Thermogenic carbon release from the Central Atlantic Magmatic Province caused major end-Triassic carbon cycle perturbations

Thea H. Heimdal¹, Morgan T. Jones¹ & Henrik. H. Svensen¹

¹Centre for Earth Evolution and Dynamics (CEED), University of Oslo, PO Box 1028, Blindern, NO-0315 Oslo, Norway.

Corresponding author: Thea Hatlen Heimdal, Centre for Earth Evolution and Dynamics (CEED), University of Oslo, PO Box 1028, Blindern, NO-0315 Oslo, Norway, +47 41548103, t.h.heimdal@geo.uio.no.

Keywords

Central Atlantic Magmatic Province, end-Triassic extinction, C-cycle perturbations, C-cycle modeling

Competing interest statement

The authors declare no competing interests.

Abstract

The Central Atlantic Magmatic Province (CAMP), the end-Triassic mass extinction (ETE), and associated major carbon cycle perturbations occurred synchronously around the Triassic-Jurassic (T-J) boundary (201 Ma). Negative carbon isotope excursions (CIEs) recorded in marine and terrestrial sediments attest to the input of isotopically light carbon, although the
carbon sources remain debated. Here we explore the effects of mantle-derived and thermogenic carbon released from the emplacement of CAMP using the Long-Term Ocean-Atmosphere-Sediment CARbon Cycle Reservoir (LOSCAR) model. We have tested a detailed emission scenario grounded by numerous complementary boundary conditions, aiming to model the full extent of the carbon cycle perturbations around the T-J boundary. These include three negative CIEs (i.e. Marshi/Precursor, Spelae/Initial, Tilmanni/Main) with sharp positive CIEs in between. We show that a total of ~ 24,000 Gt C (including ~ 12,000 Gt thermogenic C) replicates the proxy data. These results indicate that thermogenic carbon generated from the contact aureoles around CAMP sills represents a credible source for the negative CIEs. An extremely isotopically depleted carbon source, such as marine methane clathrates, is therefore not required. Furthermore, we also find that significant organic carbon burial, in addition to silicate weathering, is necessary to account for the positive δ¹³C intervals following the negative CIEs.

**Significance statement**

The Central Atlantic Magmatic Province (CAMP) is coincident with the end-Triassic extinction event and several negative carbon isotope excursions (CIEs). Sill emplacements in Brazil would have generated extensive volatiles and degassing due to the contact metamorphism of evaporites, organic-rich shales, and hydrocarbons. Thermogenic carbon release from contact metamorphism represents a plausible source for ¹²C, however this has not yet been explored from a carbon cycle approach. This study explores the effects of thermogenic carbon release from CAMP using carbon cycle modeling, and shows that it represents a credible source for the negative CIEs at the end-Triassic. It strengthens the hypothesis that the sub-volcanic part of a Large Igneous Province is of major importance for understanding carbon cycle disruptions.
Introduction

The end-Triassic extinction (ETE) was one of the largest mass extinctions in Earth history. The ETE coincided with several major carbon cycle disturbances, as evidenced by increases in atmospheric CO$_2$ concentrations$^{1-3}$ and negative carbon isotope excursions (CIEs) recorded in both carbonate and organic matter records$^{4-16}$. The first disruption of the carbon cycle is marked by the late Triassic Marshi/Precursor CIE (hereby referred to as CIE 1), followed by the Spelae/Initial CIE around the T-J boundary (CIE 2), and finally the Tilmanni/Main CIE in the earliest Jurassic (CIE 3). As the ETE coincided with the onset of the Central Atlantic Magmatic Province (CAMP), it has been hypothesized that CAMP-derived emissions were responsible for the biotic crisis and contributed to the carbon cycle perturbations$^{1-26}$. Previous carbon cycle models for the Triassic-Jurassic$^{22-24}$ suggest that the release of mantle-derived carbon from CAMP cannot solely account for the negative CIEs associated with the ETE. Alternative suggestions include instead marine methane clathrate dissociation induced by CAMP volcanism$^{22-23}$ or volcanic degassing of carbon with an anomalous $\delta^{13}$C value of $-20 \%_o$.$^{24}$

A recent study suggested that 24,000 Gt C could have been generated during CAMP sill emplacement and contact metamorphism of marl and organic-rich sedimentary rocks of the Amazonas and Solimões basins in Brazil$^{21}$. High precision U-Pb dating demonstrates that these sills were emplaced synchronously with the ETE$^{20-21}$. It has been suggested that thermogenic carbon release associated with CAMP emplacement in volatile-rich sedimentary basins could explain the negative CIEs$^{10,13,20-21}$, but this has not yet been explored from a carbon cycle modeling approach. Furthermore, previous carbon cycle models have mainly focused on the Spelae excursion (CIE 2), which was interpreted to coincide with the ETE$^{5,7,14,18}$. However, stratigraphic records demonstrate that major biotic perturbations took place prior to this CIE$^{8,25-26}$, and recent correlations, based on a variety of methods including palynology, ammonite
stratigraphy, δ$^{13}$C$_{org}$ data, and radiometric dating, suggest that the onset of the ETE coincides with the earliest Marsh excursion (CIE 1)$^9$ (Fig. S1A).

In order to explore whether release of thermogenic carbon generated from sub-surface contact metamorphism of sedimentary rocks could have contributed to the observed T-J boundary carbon cycle perturbations, we tested CAMP carbon emission scenarios using the Long-Term Ocean-Atmosphere-Sediment Carbon Cycle Reservoir Model (LOSCAR)$^{27}$. Our approach identifies realistic emission scenarios by replicating observed $p$CO$_2$ and δ$^{13}$C curves, grounded by numerous complementary boundary conditions. These include high precision U-Pb geochronology, the geology and evolution of CAMP, and gas release from both lavas and contact metamorphic sedimentary rocks.

**Timing of events**

High-precision U-Pb dating of extrusive and intrusive CAMP rocks demonstrates a total duration of CAMP activity of up to 800 kyr$^{18-21}$, which is interpreted to occur in four main phases (Fig. 1a). CAMP intrusive and extrusive rocks are traditionally defined as low- or high-Ti, following a threshold of 2.0 wt.% TiO$_2$$^{28}$. There are four available U-Pb ages of sills in the Amazonas and Solimões basins, including two low-Ti (201.525 ± 0.065; 201.470 ± 0.089 Ma) and two high-Ti (201.477 ± 0.062; 201.364 ± 0.023 Ma) sills$^{20-21}$. The low-Ti sills and one high-Ti sill overlap with the ‘main’ CAMP phase, while the second high-Ti sill is coincident with the onset of the ‘late’ phase (Fig. 1a). Based on recent data from the Amazonas Basin, high-Ti sills are exclusively limited to the lower Paleozoic series$^{21,29}$, which includes organic-rich shales (up to 8 wt.% TOC$^{30-31}$) but lacks the marls that are present in upper Paleozoic strata. Therefore, the later pulse of high-Ti magmatism is likely to have volatilized predominantly organic-derived carbon, while the earlier sill intrusions into upper Paleozoic shales and marls generated both organic- and inorganic-derived carbon$^{21}$. 
The correlation of CAMP activity and $\rho$CO$_2$ changes are constrained by data from the eastern USA. Significant $\rho$CO$_2$ increases are recorded in sediments above CAMP lava flows in the Newark (Orange Mountain, Preakness, and Hook Mountain basalts) and Hartford (Talcott, Holyoke, and Hampden basalts) basins$^{2-3}$. Flow units and sedimentary sequences can be directly correlated between the two basins$^3$, and U-Pb ages are available for the Newark lavas (or intrusive equivalents$^{19}$). These geochronological constraints suggest that the oldest lavas (Orange Mountain/Talcott) belong to the ‘main’ CAMP phase, which overlaps with the earliest Marshi CIE 1 according to recent correlations$^9$ (Fig. 1b; SI Text). The duration of the ‘main’ phase corresponds to the time span between the two low-Ti Brazilian sills (Fig. 1a). This suggests that thermogenic carbon generated during contact metamorphism was released simultaneously with the mantle-derived carbon from these CAMP lavas.

A second and third increase in $\rho$CO$_2$ is recorded following the Preakness Basalt and Hook Mountain/Hampden basalts, respectively$^{2-3}$. While the Preakness basalt correlates with the onset of the latest Tilmanni excursion (CIE 3) in the earliest Jurassic$^9$, U-Pb geochronology demonstrates (based on dating of the Butner diabase$^{19}$) that the Hook Mountain/Hampden basalts post-date all three CIEs (Fig. 1b). The youngest high-Ti Brazilian sill overlaps with the Spelae excursion (CIE 2)$^9$ (Fig. 1b), but a $\rho$CO$_2$ peak potentially accompanying this CIE is not recorded in the Newark/Hartford basins (SI Text). This sill pre-dates the remainder of the ‘late’ CAMP phase (not taking the large error bars of the Shelburne dyke into account; Fig. 1a). This suggests that thermogenic carbon generated during contact metamorphism was released slightly prior to mantle-derived carbon from the ‘late’ CAMP lavas.

Note that a fourth marked increase in $\rho$CO$_2$ is recorded in the Hartford Basin$^3$, but as it post-dates any known CAMP event and the ETE, this is not taken into account in the model. Also, as the majority of CAMP rocks representing the ‘early’ CAMP phase predate the Orange Mountain/Talcott basalts (marking the first observed peak in $\rho$CO$_2$), carbon release from this
phase is not added in any emission scenario modeled here. We stress that there are alternative correlations and interpretations for the T-J boundary carbon cycle perturbations\textsuperscript{5-7,14,18} that differ from those used here (Fig. S2). We present model runs using these correlations and boundary conditions in the SI Text (Table S1; Fig. S3).

**Results**

Carbon pulse 1, representing the ‘main’ CAMP phase, is released at model time \( t = 0 \). This pulse is interpreted to include both mantle-derived and inorganic- and organic-derived thermogenic carbon (\( \delta^{13}C \) of -13 \(^\circ\); see SI Text). The duration is set to 50 kyr, based on the minimum span in U-Pb ages of the three oldest Brazilian sills\textsuperscript{20-21}. By releasing 8,800 Gt C (Table 1), the \( p\text{CO}_2 \) increases to \( \sim 3,800 \) ppm, and the \( \delta^{13}C \) of shallow ocean sediments decreases by \( \sim 2.0 \% \) (Fig. 2), which is within the range of the magnitude of the Marshi CIE 1 (SI Text).

Carbon pulse 2, characterizing the second pulse of high-Ti magmatism, is released at model time \( t = 175 \) kyr, a representative value considering the time span between the oldest low-Ti sill from the ‘main’ phase and the youngest high-Ti sill (\( \sim 70 \) to 250 kyr). Carbon pulse 2 is thus interpreted to include organic-derived thermogenic carbon only (i.e. \( \delta^{13}C \) of -41 \(^\circ\); SI Text). The duration is set to 20 kyr, which is considered a realistic duration for a sill emplacement pulse\textsuperscript{33-34} and is within estimates of the duration of the negative excursion targeted\textsuperscript{6,15-16}. By releasing an equal amount of thermogenic carbon as modeled for carbon pulse 1 (i.e. 4,800 Gt; Table 1), the \( p\text{CO}_2 \) increases to \( \sim 3,700 \) ppm, and the \( \delta^{13}C \) of shallow ocean sediments decreases by \( \sim 3.5 \% \) (Fig. 2). Again, this correlates well with the magnitude of the Spelae CIE 2 (SI Text). As previously mentioned, a \( p\text{CO}_2 \) peak potentially associated with the Spelae CIE 2 is not recorded by the Newark/Hartford sediments when considering the correlations by ref. 9.
In the absence of any available evidence for sills emplaced during the Tilmanni CIE 3, the third carbon pulse in our model should be purely mantle-derived. However, in order to replicate a drop in $\delta^{13}C$ within the range of the Tilmanni CIE 3 with mantle-derived carbon (i.e. $\delta^{13}C$ of -5 ‰), a magnitude of ~ 15,000 Gt C is required (Fig. S4). As this value exceeds most estimates of cumulative mantle-derived carbon release from CAMP2-3,22,35,36 (Fig. S5), we consider it to be an unrealistically high value for a single pulse. While there might have been alternative processes affecting carbon isotopes at this time (e.g. methane clathrates), for the purposes of this study we assume that thermogenic carbon was responsible for the Tilmanni CIE 3. However, mantle-derived carbon from the ‘late’ CAMP phase must also be taken into account, and we assume a pulse of mantle-derived carbon of equal magnitude to carbon pulse 1 (i.e. 4,000 Gt; Table 1). Carbon pulse 3 is released directly following carbon pulse 2 at model time $t = 195$, and lasts until model time $t = 265$ kyr, which corresponds to the U-Pb age of the Preakness/Holyoke basalts. At $t = 265$ kyr, carbon pulse 4 is released, which includes 2,500 Gt organic-derived thermogenic carbon (i.e. $\delta^{13}C$ of -41 ‰; SI Text). With this release, the $pCO_2$ increases to ~ 2,900 ppm, and the $\delta^{13}C$ of shallow ocean sediments decreases by ~ 2.0 ‰ (Fig. 2), which is within the range of the magnitude of the Tilmanni CIE 3 (SI Text).

Following the previous pulses of mantle-derived carbon, carbon pulse 5, which represents the ‘final’ CAMP phase, is given a magnitude of 4,000 Gt C and is released at model time $t = 623$ kyr (SI Text). This corresponds to the age of the Hook Mountain/Hampden basalts based on U-Pb dating of the Butner diabase19. The modeled $pCO_2$ increases to ~ 3,300 ppm, while the $\delta^{13}C$ of shallow ocean sediments decreases by ~ 0.5 ‰ (Fig. 2).

**Discussion**

According to our emission scenario, a total release of ~ 24,000 Gt C (from both magmatic and contact metamorphic sources) can replicate the observed $pCO_2$ and $\delta^{13}C$ data around the T-J
boundary. The modeled \( p\text{CO}_2 \) generally plots between the minimum and mean values for the reconstructed \( p\text{CO}_2 \) (Fig. 2), which suggests that this value may be a conservative estimate. Nevertheless, it is still significantly higher than most previous estimates of released mantle-derived carbon from CAMP\(^2\text{,}3\text{,}22\text{,}23\text{,}35\text{,}36\) (Fig. S5). This suggests that additional sources, such as thermogenic carbon or methane clathrates, were involved. It is plausible that previously published magmatic carbon fluxes are underestimated (SI Text). However, our modeling further demonstrates that the release of mantle-derived carbon alone is unlikely to have generated the observed negative CIEs (Fig. S4, S6, S7), in agreement with previous carbon cycle modeling\(^{22\text{,}24}\).

The minimum \( \delta^{13} \text{C} \) value given to those pulses comprising thermogenic carbon in our model (i.e. pulses 1, 2 and 4) is -41‰ (Table 1; SI Text). Therefore, an extremely isotopically depleted carbon source, such as biogenic CH\(_4\) (-70 ‰\(^{37}\)), is not needed to replicate the negative CIEs. There is currently no U-Pb data linking sill emplacements in Brazil with the Tilmanni CIE 3 (considering the correlations by ref. 9), but only four U-Pb ages are available. A later thermogenic release coincident with this CIE can therefore not be ruled out, and the modeling constraints indicate that isotopically light carbon is needed to replicate all of the CIEs. Therefore, we propose a scenario where three main pulses of emplacement in the Brazilian basins generated and released sufficient volumes of gases. These include an initial release of organic- and inorganic-derived thermogenic carbon, followed by a second and third release of predominantly organic-derived thermogenic carbon (Fig. 3).

The total magnitude of thermogenic carbon released in the model (12,100 Gt) represents around half of the estimated magnitude of carbon gases generated in the Brazilian basins (i.e. 24,000 Gt\(^{21}\)). We stress that the ratio of trapped to released gases from volcanic basins is poorly constrained\(^{38}\). Considering that the majority of petroleum reservoirs in the Solimões Basin are linked to sill emplacements\(^{30\text{,}32\text{,}39\text{,}40}\), a value of 50 % might represent an overestimation of the
gas release. However, the value of 24,000 Gt C does not include carbon generation from contact metamorphism of lower Paleozoic shales (TOC up to 8 wt.%)\textsuperscript{30-32}, nor interactions between the sills and pre-existing hydrocarbon accumulations in the basins\textsuperscript{21,41}. The Amazonas Basin was petroleum-bearing prior to sill emplacement, which likely led to enhanced gas generation\textsuperscript{30-31,39}. The proposed release of 24,000 Gt C is therefore likely underestimated, and consequently, the emission of 12,100 Gt C represents a lower percentage of released versus trapped carbon than 50%.

It is important to assess whether the magnitudes of the individual thermogenic carbon pulses are realistic. Within the upper Paleozoic series, sills can be traced almost continuously between the margins of the Amazonas and Solimões basins, whereas sill emplacement into the lower Paleozoic unit is restricted to the eastern Amazonas Basin\textsuperscript{21,29-32,39-40}. This may suggest that the total gas generation within the upper Paleozoic series (represented by carbon pulse 1) was higher compared to the lower Paleozoic series (represented mainly by carbon pulses 2 and 4). However, as the maximum TOC content in lower Paleozoic rocks is significantly higher compared to those of the upper Paleozoic series (8 vs. < 1 wt.%, respectively\textsuperscript{30}), more carbon-bearing gases could have been generated despite the fact that sills are less widespread. Furthermore, closely spaced sills affect the same sedimentary rocks, thus a later sill emplacement pulse would generate and release less carbon than the preceding one, if the same portion of the basin is affected\textsuperscript{21}. This likely applies for the Brazilian basins considering the localized sill distribution within the lower Paleozoic series (Fig. 3). We therefore consider an identical magnitude for pulses 1 and 2 (4,800 Gt), and a reduction for pulse 4 (2,500 Gt) to be in close agreement with the available data.

Each of the three individual mantle-derived carbon pulses are set to 4,000 Gt, so that the total magnitude released represents a mean of previous estimates constrained by several methods\textsuperscript{2-3,22,35-36} (Fig. S5). The bulk of the dated CAMP rocks overlap with the ‘main’ phase,
whereas there are few CAMP rocks with ages corresponding to the ‘final’ phase (Fig. 1). However, the Newark and Hartford records demonstrate a similar magnitude of $pCO_2$ increase directly following both the Orange Mountain/Talcott (‘main’ phase) and Hook Mountain/Hampden basalts (‘final’ phase)\textsuperscript{2,3}. The present-day CAMP is dominated by sills and dykes, while preserved lava flows are rare, showing that large parts of the lavas have been removed through erosional processes. Estimating the pre-erosional volume of the province (and other LIPs) is challenging (SI Text). Therefore, it is difficult to constrain whether the magnitudes of the individual carbon pulses are proportional with the igneous volumes for the different CAMP phases.

The T-J boundary records show marked decreases in $pCO_2$ subsequent to each $pCO_2$ peak, with positive $\delta^{13}C$ intervals following the three negative CIEs\textsuperscript{4-16}. Our modeling demonstrates that significant organic carbon burial, in addition to silicate weathering, is required in order to replicate the sharp positive $\delta^{13}C$ intervals following the Marshi CIE 1 and Spelae CIE 2 (Fig. S8; SI Text). This is in agreement with previous carbon cycle modeling for the T-J boundary, and consistent with widespread anoxia and deposition of organic-rich sediments observed for the earliest Jurassic\textsuperscript{23}. It is unlikely that the positive $\delta^{13}C$ intervals following the negative CIEs represent the release of $^{13}C$-enriched carbon (e.g. inorganic thermogenic C), as the Newark and Hartford records demonstrate significant $pCO_2$ decrease following each peak\textsuperscript{2,3}. In contrast, the decreasing $pCO_2$ following the Hook Mountain/Hampden basalts and the prolonged period of increasing $\delta^{13}C$ following the Tilmanni CIE 3, which eventually return to pre-CAMP values, can be replicated by purely taking silicate weathering into account (SI Text), in agreement with previous geochemical modeling\textsuperscript{3}.

We conclude that CAMP-related volatile release can explain the carbon cycle perturbations observed around the T-J boundary. Sill heating and contact metamorphism of carbon-bearing sedimentary rocks in the Amazonas and Solimões basins represents a likely
source of $^{13}$C-depleted carbon, which can explain the observed negative CIEs. Although the exact values for the different degassing pulses represent estimates, we consider our emission scenario realistic as it incorporates the available geological knowledge of the CAMP and a realistic framework for Earth system processes during the latest Triassic and earliest Jurassic.

**Materials and Methods**

**LOSCAR modeling**

The Long-term Ocean-atmosphere-Sediment Carbon cycle Reservoir Model (LOSCAR) is a box model that simulates the cycling of carbon through atmospheric, oceanic, and sediment reservoirs on timescales from centuries to millions of years\(^\text{27}\). The model has a ‘modern’ and ‘paleo’ version, which includes three and four ocean boxes, respectively (Atlantic, Indian, Pacific ± Tethys oceans). The oceans are separated into surface (0-100 m), intermediate (100-1,000 m) and deep (> 1,000 m) boxes, as well as one high-latitude box (cold surface waters with no specific location). All ocean boxes (except the high-latitude box) are coupled to sediment boxes, and the surface ocean boxes are coupled to one atmospheric box.

The LOSCAR model is not designed specifically for the end-Triassic, and default settings do not necessarily represent T-J boundary conditions. For example, the end-Triassic paleogeography was dominated by the Panthalassic Ocean, as opposed to three or four ocean basins in the LOSCAR set up. We used default parameter settings of the LOSCAR ‘paleo’ version, and modified selected background initial conditions to better match a late Triassic steady state (Table S2). Carbonate production during the end-Triassic/early-Jurassic was likely restricted to surface epicontinental seas and shelves, which represent only a small part of the global ocean\(^\text{24,42-44}\). Following LOSCAR modeling applied for the end-Permian\(^\text{45}\), with comparable paleogeography as the end-Triassic, carbonate precipitation was set to take place predominantly over the surface ocean sediment boxes (i.e. increasing parameter FSHLF; Table
In addition, all low-latitude ocean boxes were given identical initial conditions. As seen in Fig. S9, there is little variation in modeled δ¹³C values for shallow ocean boxes between the different oceans (i.e. Atlantic, Indian, Pacific, and Tethys), suggesting that the four ocean boxes collectively could represent one major epicontinental sea or extended shelf area. Accordingly, presented δ¹³C model results represent therefore mean values of the four surface ocean sediment boxes. Furthermore, the carbon emission scenario presented in Table 1 was tested with an alternative carbon cycle model that includes paleogeography of the T-J boundary (GEOCLIM⁴⁶; SI Text). When comparing the LOSCAR vs. GEOCLIM runs, there is a close correlation in model behaviors (Fig. S10), suggesting that the LOSCAR results are robust and representative of the observed proxy data.

Other changes from the ‘paleo’ version default settings performed to better match end-Triassic conditions, included increasing the temperature for all low-latitude surface ocean boxes to 32 ºC⁴⁷-⁴⁸, and values for dissolved Ca and Mg to 17 and 32 mM respectively⁴⁹. Following several pCO₂ reconstructions for the end-Triassic²-³,⁵⁰-⁵², the initial atmospheric pCO₂ (prior to the first carbon injection from CAMP) was set to 2,000 ppm. For the remaining parameters, default settings were used (Table S2). All data discussed in the paper will be made available to readers.

Data availability

The LOSCAR source code is available for download at https://www.soest.hawaii.edu/oceanography/faculty/zeebe_files/LoscarModel.html. All data used for this paper is available in the main text and in the Supplementary Information.

Acknowledgements
This work was supported by the Research Council of Norway through its Centres of Excellence funding scheme (project number 223272), and partly through funding to The Norwegian Research School on Dynamics and Evolution of Earth and Planets (project number 249040/F60). MTJ is funded by the Research Council of Norway, project number 263000. Richard Zeebe and Yves Goddéris are thanked for their participation in discussions regarding the manuscript. We also wish to thank the Editor and three anonymous reviewers, whose contributions greatly improved the manuscript.

References


38Jones, M. T., Jerram, D. A., Svensen, H. H., & Grove, C. The effects of large igneous


**Figure 1**

A) Compilation of available high-precision U-Pb ages of CAMP rocks\textsuperscript{18-21} demonstrates a total duration of ~ 800 kyr and potentially four main phases of CAMP activity (‘early’, ‘main’, ‘late’, and ‘final’). The three oldest high- and low-Ti Brazilian sills overlap with the main phase, while the youngest high-Ti sill marks the onset of the ‘late’ phase. Note that the age for the North Mountain Basalt represents an average value after refs. 18-20.

B) Timeline of selected CAMP rocks from the Amazonas, Solimões and Newark basins, a composite carbon isotope curve\textsuperscript{9} and $p$CO$_2$ data\textsuperscript{2} spanning the T-J boundary after ref. 9. See *Fig. S1* for detailed T-J boundary $\delta^{13}$C$_{carb}$ and $\delta^{13}$C$_{org}$ curves. The ‘main’ CAMP phase, the latest high-Ti Brazilian sill and the Preakness Basalt coincide with the Marshi (‘M’; CIE 1), Spelae (‘S’; CIE 2) and Tilmanni (‘T’; CIE 3) negative excursions, respectively. PAL = Palisades Sill. PRB = Preakness Basalt. BUT = Butner diabase.

**Figure 2**

Model response of atmospheric $p$CO$_2$ (A) and $\delta^{13}$C of shallow ocean sediments (B; mean value of Atlantic, Indian, Pacific and Tethys shallow ocean boxes) to a CAMP emission scenario including five pulses of carbon release (see *Table 1*). The increases in $\delta^{13}$C following each negative excursion and decreases in $p$CO$_2$ after each $p$CO$_2$ peak reflect organic carbon burial and/or silicate weathering (see *Fig. S8*). Gray outlines/star symbols represent the range of
observed $\delta^{13}C_{\text{carb}}$ values$^{4,11-12,72}$ (see SI Text) and $pCO_2$ data$^{2-3}$ (light gray: ref. 2; dark gray: ref. 3).

**Figure 3**

Interpretation of CAMP sill emplacement and thermogenic carbon release in three main phases in the Amazonas and Solimões basins, Brazil. The first phase of sill emplacement includes both low- and high-Ti sills intruding the upper and lower Paleozoic series, releasing 4,800 Gt of mixed (50/50) inorganic and organic thermogenic carbon. The second and third phases include high-Ti sills intruding the lower Paleozoic series, releasing 4,800 and 2,500 Gt organic thermogenic carbon, respectively. The cross-section is not drawn to scale.
## Table 1
Overview and input values for CAMP emission scenario

<table>
<thead>
<tr>
<th>Carbon pulse #</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Magnitude (Gt)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Volcanic C</td>
<td>4000</td>
<td>0</td>
<td>4000</td>
<td>0</td>
<td>4000</td>
</tr>
<tr>
<td>Inorganic thermogenic C</td>
<td>2400</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Organic thermogenic C</td>
<td>2400</td>
<td>4800</td>
<td>0</td>
<td>2500</td>
<td>0</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td>8800</td>
<td>4800</td>
<td>4000</td>
<td>2500</td>
<td>4000</td>
</tr>
</tbody>
</table>

| δ¹³C (%)   |    |    |    |    |    |
| Volcanic C | -5 | n/a | -5 | n/a | -5 |
| Inorganic thermogenic C | 0  | n/a | n/a | n/a | n/a |
| Organic thermogenic C | -41 | -41 | n/a | -41 | n/a |
| **Total**   | -13 | -41 | -5 | -41 | -5 |

| Model time t (ky) | 0 | 175 | 195 | 265 | 623 |
| Duration (ky)     | 50 | 20  | 70  | 20  | 20  |

n/a = not applicable