (ESS) of 3.9 °C per doubling of CO2, if we assume ESS equals to $\Delta T/\log_{2}(\rho$CO2(peak)/$\rho$CO2(background))34. This is consistent with a previous estimate of the Permian–Triassic ESS35 and the IPCC equilibrium climate sensitivity range of 1.5 to 4.5 with a median of 3.036, suggesting slow feedbacks operated in the geologic past. However, climate model simulations reveal that the increase of SST in high latitude should be higher than low latitude37. As a result, the 10°C SST increase in low latitude might underestimate the global SST increase, which leads to an underestimate of the Earth system sensitivity during the PTME.

Comparison with previous studies and uncertainty. Previous $\rho$CO2 estimates around the Permian–Triassic boundary (Fig. 3; Supplementary Table 2) come from stomatal proxies16,17, palaeosol carbonates13,14, phytane15 and carbon cycle modelling (e.g. refs. 18–20). Published proxy-based $\rho$CO2 reconstructions are mostly for the late Permian, within long-term and very low-resolution Phanerozoic records. Stomata-based estimates from modified fossil Ginkgo stomatal index method17 gave $\rho$CO2 around 400–800 ppmv in the latest Permian, but with poor age constraint and high taxonomic uncertainty16. Latest Permian $\rho$CO2 from δ13C of palaeosol carbonates from Texas, US13, was calculated at 400 ppmv36,39 (re-calculated by ref. 38 correcting the assumed soil respired CO2 concentration), but latest Permian palaeosol carbonate record from the Karoo Basin14 shows higher
values (883–1325 ppmv). Similarly, the δ13Cphytane-based pCO2 estimates show that CO2 levels during Changhsingian could have ranged from 873 to 1085 ppmv15. The few earliest Triassic peak pCO2 estimates from stomata17 and phytane15 proxies show significant variation (600–2100 ppmv). Simulations with various climate models (e.g. carbon cycle box modelling18,19 and cGENIE20) show major variability of peak pCO2 values (1000–9380 ppmv; Supplementary Table 3), using a large range of assumed background pCO2.

Several effects, especially diagenesis40, chemical treatment41, plant taxonomic changes42 and precipitation43,44 can influence the δ13Cp and consequently affect pCO2 estimates. The original signals of δ13Cp values can potentially be altered by diagenesis during burial40 and chemical treatment during sample preparation41. However, the method we use to calculate palaeo-pCO2 considers a relative change of the Δ13C that minimizes these biases (Methods). A dramatic plants turnover occurred in southwestern China during the PTME, with a Gigantopteris flora (spore plant) in the Xuanwei and basal Kayitou formations replaced by an Tomistrobus (spore plant) and Peltaspernum (seed plant) dominated flora45. Experiments on modern plants indicate lower Δ13C in seed plants than in spore plants42. Using a plant assemblage including a mixture of different taxa and plant tissues is better than using single species and plant remains when using δ13Cp as pCO2 proxy33. In this study we used a mixture of different plant tissues (i.e. cuticle, charred wood and non-charred wood), which very likely includes different plant taxa.

An increase of the mean annual precipitation (MAP) can also increase Δ13C44,46. This effect is negligible in sites experiencing high precipitation (>1500 mm/yr)47, such as the studied area in southwestern China, which was a humid, equatorial peatland during the PTME45. The plant community changed from Gigantopteris flora-dominated rainforest ecosystem to isotelalean-dominated (lycophyte) herbaceous vegetation that inhabited the surrounding margins of coastal oligotrophic lakes, which indicate fairly constant precipitation regimes during the PTME interval48,49. The sedimentology of the Xuanwei and Kayitou formations suggests there was no significant precipitation change across the mass extinction (Supplementary Fig. 8; ref. 50). In contrast, low MAP can explain the smaller magnitude of the CIE (~3%) recorded at the PTME in the semi-arid locations of Karoo Basin and North China31. In summary, the persistently humid condition in southwestern China was unlikely to have affected plant Δ13C, thus the pCO2 estimates are considered robust. A Monte Carlo method has been applied to evaluate the uncertainties (Supplementary Information; Supplementary Fig. 9), which reveals that the uncertainty in the pCO2 increases with increasing pCO2, as seen in the previous studies51.

Potential source of 13C-depleted carbon during the PTME. The ultimate source of 13C-depleted carbon capable to trigger the observed negative CIE, is widely debated. Several climate models of varying complexities (e.g. simple box models18,19 and cGENIE20) use different light carbon sources to fit the δ13C of marine carbonates (Supplementary Table 3). Proposed 13C-depleted carbon sources include biotic or thermogenic methane (Δ13C ≈ −60% to −40%; e.g. ref. 18), CO2 from thermal metamorphism or rapid oxidation of organic-rich rock (Δ13C ≈ −25%; e.g. ref. 6,19,52,53), and
Carbon emission caused prolonged high pCO2 and high temperature (ca. 35 °C) during the earliest Triassic (H. parvus and I. isarcica zones) and may have lasted for > 500 kyr (Fig. 3). This lengthy phase of extreme warmth likely implies prolonged carbon emissions into ocean-atmosphere system from continued eruption of the Siberian Traps volcanism, and/or reduced carbon sequestration rate, potentially due to lower consumption of atmospheric CO2 through reduced organic carbon burial and the possible failure of the silicate weathering thermostat59.

Methods
Sample treatment and analysis. In total, 68 samples from Chinhae, 41 samples from ZK4703 and 40 samples from Jiucaichong were analyzed for bulk organic carbon isotopes. Samples were crushed to fine powder (<200 mesh), and ~2 g powder were weighed, placed into a centrifuge tube and treated with 3 mol/L HCl for 24 h to remove the carbonate. Then the treated samples were rinsed with ultrapure water repeatedly until neutralized and finally dried at 35 °C. For C3 plants δ13C analysis, 45 samples from Chinhae, 26 samples from ZK4703, 30 samples from Jiucaichong and 13 samples from Chahe were treated with concentrated HCl and HF, then sieved over 500 μm and a 100 μm mesh screen to get the 100–500 μm particles. C3 plant fragments, including cuticle, non-charred wood and charred wood (charcoal), were picked under the microscope. The δ13C analyses were performed by using an element analyzer (EA, Costech) coupled to an isotope ratio mass spectrometer (Thermo Delta V Advantage) at the State Key Laboratory of Biogeology and Environmental Geology of the China University of Geosciences (Wuhan). The results were calibrated using certified secondary references standards USGS40 (δ13C = −26.39‰) and UREA (δ13C = −37.32‰) and given in per mil (%) relative to Vienna Pee Dee Belemnite (VPDB) with analytical precision better than ± 0.2‰. A Multi EA 4000-analyzer was used for TOC at China University of Geosciences (Wuhan), yielding an analytical precision of 1.5%.

Carbon isotope compilation and estimate of CIE magnitude. In order to estimate a reliable magnitude of the CIE, the carbon isotope profiles that record a roughly complete CIE shape within pre-CIE and CIE body are selected in this study. The compilation consists of 69 marine carbonate carbonate isotope (δ13Ccarb) profiles and 38 terrestrial δ13Corg profiles. The δ13Ccarb profiles recording complete negative CIE are from Eastern Palaeothetics (n = 29), Western Palaeothetics (n = 19), Central Palaeothetics (n = 4), Northern Neothetics (n = 4), Southern Neothetics (n = 10) and Panthalassa (n = 3). Bulk marine organic matter δ13C records were not included, because they often represent a mix of various organic components (both marine and terrestrial). The sedimentary facies belong to a range of shallow shelf, deep shelf, and slope environments. The few reported δ13Ccarb profiles from deep basins are ignored in this compilation (e.g. Shangsi section), because the elevated water stratification and large vertical δ13C DIC gradients at deep basin sites during the Permian–Triassic crisis could large CIE magnitude.49 A total of 38 terrestrial δ13Corg records are reviewed from eight terrestrial basins including western Guizhou and eastern Yunnan in southwestern China (n = 14), Junggar Basin (n = 3), Turpan Basin (n = 1), North China (n = 3), Central European Basin (n = 3), Bowen Basin (n = 2), Sydney Basin (n = 7), Karoo Basin (n = 1) and three oceanic regions where organic matter (OM) in samples are C3 plants or a mix of organic matter dominated by C3 plants including South China (n = 1), Boreal realm (n = 2) and South Neothetics (n = 3). Among these terrestrial δ13Corg profiles, there are nine records of δ13C organics from C3 plants (5 δ13Corg and 4 δ13Cfa) from Meishan section, Amb section in South Neothetics, southwestern China, and others are all bulk δ13Corg profiles.

In order to demonstrate the difference of marine and terrestrial CIE magnitudes, carbon isotope values immediately before the CIE (δ13Cbackground) and peak values (δ13Cpeak) are used to calculate the magnitude of the CIE (Δδ13Cpeak − δ13Cbackground). Note that 31 pairs of δ13Cbackground and δ13Cpeak values are from marine sections that are well constrained by latest Permian conodont occurrences (e.g. C. changxingensis, H. praeparvus, H. latidentatus zones) and earliest Triassic conodont (H. parvus and I. isarcica zones) occurrences. To test if the discrepancy of the CIE magnitude in different substrates (marine carbonate, terrestrial bulk OM, terrestrial C3 plant tissues) is statistically significant, we used a non-parametric Kruskal-Wallis test (function kruskal.test), using R software. The Wilcoxon function wilcox.test test was performed in R software to determine whether means of two independent groups (marine vs. terrestrial) are equal or not. Boxplots were drawn to visualize discrepancy in CIE magnitude of different substrates. All statistical analyses and graphing functions were undertaken using R.

C3 plant proxy. The carbon isotope fractionation in C3 plants (δ13C) and atmospheric pCO2 is described as a hyperbolic relationship22.53.61:

$$\Delta^{13}C = \frac{(A+B)(pCO_2 + C)}{A+B(pCO_2 + C)}$$

where

\[
\Delta^{13}C = \frac{(A+B)(pCO_2 + C)}{A+B(pCO_2 + C)}
\]
The original δ¹³C signals in C₃ plant can be altered by several effects (e.g. diagenesis [1], chemical treatments [2]), that potentially influence pCO₂ calculations. In order to minimize this effect, the data set is analyzed by a relative change in the Δ¹³C value between the time of interest (t) and a reference time (t = 0), designated as ΔΔ(Δ¹³C):

\[
\Delta\Delta(\Delta^{13}C) = \Delta(\Delta^{13}C) - \Delta(\Delta^{13}C)_{t=0}
\]

We can expand this as:

\[
\Delta\Delta(\Delta^{13}C) = (A)(B)\left(\frac{pCO_{2}(t)}{pCO_{2}(0)} + C\right) - (A)(B)\left(\frac{pCO_{2}(t)}{pCO_{2}(0)} + C\right)_{t=0} + A + (B)\left(\frac{pCO_{2}(t)}{pCO_{2}(0)} + C\right)_{t=0} + A + (B)\left(\frac{pCO_{2}(t)}{pCO_{2}(0)} + C\right)
\]

By rearranging Eq. (3), pCO₂(t) at any given time can be calculated by

\[
pCO_{2}(t) = \Delta\Delta(\Delta^{13}C) \cdot A \cdot B \cdot \frac{pCO_{2}(0) + C}{A + (B)\left(\frac{pCO_{2}(0)}{pCO_{2}(0)} + C\right)_{t=0}} + 2 \cdot \Delta(\Delta^{13}C) \cdot A \cdot B \cdot \left(\frac{C_{0} + C_{1}}{A \cdot B - \Delta(\Delta^{13}C)}\right) \cdot B - \Delta(\Delta^{13}C) \cdot B \cdot pCO_{2}(0)\]

where A, B, C are curve fitting parameters. Values for A and B are 28.26 ± 0.022, respectively [3], which could produce more robust pCO₂ estimates compared with other parameter values in subsequent research [4]. The C is the function of the A and B values (C = A + 4.4/((A − 4.4) × B)). The pCO₂(t=0) is equal to the pCO₂ before the negative CIE, determined from independent stomatal proxies based on fossil conifers from the Dalong Formation in South China [5]. Because of the good age control (Clarkina changxingensis conodont zone), reliable taxonomy and calculation method, these stomatal estimates are considered as robust pCO₂ estimates before CIE. The mean value for the stomatal estimates is 425 ± 68 ppmv set as pCO₂(t=0). The δ¹³C in C₃ plant is in a South China conodont zone from Meishan are selected as standard [6]. LOESS curves with 0.5% (whose age equals to 28.26 ± 0.022) were used to fit the δ¹³C data set as pCO₂(t=0). The δ¹³C in C₃ plant is constant and independent of temperature, so that the δ¹³C value at any given time can be calculated by

\[
\delta^{13}C_{t} = \delta^{13}C_{0} \cdot \left(1 + \delta^{13}C_{carb}/\delta^{13}C_{DIC}\right)
\]

The δ¹³Ccarb data from Meishan (n = 199) is the most abundant of all the δ¹³Ccarb data (n = 707), thus, we performed a LOESS fitting based on δ¹³Ccarb data without Meishan data (Supplementary Fig. 4).

In order to calculate pCO₂(t), we need to align global marine δ¹³Ccarb profiles and δ¹³Ccarb based on same age model. The nearly same CIE curves were divided into four stages in carbonate and C₃ plants records to ensure the correlation between marine and terrestrial carbon isotope profiles. The age model for four sections is shown in Supplementary Information. In addition, the LOESS method with 0.002 Myr spacing was also performed in δ¹³Ccarb and temperature data to get the probability maximum value and standard error at each 0.002 Myr time step. The Monte Carlo method was employed to propagate input error [7] and served as propagating package in R. All the input parameters were assumed to be Gaussian distributed with mean and standard deviations listed in Supplementary Table 4. 10,000 values for each input parameters were randomly sampled to calculate 10,000 values for each pCO₂(t). The invalid pCO₂(t) values, pCO₂(t) < 100 ppmv or >130 ppmv were excluded. The 16th and 84th percentiles of the remaining estimates were determined to construct the 68% confidence interval. The positive error of the reconstructed pCO₂(t) value represents the difference between the 84th percentile value and the median, and the negative error represents the difference between the 16th percentile value and the median. The sensitivity analysis of C₃ plant proxy is discussed in Supplementary information and Supplementary Fig. 9.

Carbon isotope mass balance. This model used to evaluate the light carbon source, following mass balance equation is modified from McInerney and Wing [8]:

\[
M_{\text{added}} = \left(\delta^{13}C_{\text{peak}} - \delta^{13}C_{\text{Odstock}}\right)
\]

where M_{\text{added}} is the mass of carbon added into atmosphere-ocean system carbon emission. The δ¹³Cbackground represents initial carbon reservoir size during Permian-Triassic including ocean and atmosphere carbon inventory, but dominated by the ocean reservoir. Thus, the δ¹³Cbackground is assumed to be the initial marine DIC reservoir size ranging from 66,000 to 82,000 Gt. CIE represents the global magnitude of CIE controlled only by release of light carbon effect, which is set as a series values from −1% to −8%. The peak δ¹³C value (δ¹³Cpeak) at the event is calculated by initial isotopic composition of global carbon reservoir (δ¹³Cbackground) and CIE (δ¹³Cpeak = δ¹³Cbackground + CIE). The δ¹³Cbackground is assumed to be the initial isotopic composition of DIC reservoir 2.2% that is estimated from global marine δ¹³Ccarb profiles (age = 252-104 Ma). The δ¹³Cbackground is the δ¹³C value of the carbon source causing the CIE. Four kinds of carbon sources are involved including biogenic methane buried in permafrost or seafloor (δ¹³C = −60%), thermomorganic methane (δ¹³C = −40%), thermal metamorphism or rapid oxidation of organic-rich rock (δ¹³C = −25%) and CO₂ released from direct volcanic eruption (−6%). Finally, the increased CO₂ is estimated from M_{\text{added}} (1 Gt C = 0.3 ppmv; ref. 9), and compared with reconstructed atmospheric CO₂ levels from C₃ plant proxy.

Data availability

The authors declare that all data supporting the findings of this study are available within the paper and its supplementary file.

Code availability

R code to run the model is available from D.L. Chu on request.

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**Author contributions**

D.L.C., Y.C., and Y.Y.W. designed this study with in-depth inputs from H.J.S., P.B.W., J.D.C., and J.N.T. Y.Y.W. and D.L.C completed the data preparation and analysis with the help from H.Y.S., Y.D., and J.D.C. Y.Y.W. and Y.C. performed the calculations. Y.Y.W and D.L.C wrote the main text and supplementary materials with inputs from all authors.

**Competing interests**

The authors declare no competing interests.

**Additional information**

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