Upper limits on the extent of seafloor anoxia during the PETM from uranium isotopes

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The Paleocene Eocene Thermal Maximum (PETM) represents a major carbon cycle and climate perturbation that was associated with ocean de-oxygenation, in a qualitatively similar manner to the more extensive Mesozoic Oceanic Anoxic Events. Although indicators of ocean de-oxygenation are common for the PETM, and linked to biotic turnover, the global extent and temporal progression of de-oxygenation is poorly constrained. Here we present carbonate associated uranium isotope data for the PETM. A lack of resolvable perturbation to the U-cycle during the event suggests a limited expansion of seafloor anoxia on a global scale. We use this result, in conjunction with a biogeochemical model, to set an upper limit on the extent of global seafloor de-oxygenation. The model suggests that the new U isotope data, whilst also being consistent with plausible carbon emission scenarios and observations of carbon cycle recovery, permit a maximum ~10-fold expansion of anoxia, covering <2% of seafloor area.
The Paleocene Eocene Thermal Maximum (PETM) is one of the best studied climate perturbation events in Earth’s history, but the driving mechanisms, environmental consequences, and recovery processes are still debated. The PETM is characterized by the input of isotopically light carbon, probably representing a mixture of mantle-derived carbon from large igneous province (LIP) activity and oxidation of organic carbon reservoirs\textsuperscript{3–4}. These emissions resulted in warming of ~5°C (ref. \textsuperscript{7}) and a negative carbon isotope (δ\textsuperscript{13}C) excursion (CIE) of ~3–4‰, associated with rapid environmental deterioration including ocean acidification\textsuperscript{5}, changes in hydrology and weathering regimes\textsuperscript{6}, alteration of ocean circulation\textsuperscript{8}, locally enhanced primary productivity\textsuperscript{9,10}, and local de-oxygenation of the water column\textsuperscript{11–15}.

There is increasing recognition that expanded oceanic anoxia is a defining consequence of past global warming events\textsuperscript{16}, and that it is predicted to worsen under future climate projections\textsuperscript{17}. Indicators of local de-oxygenation are common for the PETM, with the development of low-O\textsubscript{2} conditions in open ocean locations and fully euxinic (anoxic and sulfidifying) environments in some continental shelf and restricted basin sites\textsuperscript{7,11–15,18–20}. Water column de-oxygenation is often invoked as a contributor to benthic faunal turnover and extinction patterns\textsuperscript{21,22}. Various mechanisms have been proposed to drive de-oxygenation during the PETM, including changes in deep ocean and basin ventilation, upwelling strength, temperature solubility effects, methane hydrate destabilization, and enhanced nutrient inputs from the continents\textsuperscript{9,22–24}. Many of the characteristics and hypotheses for the PETM are shared with the Mesozoic Oceanic Anoxic Events (OAEs), allowing them to be compared for insights into the driving mechanisms of environmental catastrophe but also for understanding negative feedbacks that counter greenhouse gas forcing and enable climate recovery, such as silicate weathering and organic carbon burial\textsuperscript{1,16,25}.

Despite extensive study, the overall global extent and severity of anoxia during the PETM is poorly quantified, making it difficult to meaningfully compare it to other periods of oceanic anoxia. This limitation is largely due to reliance on local redox proxies, which act as targets for spatially resolved Earth system models\textsuperscript{15,20}, but yield patchy geographical information and conflicting interpretations\textsuperscript{7}. Global-scale geochemical insights, from molybdenum (Mo) isotopes, suggest that euxinia was more prevalent during the PETM than the modern\textsuperscript{11} but less extensive than during Mesozoic OAEs\textsuperscript{26}. Sulfur (S) isotope evidence has also been used to model a 10–20-fold increase in the volume of euxinic waters (with up to 0.5 mM H\textsubscript{2}S) during the PETM compared to the modern ocean\textsuperscript{27}. Whilst these studies provide important insights into general redox characteristics, neither capture non-euxinic redox changes that are crucial for understanding faunal responses and biogeochemical feedbacks during climate perturbations.

To overcome these limitations, we present the first carbonate-associated uranium isotope data (238U/235U, reported as δ\textsuperscript{238}U\textsubscript{CAU}, see “Methods”) for the PETM interval. Uranium solubility and isotope fractionation are strongly redox sensitive (commonly as U(VI) to U(IV)), with the preferential sequestration of 238U, relative to 235U, in reducing sediments\textsuperscript{28,29}. Uranium reduction primarily occurs at or below the sediment–water interface, at a redox potential similar to that for iron reduction\textsuperscript{30}. As such, authigenic U(IV) accumulation is a result of bottom-water and pore-water anoxia, but does not require euxinic conditions within the water column. With a long residence time in the modern ocean (~320–560 kyrs\textsuperscript{31}), δ\textsuperscript{238}U can trace global-scale changes in benthic anoxia\textsuperscript{26}; events such as OAE 2 (ref. \textsuperscript{32}) and the Permo-Triassic Boundary\textsuperscript{33,34} are associated with clear shifts to lower δ\textsuperscript{238}U\textsubscript{CAU} values that are thought to reflect trends in seawater δ\textsuperscript{238}U (δ\textsuperscript{238}U\textsubscript{sw}) and can be quantitatively related to the expansion of anoxic sinks.

New δ\textsuperscript{238}U\textsubscript{CAU} and U/Ca data are presented for bulk carbonate leachates (see “Methods”) from three geographically dispersed pelagic localities that span the PETM: ODP Site 865 from Allison Guyot, equatorial Pacific Ocean (paleodepth ~1500 m), DSDP Site 401 on the continental slope of the Bay of Biscay (~2000 m), and ODP Site 690 from Maud Rise in the Southern Ocean (~2100 m)\textsuperscript{35–37}. Samples are poorly lithified carbonate-rich sediments with a mixture of benthic and pelagic foraminifera, or carbonate nannofossil ooze, and show variable contributions from detrital clays\textsuperscript{35–37}. The intermediate-depth equatorial Pacific is thought to have been mostly oxygenated before and during the PETM\textsuperscript{14,20}, and δ\textsuperscript{238}U\textsubscript{CAU} is likely to reliably record δ\textsuperscript{238}U\textsubscript{sw}. By contrast, the other two sites experienced varying degrees of de-oxygenation with observable effects on local U systematics\textsuperscript{13,20}.

The δ\textsuperscript{238}U\textsubscript{CAU} from Site 865 and published δ\textsuperscript{13}C data\textsuperscript{3} provide constraints for an established carbon–phosphorus–uranium (C–P–U) biogeochemical box model\textsuperscript{32}, based on the more comprehensive COPSE model\textsuperscript{38}, to quantify the extent of seafloor anoxia during the PETM.

Results

Trends in U/Ca and U isotopes. All three sites exhibit stable background U/Ca that increases around the onset of the PETM (Fig. 1 and Supplementary Figs. 1–3). For Site 865, the increase defines a minor peak, from 0.02 to 0.05 µmol/mol, beginning ~20 kyrs before the PETM onset and extending to ~25 kyrs after (see “Methods” for age model details). Lower U/Ca values are seen for the remainder of the record, except for a few samples (up to ~0.1 µmol/mol) in the shallowest part of the core. For Site 401, U/Ca increases from 0.05 to 0.27 µmol/mol at the PETM onset and remains elevated for the next ~120 kyrs before decreasing again for the remainder of the record, but remaining above pre-PETM values. Larger U enrichments are seen at Site 690, where U/Ca increases from a background of ~0.05 µmol/mol to over 1 µmol/mol. At Site 690, the U enrichments occur in two main phases, firstly ~100 kyrs prior to the PETM onset, and, more significantly, later during the PETM recovery interval.

Sites 865 and 401 both have largely invariant δ\textsuperscript{238}U\textsubscript{CAU} Values of ~0.36 ± 0.06‰ (2 SD), before and after the PETM interval. Site 865 shows no resolvable variation in δ\textsuperscript{238}U\textsubscript{CAU} for the entirety of the record, except for a single anomalously low sample at ~0.58‰, and there is no change in δ\textsuperscript{238}U\textsubscript{CAU} related to the minor peak in U/Ca (Fig. 1). Higher U/Ca at Sites 401 and 690 are positively correlated with δ\textsuperscript{238}U\textsubscript{CAU} (Fig. 1). Additionally, Site 690 shows lower pre-PETM values than the other sites and a return to these low values between intervals of higher U/Ca.

The relationship of δ\textsuperscript{238}U\textsubscript{CAU} to seawater. Pre-PETM U/Ca are similar to modern bulk pelagic carbonates but higher than modern foraminifera\textsuperscript{39,40}. These U/Ca match better with the higher partition coefficient of abiotic calcite compared to biogenic calcite\textsuperscript{39}, suggesting that the majority of carbonate-associated U resides in syn-sedimentary authigenic cements\textsuperscript{40}. Sites 865 and 401 demonstrate remarkable invariance and reproducibility in pre-PETM δ\textsuperscript{238}U\textsubscript{CAU}, with an average of ~0.36 ± 0.06‰ (2SD) that is similar to modern seawater (~0.39 ± 0.01‰; ref. \textsuperscript{28}). The simplest explanation for this uniformity is that these sites directly record contemporary δ\textsuperscript{238}U\textsubscript{sw}, with little or no associated isotopic fractionation, as shown for some modern biogenic and abiotic calcites\textsuperscript{41–43}. This implies that, under well-oxygenated conditions, U is incorporated into the carbonate lattice of syn-sedimentary cements as U(VI) whilst in communication with overlying seawater, and is not influenced by reducing conditions during early
The precise extent of U isotope fractionation into authigenic carbonate depends on the U(VI) speciation in the aqueous phase, which may lead to $\delta^{238}U_{\text{CAU}}$ up to ~0.1‰ heavier than seawater\(^4\), so it is possible that Paleocene $\delta^{238}U_{\text{sw}}$ could have been lower, down to ~$-0.46$‰. Site 690 appears to also show invariant baseline values before the PETM, but with a lower $\delta^{238}U_{\text{CAU}}$ of $-0.64$‰. This difference might reflect a higher proportion of adsorbed (isotopically light) U in the authigenic carbonate at Site 690, such as has been inferred for a modern pelagic bulk carbonate sample\(^{40}\), or U associated with planktic organic matter\(^{44}\) that could be incorporated into authigenic calcite. The same explanation could also apply to the single anomalously low value at Site 865.

**Diagenetic influences.** The deep-sea core samples used for this study have experienced relatively simple diagenetic histories compared to carbonates used in the majority of paleo-studies using U isotopes, which to date have focused mainly on platform carbonates and/or outcrop material\(^{32,33,45-47}\). In the case of platform carbonates, that originally consisted of metastable aragonite or high-Mg calcite, diagenetic corrections are often applied to $\delta^{238}U_{\text{CAU}}$ values\(^{33,45-47}\) to account for authigenic U(IV) uptake under reducing porewater conditions during early diagenesis (including recrystallization and cementation) that results in a high degree of scatter with a generally positively offset compared to seawater ($+0.27 \pm 0.14$‰; 1 SD)\(^{48,49}\). Previous work has suggested that some pelagic carbonates record lower $\delta^{238}U_{\text{CAU}}$ signatures, closer to seawater values compared to platform counterparts, and thus do not require correction\(^{32,41}\). Similarly, biogenic calcite from brachiopod shells is thought to better retain primary $\delta^{238}U_{\text{sw}}$ signatures than metastable matrix components\(^{43}\). The better preservation of $\delta^{238}U_{\text{sw}}$ in pelagic carbonate samples is likely due to the stable mineralogy of these sample types (low-Mg calcite), but also the generally low organic carbon accumulation rates and low fluid flow rates of pelagic settings\(^{43}\). Here, increases in $\delta^{238}U_{\text{CAU}}$ and U/Ca at Sites 690 and 401, that likely reflect early diagenetic U(IV) uptake, are systematically tied to local environmental change related to the PETM hyperthermal (see below). Such patterns therefore reflect an integration of bottom-water and pore-water conditions, carrying information on local environmental deterioration, rather than a constant diagenetic process that requires correction. Given the correspondence of pre-PETM $\delta^{238}U_{\text{CAU}}$ to near modern seawater values, we do not apply any diagenetic correction factor in order to reconstruct seawater compositions.

The main concern for the integrity of the presented datasets is for Site 865, where the planktic foraminifera are known to have undergone recrystallization, resulting in opaque calcite\(^{50}\), although the benthic species are well preserved\(^{51}\). Geochemical

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**Fig. 1** Uranium isotope and U/Ca data for pelagic PETM localities. **a** Bulk carbonate $\delta^{13}C$ data from Site 401 (blue line; ref. 1) and Site 690 (red line; ref. 4), and A. soldadoensis $\delta^{13}C$ for Site 865 (black circles; ref. 53). All data are plotted as a function of time relative to the onset of the carbon isotope excursion (CIE) using the $^3$He-based age model\(^6\) (see “Methods”). **b** U/Ca from carbonate leachates corresponding to $\delta^{238}U_{\text{CAU}}$ plotted in **c**, with symbols colored by site as in **a**. Note a log scale is used in **b**. U/Ca are not shown for samples run at the Open University (OU; gray filled circles) due to incomplete carbonate digestion (see “Methods”). Error bars for $\delta^{238}U_{\text{CAU}}$ (**c**) represent the external reproducibility of the NIST-SRM 1d limestone standard ($\pm 0.06$‰, 2 SD, $n = 14$; see “Methods”). PETM $\delta^{238}U_{\text{CAU}}$ data are compared to average modern $\delta^{238}U_{\text{sw}}$ in **c** ($-0.39 \pm 0.01$; blue shading)\(^{28}\).
alteration of the planktic foraminifera resulted in a temperature-dependent offset in planktic foraminiferal $\delta^{18}O$, whilst $\delta^{13}C$ of individual planktic specimens approach the bulk carbonate signature\textsuperscript{52} but still capture the structure of the PETM CIE\textsuperscript{53}. By contrast, boron (B) isotopes ($\delta^{11}B$) appear to be well preserved\textsuperscript{52,54} but B/Ca is lower in recrystallized specimens, suggesting that although B is lost during recrystallization the remaining calcite-bound B did not exchange with diagenetic pore fluids\textsuperscript{52}. Given the immobility of U in deeper (anoxic) burial environments, and the relatively closed nature of recrystallization at Site 865 (ref. \textsuperscript{52}), we would expect a similar preservation of primary U isotope signatures\textsuperscript{41}. Regardless of any planktic foraminiferal recrystallization, it is important to recognize that the measurements presented here were performed on bulk carbonate leachates where the majority of U is thought to be present in syn-sedimentary cements. Thus, any recrystallization of the planktic foraminifera, that contribute only a fraction of the U mass balance, should have little impact on the bulk carbonate U isotope signatures. It is noteworthy that the minor U/Ca increase at Site 865 corresponds consistently with benthic foraminifera turnover patterns that are suggestive of de-oxygenation\textsuperscript{51} (see below; Supplementary Fig. 1), implying that early authigenic carbonate signatures are preserved during burial. Moreover, the replication of pre-PETM $\delta^{238}U_{\text{CAU}}$ at Site 401 strongly suggests that the Site 865 record is not compromised by recrystallization.

**Influence of local de-oxygenation.** Close to the onset of the PETM, the three sites show very different responses in U systematics. Both Sites 690 and 401 record higher Ca/U correlated with higher $\delta^{238}U_{\text{CAU}}$ (Fig. 1). In pelagic sediments, elevated U/Ca in foraminifera is thought to semi-quantitatively track overall authigenic U enrichment\textsuperscript{53} which is also likely to be reflected in the bulk carbonate data presented here. Together with the higher $\delta^{238}U_{\text{CAU}}$ values, these trends indicate more reducing conditions and are consistent with other proxy evidence in these sediments, including benthic species turnover patterns\textsuperscript{21,56}, barium accumulation rates (BARs)\textsuperscript{10}, cerium anomaly data, and chromium isoforms\textsuperscript{20} (Supplementary Figs. 2 and 3; Supplementary Note 1). By contrast, Site 865 records only a minor increase in U/Ca, that is also coincident with local benthic faunal turnover evidence for increased food supply or lower O$_2$ conditions\textsuperscript{51} (Supplementary Fig. 1; Supplementary Note 1), but with no resolvable change in $\delta^{238}U_{\text{CAU}}$ (Fig. 1). These two behaviors can be clearly separated by examining the relationship between U concentrations and $\delta^{238}U_{\text{CAU}}$ for the three localities, a relationship that is likely a function of localized redox conditions (Fig. 2). Note that in Fig. 2, U concentrations are expressed as Ca/U, as opposed to U/Ca in Fig. 1, to better identify the relationship of U enrichments coupled to high $\delta^{238}U_{\text{CAU}}$. All the data for Site 865 and the pre-PETM data for Site 401 are expressed as $\delta^{238}U_{\text{CAU}}$ by recrystallization. The data from Site 865 (black circles) and pre-PETM data from Site 401 that plot along the flat array of Ca/U likely record $\delta^{238}U_{\text{sw}}$, $\delta^{238}U_{\text{CAU}}$ values below inferred $\delta^{238}U_{\text{sw}}$ at Site 690 could be caused by the adsorption of isotopically light U directly to calcite or repartitioned during organic matter re-mineralization. Error bars for $\delta^{238}U_{\text{CAU}}$ represent the external reproducibility of the NIST-SRM 1d limestone standard ($\pm 0.06\%$, 2 SD).

**Fig. 2 Relationship of $\delta^{238}U_{\text{CAU}}$ and U concentrations.** Note that U concentrations are expressed as Ca/U (inverted relative to Fig. 1) in order to emphasize the samples showing U enrichment (lower Ca/U) coupled to high $\delta^{238}U_{\text{CAU}}$ at Sites 690 (red circles) and 401 (blue circles). These samples demonstrate a near-linear relationship when Ca/U < 8 mol/μmol, which are consistent with a $\Delta^{238}U_{\text{sw}}$ of up to $\pm 0.6\%$ relative to inferred $\delta^{238}U_{\text{sw}}$. Data from Site 690 (black circles) and pre-PETM data from Site 401 that plot along the flat array of Ca/U likely record $\delta^{238}U_{\text{sw}}$. $\delta^{238}U_{\text{CAU}}$ values below inferred $\delta^{238}U_{\text{sw}}$ at Site 690 could be caused by the adsorption of isotopically light U directly to calcite or repartitioned during organic matter re-mineralization. Error bars for $\delta^{238}U_{\text{CAU}}$ represent the external reproducibility of the NIST-SRM 1d limestone standard ($\pm 0.06\%$, 2 SD).

**Quantifying upper limits on the global extent of seafloor anoxia.** The above discussion indicates that the Site 865 $\delta^{238}U_{\text{CAU}}$ record has not been affected by local de-oxygenation or diageneric, and likely records seawater compositions throughout the PETM. The invariance of the Site 865 record is taken to indicate the lack of a resolvable perturbation to the global U isotope mass balance during the PETM. Here we use an established biogeochemical box model\textsuperscript{32,47} to assess the quantitative samples\textsuperscript{41,42,48,58}, so it may simply be a consequence of variable mixtures of carbonate allochems and cements, despite the exclusive presence of low-Mg calcite in these samples\textsuperscript{35-37}, such that cement bound authigenic U(VI) is diluted by low-U biogenic carbonate. Variations in grain size and surface area, and therefore faunal morphologies, are also known to affect the degree of U uptake in carbonate cements\textsuperscript{55}. But, given other proxy evidence outlined above, one plausible explanation could be through enhanced authigenic U(IV) uptake with a $\delta^{238}U$ close to seawater, as seen in some modern hypoxic to suboxic continental slope localities, where minor U enrichments are thought to occur within an effectively closed system due to extreme diffusive supply limitation\textsuperscript{39,60}. Alternatively, minor variations in U/Ca (on the order of 0.01 μmol/mol) could reflect changes in the distribution coefficient for U into cements driven by local aqueous speciation or precipitation rate, potentially linked to ambient carbonate saturation state or pH\textsuperscript{38}. In particular, lower pH conditions at the PETM\textsuperscript{4,54} could be another explanation for the minor U/Ca increase at Site 865 (Fig. 1). A final possibility is that these variations reflect the partial leaching of detrital material (Supplementary Note 2) or phosphate-associated U (which might also increase during higher productivity periods), although these should largely be avoided with the methods used here\textsuperscript{40,41,61}. Irrespective of the exact reason for the variation in U concentrations at Site 865, Fig. 2 demonstrates that this had no resolvable influence on the $\delta^{238}U_{\text{CAU}}$. 
implications of this result. We use the Site 865 $\delta^{238}$U_CAU dataset and higher resolution $\delta^{13}$C data from Site 401 (ref. 1) as targets for the biogeochemical model, which calculates the coupled dynamics of C, P, and U cycling in response to hypothesized carbon emissions. In the C–P–U model, the resultant increase in atmospheric CO$_2$ and temperature drives a vegetation-mediated increase in silicate weathering which is a sink for carbon (after carbonate burial), but a source of U and P. Resultant P inputs then drive changes in global marine primary productivity and organic carbon burial, which is a sink for carbon. Burial of terrestrial organic carbon also responds directly to CO$_2$ concentrations. Oxygen demand in the ocean is controlled by primary productivity and organic matter remineralization rates, and determines the extent of seafloor anoxia (where 0.25% of the modern seafloor area has <0.5 mL/L dissolved O$_2$; ref. 62). The P cycle is also redox sensitive, accounting for the positive feedback mechanism of P recycling under anoxic conditions, which acts to further stimulate marine primary productivity. Carbon isotope mass balance tracks relative changes in sources and sinks of inorganic and organic carbon. Model U cycling and $\delta^{238}$U respond to changes in the anoxic U burial flux (scaled to benthic anoxic area) as well as weathering inputs of U, allowing us to simulate the dynamical relationship between predicted seafloor anoxia, ocean U content, and $\delta^{238}$U$_{sw}$. Solutions are sensitive to the assumed global average $\Delta^{238}$U$_{inox}$, where we use +0.4‰ and +0.6‰ based on modern isotopic mass balance constraints and observations of modern anoxic sediments$^{28,57}$ (Supplementary Note 3).

The model is set up for Late Paleocene background conditions (see “Methods”) and is perturbed by a plausible PETM carbon injection with an emission rate, magnitude and isotope composition ($\delta^{13}$C$_{inapp}$) informed by previous studies$^{1-3}$. In the C–P–U model, oceanic anoxia is determined by longer-term feedback processes (silicate weathering and seawater P concentrations) with a response time (>10 kyrs) on the order of the P residence time in the global ocean. The model does not account for temperature-related O$_2$ solubility effects, which are deemed less important for driving OAEs$^{63}$, or spatially controlled factors that could drive local anoxia on shorter timeframes. As such, the model estimates of global seafloor anoxia are insensitive to the precise rate and structure of carbon emission for a given total magnitude (Supplementary Note 4). But, oceanic anoxia is sensitive to background atmospheric $pO_2$, and previous studies suggest higher than modern levels during the Paleocene$^{38}$. Here we use $pO_2 = 1.1–1.3$ present atmospheric levels (PAL) as a starting point. The total emission magnitude was tuned (to 9600 PgC) such that the calculated $\Delta^{238}$U$_{sw}$ response for the lower $\Delta^{238}$U$_{inox}$ set up (+0.4‰) is within the uncertainty of the Site 865 $\delta^{238}$U_CAU data in the most sensitive model version ($pO_2 = 1.1$ PAL). But, for the same perturbation, the higher $\Delta^{238}$U$_{inox}$ set up (+0.6‰) results in a larger negative $\Delta^{238}$U excursion that exceeds the limits of the observations (Fig. 3). Smaller emission magnitudes (with correspondingly smaller anoxia responses) would also be compatible with the $\Delta^{238}$U$_{CAU}$ data in all $pO_2$ setups. Larger emissions would also be consistent but only if we assume higher $pO_2$ (<20,400 PgC when $pO_2 = 1.3$ PAL; <15,000 PgC when $pO_2 = 1.2$ PAL) in order to limit the absolute extent of anoxia, or use an unrealistically low $\Delta^{238}$U$_{inox}$ to limit $\Delta^{238}$U$_{sw}$ response. But emissions above ~15,000 PgC are not supported by other proxy datasets$^{1-3}$. With this approach, we place a maximum limit on the extent of seafloor anoxia permitted by the data, but do not attempt to refine estimates of the total emission magnitude$^{1-3}$.

In response to the 9600 PgC emission, enhanced silicate weathering results in a ~30% increase in oceanic P inputs, and an ~10-fold relative increase in anoxic seafloor area, regardless of $pO_2$ (Fig. 3, Table 1). In the most sensitive model set up ($pO_2 = 1.1$ PAL), benthic anoxia could have transiently occupied ~2% of the total seafloor area while remaining within the uncertainty range of the $\delta^{238}$U_CAU data. Importantly, the anoxia estimate from the dynamical model allows for a transient expansion of anoxia which is not captured by steady-state isotope mass balance solutions (Table 1), highlighting the strength of coupled biogeochemical–isotope models for investigating perturbation events. This is because anoxia is coupled to the oceanic P reservoir, which has a faster response time than the U reservoir. Furthermore, the decline in the U reservoir due to anoxic burial is muted by increased U inputs from weathering (Fig. 3) that could be important for events on the scale of the PETM.

In previous model studies examining PETM emission and recovery scenarios, marine organic carbon burial fluxes were specifically imposed, or inferred using an inverse model approach, ~30 kyrs after the PETM onset in order to help explain the “rapid” termination of the PETM CIE, which cannot be fully
Table 1 Model results comparing steady-state and dynamic model solutions for anoxia and U cycle responses.

|                        | Steady-state model<sup>a</sup> | Dynamic biogeochemical model<sup>b</sup> |
|------------------------|--------------------------------|---------------------------------|
|                        | $pO_2 = 1.1$                  | $pO_2 = 1.2$                  | $pO_2 = 1.3$                  |
| Seafloor anoxia        |                                |                                |                                |
| Modern                 | 0.25%<sup>c</sup>            | 0.25% for modern setup, $pO_2 = 1$ | 0.07%                        |
| pre-PETM               | 0.09 – 0.16<sup>d</sup>      | 0.25%                          | 0.02%                        |
| PETM                   | 0.39 – 0.71<sup>d</sup>      | 2.03%                          | 0.60%                        |
| Negative $\delta^{238}$U excursion | 0.08%<sup>e</sup> | 0.08 – 0.12<sup>d</sup>      | 0.02 – 0.03<sup>d</sup>    | 0.005 – 0.008<sup>d</sup> |

<sup>a</sup>The steady-state model implicitly accounts for changes in $pO_2$ by deriving an extent of anoxia consistent with pre-PETM seawater of $\delta^{13}$C = −0.36‰.

<sup>b</sup>Dynamic solutions are given for a fixed carbon emission of 9600 PgC.

<sup>c</sup>Modern seafloor anoxia constraints from ref. 52.

<sup>d</sup>Min and max estimates are for $\Delta^{238}$U<sub>anox</sub> = −0.4‰ and −0.6‰, model runs.

<sup>e</sup>$\Delta^{238}$U excursion of −0.08‰, matches the dynamic model $pO_2 = 1.1$, $\Delta^{238}$U<sub>anox</sub> = −0.4‰, scenario.

explained by silicate weathering alone<sup>1,3,64</sup>. By contrast, a key feature of the C–P–U model is the mechanistic response of the P cycle (and hence C and U cycles) to global temperature changes and weathering activity (Supplementary Note 5). Thus, the model calculates changes in marine organic carbon burial fluxes (and terrestrially derived organic carbon burial), in addition to silicate weathering carbon fluxes, in an internally consistent manner. With this forward modeling approach, the ~9600 PgC input flux results in a good fit to the PETM CIE, and, more importantly, the recovery of $\delta^{13}$C is well captured (Fig. 3). The cumulative burial of marine organic carbon is ~3800 PgC under this scenario, which is within the range of published estimates<sup>1,3,64</sup>, but with lower burial rates (~0.02 PgC/yr) spread over the extended PETM and recovery interval (Fig. 3). In comparison, silicate weathering accounts for a larger proportion of carbon burial, at ~4900 PgC. Together, these diverse model approaches highlight the importance of both silicate weathering and marine organic carbon burial in driving the recovery from the PETM.

Discussion

Uranium isotope records in carbonate sediments can provide unique insights into the timing and severity of both local and global-scale changes in ocean de-oxygenation, depending on the oceanographic setting of the sample locality. Site 690 and 401 demonstrate the utility of $\delta^{238}$U<sub>CAU</sub> as a local redox indicator in carbonate sediments, one that likely integrates bottom-water and pore-water conditions. When interpreted within a wider geochemical and paleontological framework, these data can provide information on subtle changes at and below the water sediment interface. For Sites 401 and 690, $\delta^{238}$U<sub>CAU</sub> indicates the presence of anoxia, at least in pore-waters, and benthic faunal patterns suggest low-O2 bottom waters<sup>21</sup>, (e.g. buliminids and bolivinids survive with <0.1 to 3 mL/L dissolved O2 in modern settings<sup>65</sup>). Conditions appear to be more severely, or more permanently, reducing at Site 690 during the PETM recovery interval, where $\delta^{238}$U<sub>CAU</sub> reach values similar to modern anoxic or euxinic environments<sup>28,37</sup>. But the continued presence of low-O2 tolerant foraminifers<sup>21</sup> suggests either temporally fluctuating conditions, or extremely low dissolved oxygen, and the preservation of marine barite implies a lack of euxinic pore-waters<sup>8</sup>. The initial rise in U/Ca at Site 690 suggests that de-oxygenation may have begun ~100 kys before the PETM CIE, which could be an expression of background variability at the locality given that the abundance of buliminids also varies deeper in the core<sup>21</sup> (Supplementary Fig. 2), or an indicator of early environmental change related to the PETM hyperthermal. Unfortunately, for Site 401, we cannot precisely constrain the timing or severity of local redox changes during the initial phase of the CIE (including the acidification interval) using U/Ca and $\delta^{238}$U<sub>CAU</sub>. But, U enrichment factors, measured on bulk sediment digests that were sampled at higher resolution<sup>13</sup>, and foraminiferal chromium isotope data<sup>20</sup> suggest a brief interval of relatively more reducing conditions at the PETM onset compared to the rest of the measured interval (Supplementary Fig. 3).

In contrast to Sites 690 and 401, Site 865 remained predominantly oxygenated throughout the PETM with no evidence for the modification of $\delta^{238}$U<sub>CAU</sub> by authigenic U(IV) uptake, suggesting that this locality records changes in open ocean $\delta^{238}$U<sub>anox</sub>, and hence reflects changes in the global extent of seafloor anoxia. The invariance in this new $\delta^{238}$U<sub>CAU</sub> dataset therefore provides robust upper limits on the degree of environmental anoxia during the PETM, indicating that any oceanic de-oxygenation must have been less than that required to perturb the global U isotopic mass balance. This result is in contrast to interpretations of published barite S isotope data in terms of a perturbation of the global S-cycle related to the expansion of sulfidic waters during the PETM<sup>27</sup>. The unexpected decoupling of these proxy records requires further investigation. The barite S isotope record is hypothesized to reflect an increase in water column microbial sulfate reduction that generated a transient sea-water sulfide reservoir within an oxygen minimum zone (OMZ), rather than sub-sediment processes that lead to pyrite burial and S removal<sup>26,66</sup>. Under this scenario, U reduction would occur in a much smaller area where the OMZ impinged on the seafloor, whilst more extensive sulfate reduction could occur in open waters. The different responses might also imply an even lower sulfate reservoir than previously assumed (5 mM<sup>27</sup>), one that would more easily see a perturbation of S isotopic mass balance.

Based on the invariance of the Site 865 record, we estimate a maximum ~10-fold relative increase in anoxic U sinks across the PETM, with anoxia likely restricted to <2% of seafloor area. Anoxia expansion above this estimate would require the effective isotope enrichment factor to be less than +0.4‰, which is inconsistent with observations from modern anoxic settings<sup>28</sup>. Any expansion of anoxic conditions would be accompanied by an expansion of low-O2 (i.e. suboxic) conditions (e.g. ref. 20), which would be associated with only minor U isotope fractionation and a lesser degree of authigenic U enrichment<sup>28,59,60</sup>, and therefore still compatible with the $\delta^{238}$U<sub>CAU</sub> invariance observed at Site 865.

The model approach implicitly assumes that enhanced nutrient inputs were the primary driver of oceanic anoxia on a global scale, in a manner that is consistent with previous model results for Mesozoic OAEs<sup>32,63</sup>, model-data misfits for the PETM<sup>24</sup>, and proxy evidence in some regions of the ocean<sup>2</sup>. But the limits provided by U isotopes stand independent of this assumption. The extent of seafloor anoxia during the PETM may have been an order of magnitude lower than the most severe of the Mesozoic
Methods
Chemical preparation. Samples were processed and measured at ETH Zurich (ETHZ; Sites 865, 401, and 690) and the Open University (OU; Site 865). At ETHZ, ~600 mg of dry sample powder was leached using 2 × 40 mL of 1 M ammonium acetate (pH 5) for 24 h at room temperature in order to selectively dissolve the bulk carbonate fraction46. The excess volume of acid ensures near complete carbonate dissolution with this method. At the OU, selective carbonate leaching was performed using 0.5 M acetic acid and dissolved ~50% of the carbonate. The IRMM-3636 U double spike was added before column chemistry for all samples.

At ETHZ, leachates were oxidized and dried down with excess concentrated HNO3. The salt precipitates were then converted to chloride form with 7 M HCl, dried and re-dissolved in 10 mL of 1 M HCl. The solution was loaded onto ~0.2 mL of 50–100 µm pre-cleaned RE-Resin (Triskem Technologies). Columns and resin were pre-cleaned with 2 mL of 0.2 M HCl + 0.3 M H2O2, rinsed with MilliQ water, and pre-conditioned with 2 × 2 mL of 1 M HCl. Matrix elements were eluted during the first column pass using 2 mL of 1 M HCl and U collected using 2 mL of 0.2 M HCl + 0.3 M H2O2. The samples were dried down and re-dissolved in 1 mL of 1 M HCl. Column washes pass on a fresh column passed a fresh resin bed. U was eluted on the second pass using 4 mL of 1 M HCl followed by 2 mL of 0.2 M HCl and U was collected using 2 mL of 0.2 M HCl + 0.3 M H2O2. Before analysis, samples were vigorously oxidized overnight with concentrated HClO4 and HNO3, in order to break down minor resin contribution, and re-dissolved in 0.2 M HCl. Total procedural blanks are estimated at ~30 pg for U.

At the OU, leachates were similarly oxidized and dried down with excess concentrated HNO3. The samples were then taken up in 7.5 M HNO3 and loaded on a 2-mL pre-column filled with Bio-rad AG1X8 resin. The U fraction was eluted with 0.5 M HCl, before being dried down and taken up in 3 M HNO3 for loading onto a 0.2-mL column packed with UTEVA resin (Triskem Technologies). The remaining matrix was eluted with 5 M HCl before the U fraction was collected using 0.5 M HCl. Similarly to the ETHZ protocol, samples were oxidized overnight with concentrated HClO4 and HNO3, before being taken up in 2% HNO3 for MC-ICPMS measurements. Total procedural blanks measured by isotope dilution were between 6 and 15 pg of U.

Trace and major element concentrations. Leachates were directly aliquoted and diluted 200–400 times in 2% HNO3. Major and trace elements were measured using a Thermo–Finnigan Element XR ICP–MS with an internal indium (In) standard and a blank correction applied. Concentrations were calculated relative to an in-house, well characterized and gravimetrically produced artificial standard with major, trace, and isotopic reference materials, and reported normalized to Ca (µmol/mol). U/Ca are not reported for OU samples as they represent incomplete carbonate dissolution. Uncertainties on metal concentrations and ratios are twice the relative standard deviation (RSD) of a carbonate standard which are typically between 10 and 15%.

Isotope measurements. Isotope ratios were measured on a Neptune Plus (Thermo–Finnigan) MC–ICPMS equipped with an Aridus II DSN (CETAC) and using a一样化器 and spray chamber (CPI) sample introduction system at ETHZ or OU. Given the low concentrations in the leachates “jet + X-cones” were used. Uranium isotope ratios are reported relative to the standard CRMY-145 = 0‰ for 238U/235U and secular equilibrium for 234U/238U, and presented as:

\[
\delta^{238}U = \left( \frac{238U}{235U_{\text{sample}}} / 238U_{\text{CRMY-145}} \right) - 1 \times 1000.
\]

\[
\delta^{234}U = \left( \frac{234U}{238U_{\text{sample}}} / 234U_{\text{sec}} \right) - 1 \times 1000.
\] (1)

Internal errors (2 SE) for δ238U measurements were typically 0.02–0.04‰ at ion beam intensities of ~35–40 V (using 1011 ohm resistors) for ~40 p.p.m. solutions. At ETH, we used two secondary standards to assess external reproducibility and accuracy. First, a uraninite standard, CZ-1, used previously at ETH Zurich, was run between five unknown samples and gives a δ235U of −0.04 ± 0.05‰ (n = 68, 2 SD) during the measurement period, which is identical to values reported elsewhere29,40. To determine the external reproducibility of a carbonate sample, we used the NIST SRM-1d argillaceous limestone that was processed in two sample batches in the same manner as samples, but dissolved using 1 M HCl. NIST SRM-1d gave a δ235U of −0.12 ± 0.06‰ (2 SD, n = 14) which is identical to values reported elsewhere42. At the OU, SRM 950A was run between every three unknown samples and gives, over the analysis period, a δ235U of −0.30 ± 0.03‰ (n = 7, 2 SD), within error of published values28. Additionally, two natural standards, processed through column chemistry, were run for assessing accuracy and long-term reproducibility: Seawater, yielding a δ235U of −0.39 ± 0.05‰ (n = 8, 2 SD), similar to published values (−0.39 ± 0.01‰)26; BE-N basal powder yielding a δ235U of −0.30 ± 0.03‰ (n = 7, 2 SD), within error of published values (−0.33 ± 0.03‰)26.

Age model. Here we use the He age model48 rather than the cyclostrophic ages49 for all sites. The He model is favored because of the consistency of de-oxygenization in δ234Ucalculated with barium isotope and BAR records at Site 690 when using this age model48 (Supplementary Fig. 2; Supplementary Note 2). We use age tie points for the CIE onset from Röhl et al.69 as used for Sites 401 and 690 in previous studies1. The age model for Site 865 is less well constrained compared to the Site 690 and 401 records. Here we match the foraminiferal δ18O data from Site 865 to the bulk carbonate record for Site 690, and check this against planktic records from Site 401, as shown in Supplementary Fig. 4. There is a small gap in the Site 865 record that is thought to be caused by winnowing. This has little bearing on the completeness of the δ238Ucalculated record and subsequent interpretations as the U isotope response to the PETM perturbation is expected to be preserved later due to the non-linearities of anoxia and the large seawater U reservoir (Fig. 3). The uncertainties in the age model have no significant bearing on the findings of anoxia (Supplementary Note 4).

Dynamic biogeochemical model. The C–P–U model is informed by previously established biogeochemical models (COPE65) and calculates the coupled dynamics of C, P, and U cycling associated with changes in temperature, weathering, and oceanic anoxia, in response to hypothesized CO2 perturbations. Full model details are given in ref.32. We make two changes to the model described there: (1) We do not account for changes in O2 due to changes in the burial of organic carbon (because the timescale is too short for significant changes in O2 or we want to treat O2 as an uncertain boundary condition); (2) We increase the climate sensitivity to doubling atmospheric pCO2 to a value of ~5 °C, appropriate for the PETM29,60 in order to better capture the weathering response to CO2 perturbation at this time. The model is set up with appropriate boundary conditions for background degassing (D) uplift and erosion (E) and weatherability (W) based on COPE estimates65. With the revised climate sensitivity, these baseline settings result in atmospheric pCO2 of 1.65 PAL (PAL = −300 p.p.m.), in reasonable agreement with constraints from other PETM studies3. We use a total C emission scenario informed by previous estimates3–5 with a total duration of 75 kyrs. Peak carbon inputs reach 0.54 PgC/yr and are followed by a prolonged C leak at a rate of 0.07 PgC/yr. The isotopic composition of the C source is varied between −3 and −25‰, representing a mix of C from mantle sources and organic C reservoirs (e.g. methane or oxidized organic C), in order to fit as well as possible the planktic δ13C record from Site 401. The δ13C of organic carbon (δ13Corg) is assumed to be −25‰ throughout for simplicity.

U isotopic mass balance. The U-cycle model is as presented in ref.32. It assumes that the riverine input of U is driven by silicate weathering (Fsil), all U sinks scale linearly with U concentration, the hydrothermal sink of U additionally scales with degassing from seafloor spreading (D), and the anoxic sink of U also scales with fmonoxic, the fraction of anoxic seafloor:

\[
dU/\, dt = F_{\text{in}} - F_{\text{surf}} - F_{\text{monoxic}} - F_{\text{other}}.
\]

(3)

\[F_{\text{in}} = k_{\text{in}} \times F_{\text{sil}} / F_{\text{vol}}.
\]

(4)

\[F_{\text{surf}} = k_{\text{surf}} \times D \times (U/\, U_{\text{in}}).
\]

(5)

\[F_{\text{monoxic}} = k_{\text{monoxic}} \times (U/\, U_{\text{in}}) \times f_{\text{monoxic}} / f_{\text{monoxic}}^{\text{ref}}.
\]

(6)

\[F_{\text{other}} = k_{\text{other}} \times (U/\, U_{\text{in}}),
\]

(7)

where the subscript "0" denotes present day values. For a modern U, 1.85 × 1013 mol, we use estimates of kmonoxic = 6.2 × 106 mol/yr, Fsurf = 5.7 × 106 mol/yr, kother =...
\[ \frac{d(U \times \Delta V)}{dt} = F_{\text{flx}} \times \Delta V_hyd - F_{\text{atm}} \times (\Delta V_{\text{flx}} + \Delta V_{\text{atm}}) - F_{\text{sea}} \times (\Delta V_{\text{flx}} + \Delta V_{\text{sea}}). \]  

From the chain rule:

\[ \frac{d(V_{\text{atm}})}{dt} = \left( F_{\text{flx}} \times (\Delta V_{\text{flx}} - \Delta V_hyd) - F_{\text{atm}} \times (\Delta V_{\text{flx}} - \Delta V_{\text{atm}}) - F_{\text{sea}} \times (\Delta V_{\text{flx}} - \Delta V_{\text{sea}}) \right) / U. \]

For the modern, we assume \( \Delta V_{\text{flx}} = 0.29\% \), \( \Delta V_hyd = 0.2\% \), \( \Delta V_{\text{atm}} = 0.5\% \), and \( \Delta V_{\text{sea}} = 0.39\% \), requiring \( \Delta V_{\text{atm}} = 0.0153\% \) for steady state. Solving for steady state and rearranging gives:

\[ \Delta V_{\text{atm}} = \frac{\delta f_{\text{atm}}}{\left( \frac{\Delta V_{\text{flx}}}{\Delta V_{\text{atm}}} + \frac{\delta f_{\text{atm}}}{\Delta V_{\text{atm}}} \right)}. \]

The present-day anoxic fraction \( \left( \delta f_{\text{atm}} \right) \) is estimated to be 0.2–0.3% of seafloor area. In the full model, \( \delta f_{\text{atm}} = 0.002473 \) in the middle of this range. Equation (9) allows us to derive steady-state anoxic fraction for a given set of parameters and seawater \( \Delta V_hyd \). At 55 Ma, we assume no change in \( \Delta V_{\text{flx}} = 0.29\% \), \( \Delta V_hyd = 0.2\% \), \( \Delta V_{\text{atm}} = 0.5\% \), and \( \Delta V_{\text{sea}} = 0.39\% \). We assume elevated degassing \( D = 1.4 \) (ref. 39) and explore \( \Delta V_{\text{atm}} = 0.4\% \) as well as 0.6%, for a pre-PETM seawater of \( \Delta V_{\text{atm}} = 0.36\% \). In the steady-state solution, an excursion magnitude of \( \Delta V_{\text{flx}} \) is defined by the dynamic model for \( \delta V_{\text{flx}} = 1.1 \). Anoxic fraction was 0.04% in scenario, in order to compare the two approaches.

**Data availability**

All data generated or analyzed during this study are included in this published article (and its Supplementary information files). Source data are provided with this paper.

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**References**

1. Gutjahr, M. E. et al. Very large release of mostly volcanic carbon during the Paleocene-Eocene Thermal Maximum. *Nature* **548**, 573 (2017).

2. Haynes, L. L. & Hönisch, B. The seawater carbon inventory at the Eocene thermal maximum. *Nature* **470**, 202–212 (2011).

3. Dunkley Jones, T. et al. Climate model and proxy data constraints on ocean warming across the Paleocene-Eocene Thermal Maximum. *Earth-Sci. Rev.* **125**, 123–145 (2013).

4. Zachos, J. C. et al. Rapid acidification of the ocean during the Paleocene-Eocene thermal maximum. *Science* **313**, 1252–1255 (2006).

5. Carignan, M. J. et al. Hydrological and associated biogeochemical consequences of rapid global warming during the Paleocene-Eocene Thermal Maximum. *Global Planet. Change* **157**, 114–138 (2017).

6. Bains, S., Corfield, R. M. & Norris, R. D. Mechanisms of climate warming at the end of the Paleocene. *Science* **285**, 724–727 (1999).

7. Dunkley Jones, T. et al. Climate model and proxy data constraints on ocean warming across the Paleocene-Eocene Thermal Maximum. *Earth-Sci. Rev.* **125**, 123–145 (2013).

8. Bains, S. & Norris, R. D. Abrupt reversal in ocean overturning during the Paleocene-Eocene warm period. *Nature* **439**, 60–63 (2006).

9. Bains, S., Norris, R. D., Corfield, R. M. & Faul, K. L. Termination of global warmth at the Paleocene/Eocene boundary through productivity feedback. *Nature* **407**, 171–174 (2000).

10. Luyendyk, B. L., Haq, B. U., Dickens, G. R., Eicher, J. L. & Mutter, C. J. Oxygenation of the Paleocene ocean. *Geology* **30**, 330–333 (2002).

11. Dickson, A. J., Cohen, A. S. & Coe, A. L. Seawater oxygenation during the Paleocene-Eocene thermal maximum. *Geology* **30**, 439–442 (2002).

12. Dickson, A. J. et al. The spread of marine anoxia on the northern Tethys margin during the Paleocene-Eocene Thermal Maximum. *Paleoceanogr.* **29**, 964–975 (2014).

13. Zhou, X., Thomas, E., Rickaby, R. E., Winguth, A. M. & Lu, Z. I/Ca evidence for upper ocean deoxygenation during the PETM. *Paleoceanogr.* **29**, 964–975 (2014).

14. Zhou, X., Thomas, E., Rickaby, R. E., Winguth, A. M. & Lu, Z. I/Ca evidence for upper ocean deoxygenation during the PETM. *Paleoceanogr.* **29**, 964–975 (2014).

15. Zhou, X. et al. Expanded oxygen minimum zones during the late Paleocene-Eocene: Hints from multiproxy comparison and ocean modeling. *Paleoceanogr.* **31**, 1532–1546 (2016).
shells. *Geochim. Cosmochim. Acta* [https://doi.org/10.1016/j.gca.2020.06.002] (2020).

44. Holmden, C., Amini, M. & Francois, R. Uranium isotope fractionation in Saanich Inlet: a modern analog study of a paleoredox tracer. *Geochim. Cosmochim. Acta* 153, 202–215 (2015).

45. Zhang, F. et al. Global-ocean redox variations across the Smithian-Spathian boundary linked to concurrent climatic and biotic changes. *Earth-Sci. Rev.* 195, 147–168 (2019).

46. Zhang, F. et al. Multiple episodes of extensive marine anoxia linked to global warming and continental weathering following the latest Permian mass extinction. *Sci. Adv.* e1602921 (2018).

47. Zhang, F. et al. Uranium isotopes in marine carbonates as a global ocean paleoredox proxy: a critical review. *Geochim. Cosmochim. Acta* [https://doi.org/10.1016/j.gca.2020.06.002] (2020).

48. Chen, X. et al. Diagenetic effects on uranium isotope fractionation in carbonate sediments from the Bahamas. *Geochim. Cosmochim. Acta* 237, 294–311 (2018).

49. Romaniello, S. J., Herrmann, A. D. & Anbar, A. D. Uranium concentrations and $^{387}\text{U}/^{235}\text{U}$ isotope ratios in modern carbonates from the Bahamas: assessing a novel paleoredox proxy. *Chem. Geol.* 362, 305–316 (2013).

50. Pearson, P. N. et al. Warm tropical sea surface temperatures in the Late Cretaceous and Eocene epochs. *Nature* 413, 481–487 (2001).

51. Arreguin-Rodriguez, G. J., Aleger, L. T. & Thomas, E. Late Eocene-middle Eocene benthic foraminifera on a Pacific seamount (Allison Guyot, ODP Site 865): Greenhouse climate and superimposed hyperthermal events. *Paleoceanography*. *Paleoceanography*. 31, 346–364 (2016).

52. Edgar, K. M., Anagnostou, E., Pearson, P. N. & Foster, G. L. Assessing the impact of diagenesis on $\delta^{27}\text{B}$, $\delta^{13}\text{C}$, $\delta^{18}\text{O}$, Sr/Ca and B/Ca values in fossil planktonic foraminiferal calcite. *Geochim. Cosmochim. Acta* 166, 189–209 (2015).

53. Borel, T. et al. Late Paleocene to Eocene paleoceanography of the equatorial Pacific Ocean: stable isotopes recorded at Ocean Drilling Program Site 865. *Allison Guyot. Paleoceanog. Paleoceanol. Acta*. 10, 841–865 (1995).

54. Pennan, D. E., Hönisch, B., Zeebe, R. E., Thomas, E. & Zachos, J. C. Rapid and sustained surface ocean acidification during the Paleocene-Eocene Thermal Maximum. *Paleoceanog. Paleoceanol. Acta*. 29, 357–369 (2014).

55. Boitou, R., Greaves, M. & Elderfield, H. Authigenic uranium in foraminiferal coatings: a proxy for ocean redox chemistry. *Paleoceanog. Paleoceanol. Acta*. 27 (2012).

56. Yamaguchi, T. & Norris, R. D. Deep-sea ostracode turnovers through the Paleocene–Eocene thermal maximum in DSDP Site 401, Bay of Biscay, North Atlantic. *Mar. Micropaleontol*. 86, 32–44 (2012).

57. Andersen, M. et al. A modern framework for the interpretation of $^{238}\text{U}/^{235}\text{U}$ in studies of ancient ocean redox. *Earth Planet. Sci. Lett.* 400, 184–194 (2014).

58. Russell, A. D., Hönisch, B., Spero, H. J. & Lea, D. W. Effects of seawater carbonate ion concentration and temperature on shell U, Mg, and Sr in cultured planktonic foraminifera. *Geochim. Cosmochim. Acta* 68, 4347–4361 (2004).

59. Weyer, S. et al. Natural fractionation of $^{238}\text{U}/^{235}\text{U}$. *Geochim. Cosmochim. Acta* 72, 345–359 (2008).

60. Andersen, M. et al. Closing in on the marine $^{238}\text{U}/^{235}\text{U}$ budget. *Chem. Geol.* 420, 11–22 (2016).

61. Dahl, T. W. et al. Uranium isotopes distinguish two geochemically distinct stages during the later Cambrian SPICE event. *Earth Planet. Sci. Lett.* 401, 313–326 (2014).

62. Helly, J. J. & Levin, L. A. Global distribution of naturally occurring marine hypoxia on continental margins. *Deep Sea Res. I Oceanogr. Res. Pap.* 51, 1159–1168 (2004).

63. Monteiro, F., Pancost, R., Ridgwell, A. & Donnadieu, Y. Nutrients as the dominant control on the spread of anoxia and euxinia across the Cenomanian–Tomian oceanic anoxic event (OAE 2): model-data comparison. *Paleoceanog. Paleoceanol. Acta*. 27 (2012).

64. Bowen, G. J. & Zachos, J. C. Rapid carbon sequestration at the termination of the Paleocene-Eocene Thermal Maximum. *Nat. Geosci.* 3, 866 (2010).

65. Sen Gupta, B. K. & Machain-Castillo, M. L. Benthic foraminifera in oxygen-poor habitats. *Mar. Micropaleontol.* 20, 183–201 (1993).