

1 **Uranium isotope evidence for two episodes of deoxygenation during Oceanic Anoxic Event 2**

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27 **Abstract**

28 Oceanic Anoxic Event 2 (OAE 2), occurring ~94 million years ago, was one of the most extreme carbon
29 cycle and climatic perturbations of the Phanerozoic Eon. It was typified by a rapid rise in atmospheric
30 CO₂, global warming, and marine anoxia, leading to the widespread devastation of marine ecosystems.
31 However, the precise timing and extent to which oceanic anoxic conditions expanded during OAE 2
32 remains unresolved. We present a new record of global ocean redox changes during OAE 2 using a
33 combined geochemical and carbon cycle modelling approach. We utilize a continuous, high-resolution
34 record of uranium isotopes in pelagic and platform-carbonate sediments to quantify the global extent
35 of seafloor anoxia during OAE 2. This new dataset is then compared to a dynamic model of the coupled
36 global carbon, phosphorus and uranium cycles to test hypotheses for OAE 2 initiation. This unique
37 approach highlights an intra-OAE complexity that has previously been under constrained,
38 characterized by two expansions of anoxia separated by an episode of globally significant
39 reoxygenation coincident with the 'Plenus Cold Event'. Each anoxic expansion event was likely driven
40 by rapid atmospheric CO₂ injections from multi-phase Large Igneous Province (LIP) activity.

41

42 **Significance Statement**

43 *Past 'Oceanic Anoxic Events' (OAEs) represent important carbon cycle perturbations that offer the*
44 *opportunity to study the Earth's response to extreme climate warming. A fundamental limitation for*
45 *understanding OAEs is quantifying the timing and total extent of ocean anoxia. We present a*
46 *quantitative account of global redox conditions for OAE 2 (~94 million years ago), using a high-*
47 *resolution record of uranium isotopes combined with a biogeochemical model. We present new*
48 *evidence for two discrete intervals of globally extensive anoxia that were coupled to enhanced*
49 *terrestrial weathering, within the typically defined OAE interval. These anoxic intervals were separated*
50 *by ocean reoxygenation and the temporary recovery of the carbon cycle.*

51

52 **Introduction**

53 Rapid climatic warming events have repeatedly punctuated Earth's history and have often
54 been associated with prolonged episodes of widespread oceanic anoxia. These intervals represent
55 major disturbances to the global carbon cycle, which contributed to marine faunal turnover and mass
56 extinctions (1). In particular, the Mesozoic Era was characterized by numerous prolonged intervals of
57 globally enhanced anoxia, termed 'Oceanic Anoxic Events' (OAEs) (1). The Mesozoic OAEs are
58 considered the model expression of oceanic anoxia in the Phanerozoic geological record, thus
59 informing studies of past hyperthermal events and future environmental change (2, 3). Despite
60 extensive study, the magnitude and timing of oceanic anoxia for OAEs are still poorly constrained,
61 representing a fundamental limitation for defining these events. We address this shortcoming by
62 presenting a new quantitative record of oceanic anoxia for OAE 2.

63 OAE 2 is the most widely recognised and severe OAE, occurring at the Cenomanian–Turonian
64 Boundary (~94 Ma) (1) and lasting for up to ~900 thousand years (kyr) (4). A leading hypothesis for
65 the initiation of OAE 2 is through the emplacement of one or more of the Caribbean, Madagascar and
66 High-Arctic Large Igneous Provinces (LIPs) (1, 5-7) and an associated outgassing of CO₂ that caused
67 climate warming and accelerated weathering (8). Such processes likely drove an increase in the supply
68 of nutrients, especially phosphorus (P), to the ocean, and thereby increased productivity and oxygen
69 demand through the aerobic degradation of organic matter (1, 5, 9). Productivity was potentially
70 further amplified by the effective recycling of P from sediments overlain by anoxic waters (10-12).
71 Ultimately, this chain of events led to anoxic water-column conditions and the enhanced preservation
72 of organic carbon, which formed distinct black shales in marine sedimentary successions in many
73 ocean basins. The preferential burial of isotopically light carbon (C) also contributed to a broad
74 positive carbon-isotope excursion (CIE) across OAE 2 that is utilized as a global chemo-stratigraphic
75 marker of the event (1). Silicate weathering and organic- carbon burial are vital components of the

76 global carbon cycle and represent important negative feedback mechanisms that sequester
77 atmospheric CO₂ and drive climate stabilization during global warming events (8, 12, 13). However,
78 the detailed behaviour of these recovery processes is currently limited for numerous carbon cycle
79 perturbations in Earth history, including OAEs.

80 For OAE 2 there is widespread evidence of regionally extensive anoxia and euxinia (anoxic-
81 sulfidic water-column conditions) in disparate localities, leading to benthic faunal extinctions, globally
82 enhanced organic-carbon burial and the sequestration of redox-sensitive and chalcophilic trace metals
83 (1, 9, 12, 14, 15). Oceanic anoxia is, however, a spatially heterogeneous condition, with a highly
84 variable geochemical and sedimentary expression. A dependence on local redox indicators has
85 significantly limited the integration of this important phenomenon into a global carbon cycle
86 framework. To address this limitation, we present a quantitative record of the global extent of anoxia
87 for OAE 2 using new, high-resolution uranium-isotope (²³⁸U/²³⁵U, reformulated as δ²³⁸U, see SI) data,
88 in combination with a biogeochemical model that calculates changes in the coupled global C, P and U
89 cycles. These new geochemical data come from well-preserved and extensively studied carbonate
90 sediments that were deposited under oxygenated conditions in the European shelf sea and Tethyan
91 continental margin during OAE 2. The studied sections (Fig. 1) include deeper shelf pelagic
92 foraminiferal-nannofossil carbonates (chalks) from Eastbourne and South Ferriby (United Kingdom),
93 and shallow-water platform carbonates from Raia del Pedale (Italy). We also present additional
94 lithium-isotope data (⁷Li/⁶Li, reformulated as δ⁷Li (8)) for Eastbourne which supplement previously
95 published results (8) to give a more highly resolved complementary record of changes in the global
96 weathering regime during OAE 2.

97 Under anoxic conditions, oxidized and soluble U(VI) is drawn down from the water column
98 and deposited in sediments as reduced and immobile U(IV), driving seawater U concentrations ([U])
99 to lower values (16, 17). The primary isotopic fractionation of U is also strongly redox-dependent,
100 whereby ²³⁸U is preferentially sequestered in anoxic sediments, leaving the residual seawater
101 relatively enriched in ²³⁵U (18-25). This geochemical behaviour generates lower δ²³⁸U signatures in
102 seawater that can be recorded in contemporary sediments. Moreover, because of the long residence
103 time of U in the modern ocean (320–560 kyrs (16)) sedimentary δ²³⁸U signatures can resolve global
104 changes in oceanic redox conditions (14, 26-29), which cannot be achieved by more traditional
105 proxies.

106 For OAE 2, application of the U-isotope paleo-redox tracer is limited to a single study of
107 organic-rich black shales (δ²³⁸U_{shale}), wherein the data are suggestive of a three-fold increase in the
108 extent of anoxia across the event (14). These data (Fig. 1) are of insufficient temporal resolution to
109 provide insights into the rate and magnitude of ocean redox changes across the event or elucidate the

110 temporal relationship of ocean anoxia to other aspects of the Earth system response. Furthermore,
111 the fractionation of U into black shales is strongly influenced by local geochemical controls (30) that
112 complicate attempts to infer the 'global' U-isotope signature of oceanic redox changes. By contrast,
113 the U-isotope signature of primary carbonate precipitates ($\delta^{238}\text{U}_{\text{carb}}$) records the seawater isotopic
114 composition ($\delta^{238}\text{U}_{\text{sw}}$), with minimal additional fractionation under most environmental conditions
115 (21, 25, 31, 32).

116 The new datasets presented here are used to test the current hypotheses for OAE 2 initiation,
117 and the resulting biogeochemical changes, by acting as target outputs for the C–P–U model (see
118 Methods and SI). This approach differs from other efforts to model U-isotope data in deep time (14,
119 26-29), in that it calculates dynamic changes to the wider Earth system, including temperature,
120 weathering, nutrient inputs, anoxia and carbon burial, driven by hypothesized perturbations to the
121 carbon cycle. These changes, in turn, affect the oceanic U cycle. The benefit of this approach is that it
122 gives an internally consistent, quantitative assessment of coupled changes in the carbon cycle.

123

124 **Results and Discussion**

125 **Geochemical Reconstruction of Oceanic Anoxia.** The stratigraphically expanded Eastbourne
126 section has been extensively studied for a multitude of paleo-environmental proxies owing to its lack
127 of significant diagenetic alteration (1, 8, 15, 33, 34) (see SI). In particular, the carbonates from
128 Eastbourne have very low values of total organic carbon (TOC) (34) and were not subject to
129 appreciable diagenetic sulfate reduction in pore waters (33), which should have minimized alteration
130 of the primary $\delta^{238}\text{U}_{\text{carb}}$ signatures. As such, the $\delta^{238}\text{U}_{\text{carb}}$ record for Eastbourne should closely reflect
131 variability in $\delta^{238}\text{U}_{\text{sw}}$ throughout the OAE 2 interval and is hence used as the primary reference curve
132 in the present study. The $\delta^{238}\text{U}_{\text{carb}}$ record for Eastbourne has the highest resolution of the three sites
133 and systematic temporal trends are statistically identified using a locally weighted polynomial
134 smoothing function (LOESS) in order to prevent undue emphasis being placed on single data points.
135 The South Ferriby section, although also showing a low degree of diagenetic alteration (see SI), is more
136 lithified than Eastbourne, raising the possibility of greater secondary influences on $\delta^{238}\text{U}_{\text{carb}}$. It also
137 contains a significant unconformity (Fig. 1) which is overlain by an organic-rich black shale.
138 Mobilization of U from this horizon could potentially explain the comparatively high $\delta^{238}\text{U}_{\text{carb}}$ values
139 above the unconformity that are more typical of anoxic sediments (18, 19, 23, 30). Deposits at Raia
140 del Pedale show evidence for partial recrystallization and dolomitization of originally mixed aragonite
141 and high-magnesium-calcite producers (see SI). In particular the lowest $\delta^{238}\text{U}_{\text{carb}}$ of the section are
142 from dolomitized samples with anomalously high U/Ca (Fig. 1 and Fig. S3), consistent with the negative
143 $\delta^{238}\text{U}_{\text{carb}}$ offset and U enrichment observed in modern dolomites (21, 31).

144 Multiple lines of evidence support the fidelity of the Eastbourne $\delta^{238}\text{U}_{\text{carb}}$ record. First, despite
145 potential diagenetic influences in the Raia del Pedale and South Ferriby sections, stability is seen in
146 the $\delta^{238}\text{U}_{\text{carb}}$ records post-dating OAE 2 at all three sites (Fig. 1). It is important to note that, during this
147 stable post-OAE 2 interval, $\delta^{238}\text{U}_{\text{shale}}$ values from the organic-rich black shales of Ocean Drilling
148 Program (ODP) site 1261 (14) are $\sim 0.5\text{‰}$ higher than contemporaneous $\delta^{238}\text{U}_{\text{carb}}$ (Fig. 1). This offset is
149 consistent with the observed magnitude and direction of U-isotope fractionation during U reduction
150 in modern organic-rich sediments under an anoxic water column (18, 19, 23, 30) and therefore
151 suggests that Eastbourne $\delta^{238}\text{U}_{\text{carb}}$ closely reflects absolute values in $\delta^{238}\text{U}_{\text{sw}}$. Furthermore, during this
152 stable period, $\delta^{238}\text{U}_{\text{carb}}$ from Raia del Pedale and South Ferriby are ~ 0.2 to 0.3‰ higher than
153 Eastbourne which is consistent with observations from some modern bulk carbonate sediments that
154 show higher $\delta^{238}\text{U}_{\text{carb}}$ than seawater due to early diagenesis (31). These offsets, compared to
155 Eastbourne, therefore likely reflect the diagenetic gradient between the carbonate sites and support
156 a near-primary $\delta^{238}\text{U}_{\text{sw}}$ signature at Eastbourne. In addition, there are two systematic decreases in
157 $\delta^{238}\text{U}_{\text{carb}}$ at Eastbourne (Anoxia Expansion (AE) 1 and AE2, Fig. 2) which are associated with the
158 progressive lowering of U/Ca and other redox-sensitive or chalcophilic trace metals (TM_{redox}) at
159 multiple localities, consistent with the removal of these elements into anoxic sediments. Finally, there
160 is a broad coupling between two decreases in the $\delta^{238}\text{U}_{\text{carb}}$ and $\delta^7\text{Li}$ records (Fig. 2), which would not
161 be preserved during diagenesis, and substantiates the veracity of both datasets.

162 At Eastbourne, the onset of the CIE, which typically defines the beginning of OAE 2, is closely
163 associated with a systematic decrease in $\delta^{238}\text{U}_{\text{carb}}$ from a maximum of $-0.24 \pm 0.09\text{‰}$ (2SE) to a
164 minimum of $-0.71 \pm 0.06\text{‰}$ (AE1). This drop in values is accompanied by a decrease in TM_{redox} in
165 multiple localities (35-37), including Eastbourne (Fig. 1 & 2) and Raia del Pedale (Fig. 1), suggesting the
166 global expansion of oceanic anoxic conditions. After AE1, $\delta^{238}\text{U}_{\text{carb}}$ rapidly increases to $-0.41 \pm 0.07\text{‰}$
167 suggesting a decrease in the extent of oceanic anoxia. This dramatic change directly corresponds to
168 the onset of the Plenus Cold Event (PCE), an interval characterized by cooler temperatures at a range
169 of latitudes (35, 38-45). Higher $\delta^{238}\text{U}_{\text{carb}}$ values averaging $-0.38 \pm 0.12 \text{‰}$ (1sd) were maintained
170 throughout the PCE and correspond to widely observed TM_{redox} enrichments (35-37) (Fig. 1 & 2),
171 indicating the decreased sequestration and oxidative liberation of these metals under more
172 oxygenated conditions (35). The rise in $\delta^{238}\text{U}_{\text{carb}}$ is also consistent with observations of local benthic
173 faunal re-population and local reoxygenation in some basins, during the PCE (35, 36, 44-47).
174 Importantly, the new $\delta^{238}\text{U}_{\text{carb}}$ dataset indicates that the PCE reoxygenation was globally significant,
175 returning the ocean to a redox state similar to pre-perturbation conditions.

176 The end of the PCE corresponds to a second systematic decrease in $\delta^{238}\text{U}_{\text{carb}}$ (AE2). Again, the
177 concentrations of TM_{redox} decrease at both Eastbourne (Fig. 2) and Raia del Pedale (35) indicating a

178 return to more expanded oceanic anoxia. Interestingly, the recovery of $\delta^{238}\text{U}_{\text{carb}}$ from this decrease is
179 poorly defined and appears more gradual, unlike the rapid changes in $\delta^{238}\text{U}_{\text{carb}}$ observed for the PCE.
180 The lack of a distinct post-OAE 2 recovery may indicate a lingering degree of anoxia after the CIE, or
181 that insufficient time is recorded in these sections to observe the complete re-stabilization of the U
182 and C cycles.

183 Both AE1 and AE2 are coupled to increased global temperatures, as indicated by $\delta^{18}\text{O}$ in
184 Eastbourne (Fig. 2) and organic geochemical proxies in North Atlantic sites (34, 42, 43, 45). Moreover,
185 there is a coupled relationship between decreasing $\delta^{238}\text{U}_{\text{carb}}$ and $\delta^7\text{Li}$ during both AE1 and AE2 (Fig. 2),
186 which is suggestive of a shift to a more congruent weathering regime (greater primary mineral
187 dissolution than secondary mineral formation (8, 48) during these two anoxic intervals. These two
188 intervals are also accompanied by independent evidence for an enhanced hydrological cycle (42),
189 together suggesting two discrete periods of enhanced global weathering flux. In contrast to AE1 and
190 AE2, the increasing $\delta^{238}\text{U}_{\text{carb}}$ and reoxygenation during the PCE corresponds to cooler temperatures
191 and drier conditions (34, 42, 45), as well as the recovery of $\delta^7\text{Li}$ to more positive values. The coupling
192 of datasets therefore empirically supports the hypothesized link between enhanced weathering
193 activity under warmer climates, increased nutrient fluxes, heightened productivity and expanded
194 oceanic anoxia (1, 8, 9, 42). Furthermore, the first interval of decreasing $\delta^7\text{Li}$ and $\delta^{238}\text{U}_{\text{carb}}$ (AE1) also
195 temporally corresponds with a prominent decrease in osmium (Os) isotopes to more unradiogenic,
196 basalt-like signatures that marks the main phase of LIP emplacement associated with the CIE (5-8).
197 The return to more radiogenic Os-isotope signatures is not seen until after the PCE (5-7), reflecting a
198 degree of continued volcanism across the cooling event, although two separate peaks in sedimentary
199 Os concentrations may indicate multiple phases of heightened volcanic activity (5). LIP activity is
200 therefore likely to have driven the two intervals of enhanced weathering activity and resulted in
201 expanded oceanic anoxia.

202 **A biogeochemical model for OAE 2.** We use a C–P–U model (see SI) to examine if plausible
203 fluxes of CO_2 from LIP activity can indeed drive the magnitude of changes observed in $\delta^{238}\text{U}_{\text{carb}}$, U/Ca
204 and $\delta^{13}\text{C}$ at Eastbourne. The model is informed by previously established biogeochemical models (11,
205 13, 49-52) and calculates the coupled dynamics of C, P and U cycling associated with changes in
206 temperature, weathering and oceanic anoxia, in response to hypothesized CO_2 perturbations. The C,
207 P and U cycles are coupled together with the following processes (Fig. S2): (i) LIP CO_2 emission, which
208 acts as a source of C; (ii) seafloor spreading, which acts a source of C and a sink of U; (iii) silicate
209 weathering, which is a sink of C and a source of P and U; (iv) primary production of organic matter,
210 which is controlled by P availability and whose burial is a sink of C; (v) oxygen demand in the ocean,
211 which is controlled by primary productivity and organic-matter remineralization rates, and determines

212 the extent of seafloor anoxia and the burial of U(IV) in anoxic sinks. The burial of organic C also results
213 in an input of O₂ to the atmosphere. The U-isotope mass-balance is included in the model, allowing
214 calculated $\delta^{238}\text{U}_{\text{sw}}$ to trace the predicted extent of seafloor anoxia, whilst accounting for changes in U
215 inputs linked to silicate weathering. Similarly, $\delta^{13}\text{C}$ traces relative changes in the sources and sinks of
216 inorganic and organic C.

217 The model is set up for mid-Cretaceous boundary conditions and the sensitivity to uncertainty
218 in these parameters is discussed in the SI. Background atmospheric O₂ is a particularly important
219 boundary condition. Varying O₂ in the model serves as a convenient way to account for uncertainty in
220 the pre-existing degree of anoxia during the pre-OAE 2 interval, and hence the size of the ocean U
221 reservoir (Fig. 3) and the sensitivity of the U cycle to CO₂ perturbations. Estimates of mid-Cretaceous
222 atmospheric O₂ suggest greater than modern levels (49, 51, 53). Here, three baseline settings (O₂ =
223 1.0, 1.1 and 1.2 times present atmospheric levels; PAL) are used to provide an envelope of uncertainty
224 that reasonably captures pre-OAE 2 conditions, satisfies U cycle constraints, and covers the range of
225 uncertainty in $\delta^{238}\text{U}_{\text{carb}}$ (Fig. 2). The highest O₂ concentration setting (1.2 PAL) is the least sensitive to
226 CO₂ perturbations and predicts modern levels of pre-OAE seafloor anoxia (0.3%) with the largest
227 oceanic U reservoir (120% of today's size (16), and a higher baseline $\delta^{238}\text{U}_{\text{sw}}$ of -0.30‰ (compared to
228 modern seawater of $-0.39 \pm 0.01\text{‰}$ (18, 21, 25, 30, 54).

229 **AE1.** Two CO₂ emission pulses were modelled, corresponding to AE1 and AE2, where the rate
230 of CO₂ injection was kept constant for two discrete intervals of 150 kyrs. According to the C–P–U
231 model, the $\delta^{238}\text{U}_{\text{carb}}$ excursion during AE1 can be explained by a CO₂ degassing event of $0.9\text{--}2.25 \times 10^{18}$
232 mol C (i.e. $0.6\text{--}1.5 \times 10^{13}$ mol C yr⁻¹), which represents ~7 to 19% of current total degassing estimates
233 for the combined Caribbean and Madagascar LIPs (see SI). This perturbation results in an increase of
234 atmospheric CO₂ to 3.4–4.8 PAL (Fig. 3) and warming of 1.1–2.3°C, similar to independent proxy
235 reconstructions (39, 41, 45). The resultant increase in silicate weathering drives a ~16–39% increase
236 in P input (F_{Pw} ; Fig. S4), thereby increasing seawater P concentrations ([P]), promoting the expansion
237 of anoxia and the sequestration of reduced U(IV), and preferentially ²³⁸U, in anoxic sediments. The
238 response of ocean nutrient levels, anoxia, and $\delta^{238}\text{U}_{\text{sw}}$ are delayed compared to the CO₂ emission, with
239 the maximum extent of seafloor anoxia occurring up to ~100 kyr after the end of the CO₂ injection,
240 and covering between 8 and 15% of the total seafloor area (Fig. 3 and S4). This temporal lag is a
241 function of the positive P feedback mechanism (see Fig. S4). When overlain by anoxic waters,
242 sedimentary P burial efficiency is decreased (10, 11, 50, 51) due to i) the decreased flux of P absorbed
243 to Fe-oxides (F_{FeP}) and ii) an increased C/P of buried organic matter ($(\text{C/P})_{\text{organic}}$), which suppresses
244 productivity-driven increases in the burial of organic bound P (F_{OrgP}). These processes inhibit seawater

245 P depletion in the model and fuel further productivity even when external nutrient inputs are
246 declining, a result which is consistent with published sedimentary P data for OAE 2 (12).

247 **PCE.** Carbon sequestration from silicate weathering and the burial of marine-derived organic
248 matter increase during AE1 by 16–30% and 20–33% respectively, compared with pre-OAE 2 conditions
249 (Fig. S4). These processes drive the decrease of atmospheric CO₂ after the LIP emission ended,
250 resulting in a temperature and weathering decrease, and hence also a decrease in F_{Pw} (Fig. S4). The
251 combination of decreased F_{Pw}, continued high Ca-bound P burial (F_{CaP}) and high F_{OrgP} burial, leads to a
252 [P] decrease (Fig. S4), and hence biological productivity also decreases. The burial of organic matter
253 under anoxic conditions further drives a minor (~0.02 PAL) net increase in atmospheric O₂ during AE1
254 (Fig. S4). As a result of these controls, ocean oxygenation progresses and is positively re-enforced by
255 the enhanced P removal from increased F_{FeP} burial and decreased (C/P)_{organic} (Fig. S4). In response,
256 δ²³⁸U_{sw} recovers relatively quickly as oceanic anoxia declines, due to the depleted global U reservoir,
257 whilst [U] recovers more slowly, reaching maximum values up to 100 kyrs later (Fig. 3). This process
258 could explain the stratigraphic lag between increasing δ²³⁸U_{carb} and U/Ca during the PCE (Fig. 1 and 2;
259 see the below discussion). During the onset of oxygenation and increase in δ²³⁸U_{sw}, δ¹³C also continues
260 to rise owing to the slower dynamics of the larger C reservoir, again consistent with the trends shown
261 at Eastbourne for the start of the PCE (Fig. 2). Eventually, the decreased burial of organic matter, and
262 thus isotopically light carbon, during the reoxygenation episode begins to generate the carbon-isotope
263 ‘trough’ that is used for the chemo-stratigraphic correlation of the PCE (35, 45, 55, 56).

264 Despite the δ²³⁸U_{carb} increase during the PCE being consistent with the observations of cooler
265 conditions and local benthic reoxygenation (35, 36, 44-47, 57), the magnitude and abrupt rise of the
266 δ²³⁸U_{carb} at Eastbourne is difficult to simulate with the C–P–U model, assuming that this feature is not
267 an artifact of a condensed sedimentary interval. Indeed, the modelled δ²³⁸U_{sw} recovery better
268 corresponds to the later phase of the PCE, when TM_{redox} also recover. The rapidity of the δ²³⁸U_{carb}
269 increase, and delay of the TM_{redox} enrichments, implies a greater sensitivity of δ²³⁸U_{carb} to changing
270 redox conditions, compared to trace-metal concentrations. Such sensitivity was likely driven by an
271 additional decrease in the dynamic residence time of U, already greatly reduced to a minimum of ~45
272 kyrs (O₂ = 1 scenario) during AE1, due to the burial of U(IV) in anoxic sinks. This reduction might have
273 been accomplished by further U drawdown in the ‘hypoxic’ U sink (i.e. sediments that are reducing at
274 depth, beneath a fully, or poorly oxygenated water column (19)), which is not accounted for in the
275 model, and would have had little impact on δ²³⁸U_{carb}. Under such conditions δ²³⁸U_{carb} would more
276 precisely capture the onset of reoxygenation than TM_{redox}. Alternatively, this model–data discrepancy
277 may be explained by the oxidation and loss of U from shales previously deposited under anoxic
278 conditions, as has been observed in modern continental-margin sediments (58), and is similar to the

279 mechanism proposed to explain transiently light $\delta^{34}\text{S}_{\text{CAS}}$ values during the PCE (15, 35). This additional
280 supply of isotopically heavy U to the ocean would act to amplify the $\delta^{238}\text{U}_{\text{sw}}$ increase due to the
281 decrease in anoxic sinks, and could further explain the faster response compared to TM_{redox} . It is also
282 plausible that reoxygenation could have been accelerated by mechanisms not considered in our
283 model, such as sea level or temperature changes (9, 59), or the additional supply of O_2 caused by pyrite
284 burial (49, 53).

285 **AE2.** An assumed second CO_2 injection, of lesser magnitude ($0.4\text{--}1.05 \times 10^{13} \text{ molC yr}^{-1}$) but of
286 the same duration as the first event, generates a return to warmer conditions after the PCE. This
287 increase in temperature results in a return of widespread anoxia and a smaller negative shift in $\delta^{238}\text{U}_{\text{sw}}$
288 by -0.2% , equivalent to an expansion of oceanic anoxia of 3–8% of the total seafloor area. This
289 modelled CO_2 injection also results in the addition of isotopically light carbon to the ocean-
290 atmosphere system that augments the decrease in ^{12}C -enriched carbon burial during the PCE
291 reoxygenation episode, thereby accelerating the development of the $\delta^{13}\text{C}$ ‘trough’ of the PCE. Despite
292 this minor negative excursion, relatively elevated $\delta^{13}\text{C}$ values occur throughout the entire OAE 2
293 interval owing to the slower dynamics of the relatively large carbon reservoir.

294 **Wider Implications for OAEs.** Characterizing the timing, duration and extent of oceanic anoxia
295 during OAE 2 is fundamental to understanding the driving mechanisms of these extreme perturbations
296 to the carbon cycle and climate system. In an attempt to replicate the distribution of anoxia indicated
297 by local redox proxies, Earth System modelling suggests 50% of the global ocean volume was dysoxic
298 or anoxic (9) whilst the shale-derived $\delta^{238}\text{U}$ estimate suggests only 1–2% of the seafloor was overlain
299 by anoxic waters (14). By contrast, the new carbonate-derived $\delta^{238}\text{U}$ dataset presented here provides
300 a better systematic resolution of the timing and magnitude of ocean deoxygenation (Fig. 1), suggesting
301 that between 8% and 15% of the seafloor was overlain by an anoxic water column. This reconstruction
302 is in close agreement with recent estimates from thallium isotopes (60) and is compatible with 2–5%
303 of seafloor being euxinic, based on modelling of the sulfur- and molybdenum-isotope systems (15, 46,
304 61), as this extreme chemical state would only make up a fraction of the total anoxic ocean, and $\delta^{238}\text{U}$
305 does not discriminate between anoxic and euxinic redox states. The importance of differentiating
306 these two redox states is increasingly being recognized for periods of expanded anoxia (e.g. 62).
307 Crucially, the high resolution of the $\delta^{238}\text{U}_{\text{carb}}$ data uniquely resolves at least two fluctuations between
308 episodes of globally expanded and contracted anoxia during the traditionally defined OAE 2 interval.
309 Such findings are likely applicable for other OAEs where there are suggestions of local redox and
310 temperature variability, such as the Aptian OAE 1a (1).

311 The C–P–U model successfully generates the magnitude of U-isotope excursions through fairly
312 conservative CO_2 degassing levels and subsequent changes in global temperature. These amounts of

313 CO₂ degassing probably represent upper estimates since the model does not consider many
314 contributory deoxygenation mechanisms, including the inverse relationship between temperature
315 and O₂ solubility, or changes in O₂ supply through variable ocean ventilation rates (9). We suggest that
316 these additional mechanisms are of second-order importance to changes in terrestrial weathering for
317 driving OAE 2, although they could help refine our understanding of the precise temporal and
318 geographical distribution of oceanic redox changes.

319 In summary, the coupled geochemical and modelling approach adopted here for OAE 2 allows
320 for an internally consistent examination of global-scale deoxygenation during carbon cycle
321 perturbation events. The new U-isotope dataset highlights globally significant oscillations in oceanic
322 redox conditions within the traditionally defined OAE 2 interval that are closely coupled to changes in
323 global temperature and hydrological regime. The C–P–U model successfully reproduces the trends in
324 $\delta^{238}\text{U}_{\text{carb}}$ through discrete atmospheric CO₂ injections, suggesting that the exceptional longevity of OAE
325 2 was the result of multi-phase LIP activity.

326

327 **Materials and Methods**

328 **Uranium-isotope analysis.** Carbonate samples were mechanically cleaned for weathered
329 surfaces. Sample powders underwent oxidative-reductive cleaning (63, 64), before selective digestion
330 using 1 M buffered sodium acetate. U-isotope composition was determined using a ²³⁶U-²³³U double
331 spike to correct for instrumental mass fractionation, where uranium was pre-concentrated by co-
332 precipitation and purified by ion-exchange chromatographic procedures using a single column packed
333 with UTEVA resin (Eichrom Technologies, USA) (20, 21, 23, 65, 66). Uranium-isotope measurements
334 were performed by multiple-collector inductively coupled plasma mass spectrometer (MC-ICP-MS) at
335 the Centre for Trace Element Analysis, University of Otago, New Zealand (19, 21-23). **Lithium-isotope**
336 **methods.** An additional 8 samples were analyzed from the Eastbourne section for Li isotopes, to
337 increase the resolution of the already existing dataset, following the same original protocols (8).
338 Carbonate samples were leached using 0.1 M HCl for 1 hr and sample solutions were purified using a
339 two-stage cation-exchange procedure. Analyses were performed by MC-ICP-MS at Oxford University,
340 U.K. (8). **Coupled C–P–U model.** The box-model is a simplified version of elements of the Geochemical
341 Carbon cycle model (GEOCARB) and the Carbon, Oxygen, Phosphorus, Sulphur and Evolution (COPSE)
342 Earth System model (11, 13, 49-52), focusing exclusively on the shorter term processes of the carbon
343 and phosphorus cycles. The model incorporates U reservoirs and the isotopic mass-balance equations
344 for C and U isotopes as tracers of the predicted outcomes to assumed CO₂ emission perturbations.

345

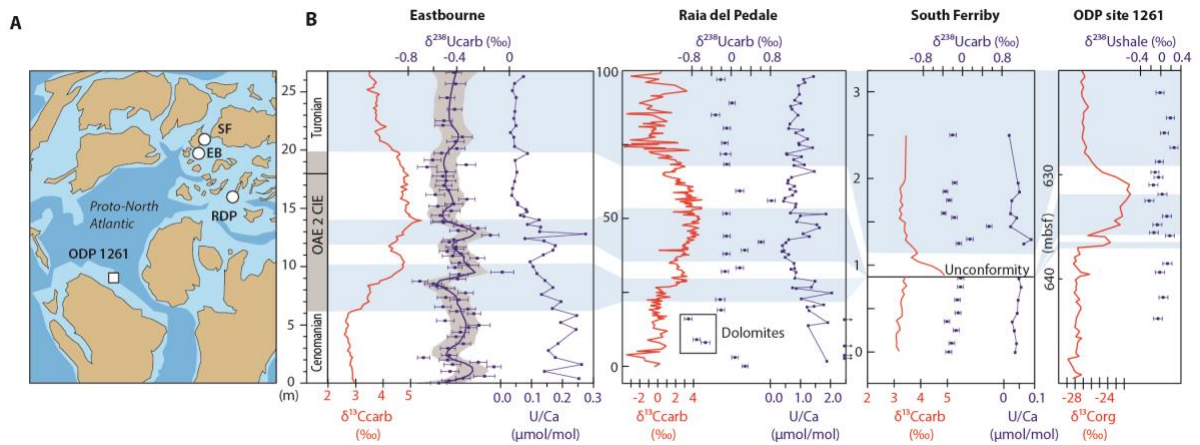
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353 CONTROLPASTCO2.

354

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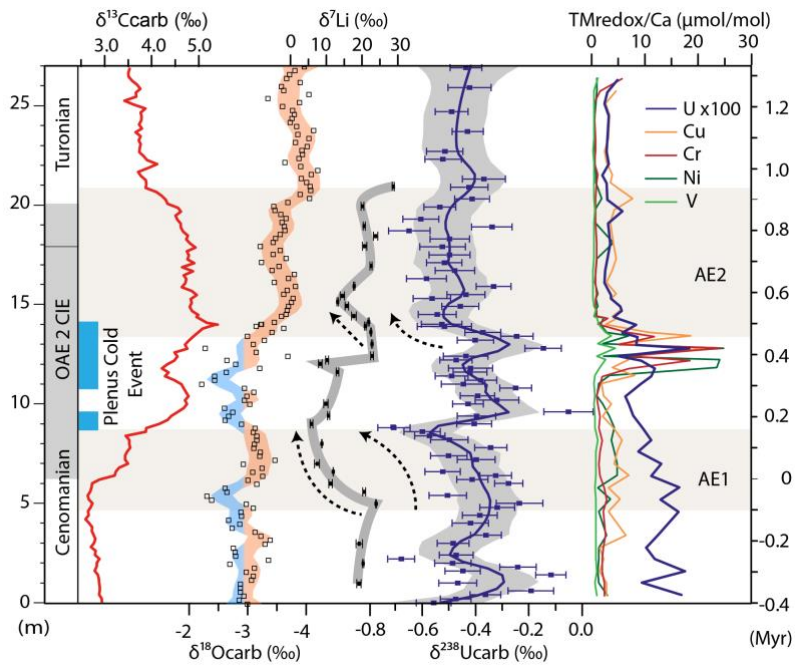
356



357

358 **Fig. 1:** A. Paleogeographic reconstruction showing locations of Eastbourne (EB), Raia del Pedale (RDP),
359 South Ferriby (SF) and ODP site 1261, modified with permission from (67) B. $\delta^{13}\text{C}$ (15, 34, 68), $\delta^{238}\text{U}_{\text{carb}}$
360 ($\pm 2\text{SE}$) and U/Ca (35 and this study) for EB, RDP and SF in comparison to $\delta^{13}\text{C}_{\text{Org}}$ and $\delta^{238}\text{U}$ from black
361 shales at ODP site 1261 (14, 55). Correlations (blue shading) are from previous work (8, 15, 35).

362

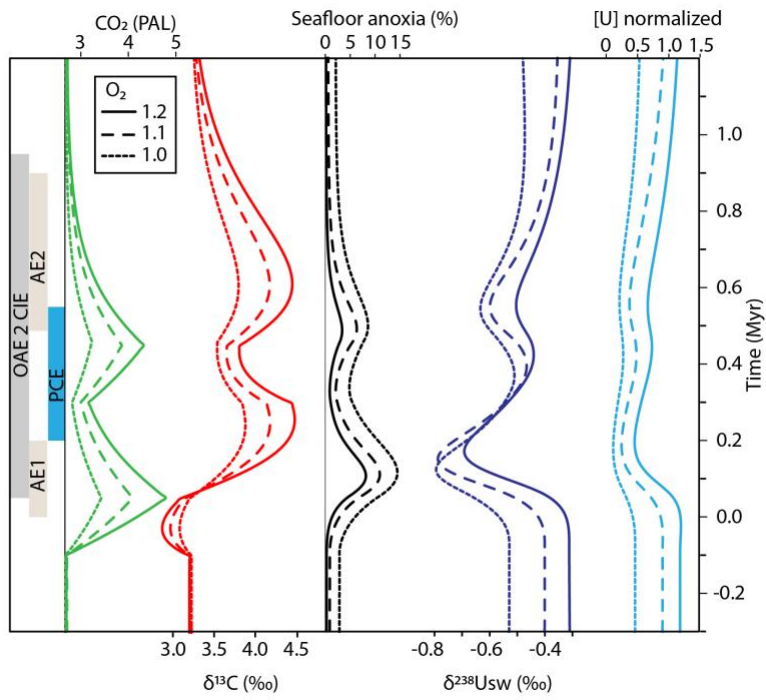


364

365 **Fig. 2:** Complementary geochemical datasets ($\delta^{13}\text{C}$, $\delta^{18}\text{O}$ and $\delta^7\text{Li}$) (8, 34), together with the $\delta^{238}\text{U}_{\text{carb}}$
 366 and TM_{redox} datasets for Eastbourne (35). ‘Anoxic Expansion’ episodes (AE1 and AE2) are identified
 367 from coupled negative $\delta^{238}\text{U}_{\text{carb}}$ excursions and TM_{redox} decreases. Unusual boreal faunas (blue bar)
 368 and less negative $\delta^{18}\text{O}$ are indicative of cooler waters and mark the ‘Plenus Cold Event’ (PCE) (35, 38,
 369 56). Black dashed arrows indicate periods of broad coupling between $\delta^7\text{Li}$ and $\delta^{238}\text{U}_{\text{carb}}$. The positive
 370 $\delta^{13}\text{C}$ excursion (34) is traditionally used to define OAE 2 (grey shading). Timescale (right axis) is based
 371 on estimates for OAE 2 of up to ~ 900 kyrs (4) and a linear age model, with 0 kyrs marking the start of
 372 the CIE.

373

374



375

376

377 **Fig 3:** C–P–U model outputs in relation to events of OAE 2 identified in the $\delta^{238}\text{U}_{\text{carb}}$ dataset of this

378 study (upper panel). Time (vertical axis) is given relative to the approximate start of the CIE, as in Fig.

379 2.

380

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597

598 **Supplementary Information**

599 **SI Materials and Methods**

600

601 **Reagents.** All reagents used were of ultra-high purity and prepared in-house by sub-boiling quartz
602 and/or teflon distillation (HF, HNO₃, HCl, CH₃COOH) or purchased commercially from Thermo Fisher
603 or Sigma Aldrich, NZ.

604

605 **Sample Preparation.** Weathered surfaces were removed from samples under dry conditions using a
606 Dremel microdrill before sonication in high-purity Milli-Q water (>18.2 MΩ cm; Millipore, USA) to
607 remove surface contaminants. Samples were dried in an oven at 40°C and then powdered in a laminar
608 flow hood (Class 100; ISO 5) using an agate pestle and mortar, which was cleaned between samples
609 using 99.995% SiO₂ sand and high-purity ethanol. Powdered samples, comprising 1–2 g, were subject
610 to reductive-oxidative cleaning (63, 64) using 1 M hydrous hydrazine + 0.25 M citric acid in 16 M
611 ammonia solution, followed by 1% H₂O₂ in 0.1 M NaOH. Repeated MilliQ rinses were performed
612 between cleaning steps. Such cleaning techniques remove potential contamination from residual
613 organic matter and ferro-manganese coatings and are particularly important for extracting primary
614 U/Ca ratios in foraminifera tests (63) and δ²³⁸U signatures from basalts (69). Cleaned powders were
615 digested using 0.6 M or 1 M sodium acetate solution (buffered to a pH of 5) (70) at room temperature
616 for 48 hrs in order to selectively digest carbonate phases and avoid U contributions from detrital
617 silicates

618

619 **Ion Exchange Chemical Preparation of Samples for Uranium-Isotope Analysis.** Uranium-isotope
620 composition was determined using a ²³⁶U-²³³U double spike to correct for instrumental mass
621 fractionation by adapting methods reported previously (20, 21, 23, 65, 66). The digest supernates
622 were evaporated to dryness in the presence of excess 7 M HNO₃ to prevent the formation of sodium
623 acetate precipitates. The final nitrate precipitates were then dissolved in 1.5 M HNO₃. This digest was
624 subsampled, diluted and analyzed via quadrupole ICP-MS using an Agilent 7500 instrument (Agilent
625 Technologies, USA) at the Centre for Trace Element Analysis, University of Otago, New Zealand to
626 determine the approximate U concentration in order to inform the double-spike addition calculation.
627 Based on this U concentration, samples were again subsampled into PFA vials to achieve a total
628 uranium mass of 30 to 150 ng and double spiked to give a ²³⁶U/²³⁵U ratio of approximately 3. Solutions
629 were refluxed for 24 hrs to ensure complete spike-sample equilibration. Due to the high matrix
630 content of the carbonate digests, co-precipitation using pre-cleaned FeCl₃ and ammonia solution was
631 conducted to remove the majority of Na and Ca from the sample and prevent overwhelming the ion-

632 exchange resin with matrix elements during the separation and purification of U. The resulting
633 precipitates were separated from the high-matrix supernates through centrifugation and were then
634 fully dissolved in 6 M HCl. After evaporation to dryness, the samples were re-dissolved in 3 M HNO₃
635 and loaded onto pre-cleaned heat shrink Teflon columns containing UTEVA resin (Eichrom
636 Technologies, USA). Uranium separation and purification was achieved following previously published
637 protocols (20, 21, 23, 65, 66). Samples were evaporated to dryness and oxidised using H₂O₂ + HNO₃
638 for 24 hrs, to eliminate any organic residues from the resin, then evaporated once more. The purified
639 U fractions were re-dissolved in 0.25 M HCl + 0.05 M HF in preparation for U isotope analysis. Uranium
640 yields were typically > 95% and the total procedural blank was approximately 10–20 pg, which is
641 negligible for U sample sizes of between 30 to 150 ng.

642

643 **Uranium-Isotope Analysis.** Uranium isotopic measurements were performed by multiple collector
644 ICP-MS (MC-ICPMS) using a Nu-Plasma HR MC-ICPMS (Nu Instruments, UK) coupled to a DSN-100
645 desolvation system, at the Centre for Trace Element Analysis, University of Otago, following previously
646 published procedures (22, 23, 66). The ²³⁸U/²³⁵U composition is presented in delta-notation (δ)
647 following Eq. (1).

648

$$649 \quad \delta^{238}\text{U} = \left(\frac{(^{238}\text{U}/^{235}\text{U})_{\text{sample}}}{(^{238}\text{U}/^{235}\text{U})_{\text{CRM-145}}} - 1 \right) \times 1000 \quad (1)$$

650 The reference material CRM-145 (National Institute of Standards and Technology, U.S.A.) is
651 the internationally recognized δ-zero standard (19) with a ²³⁸U/²³⁵U ratio of 137.837 ± 0.015. Each
652 sample δ²³⁸U value was calculated relative to two bracketing CRM145 standards, double spiked in the
653 same way as samples, and with ion beam intensities matched to within 10%. Analytical uncertainties
654 are given as twice the standard error (2SE), and reflect the internal instrumental error in the
655 measurements of the sample and bracketing standards using standard techniques of error
656 propagation (23). On average, 2SE was equivalent to ±0.07%, and is comparable to the external
657 reproducibility (2SD) of repeat measurements of the same sample (21). Analytical performance has
658 been verified previously, based on the analysis of certified reference materials (20-23, 66). Replicate
659 analyses performed on samples within and between sessions showed excellent agreement, within the
660 magnitude of their 2SE analytical uncertainties (Fig. S1).

661

662 **Elemental Composition Analysis.** For spiking purposes and in order to identify diagenetic alteration
663 of samples, aliquots of samples were diluted appropriately and measured for their elemental
664 concentrations by quadrupole ICP-MS at the Centre for Trace Element Analysis, University of Otago,

665 New Zealand. Suitable dilutions of a NIST traceable multi-element standard (Choice Analytical Pty Ltd,
666 Australia) was used for instrumental calibration purposes, and an internal standard comprising a
667 cocktail of 7 reference elements was added to all samples to correct for instrumental drift during
668 measurement. The performance of the elemental analysis methods was checked by analysis
669 of samples spiked with elements of interest. Average recoveries of $102 \pm 4\%$ were obtained for the
670 elements reported here. In addition to spiked samples the ICP-MS method has been validated by
671 analysis of digested Basalt, Columbia River (BCR-2) and Basalt, Hawaiian Volcanic Observatory (BHVO-
672 2) certified reference materials (USGS), recoveries averaged $99\% \pm 5\%$. A change in digestion protocol
673 from 0.5 to 1M Sodium Acetate may have introduced variability in the U/Ca data, but does not affect
674 the measured $\delta^{238}\text{U}_{\text{carb}}$. To minimise this effect, secular trends in elemental concentrations for
675 Eastbourne and Raia del Pedale are identified using elemental composition data produced in a
676 previously published study (35) where 5 mg powders were digested using 2 ml of 0.5 M acetic acid.
677 Measurements were made using a Thermo Scientific Element 2 ICP-MS at the University of Oxford.

678

679 **Regression Analysis**

680 To capture the main features of the $\delta^{238}\text{U}_{\text{carb}}$ dataset at Eastbourne, without placing undue emphasis
681 on single outlying data points, a LOESS smoothing function has been applied to the U-isotope dataset
682 (Fig. 1 and 2). This was performed using the 'ggplot2' package in R, with the 'geom_smooth' function,
683 a default polynomial setting of 2, and $\alpha=0.15$. This also calculates the 95% confidence interval of the
684 LOESS regression. The majority of data points overlap with the 95% confidence interval of this
685 regression (grey shading Fig. 1 and 2), if their analytical uncertainties are taken into account,
686 suggesting that robust systematic trends can be identified. A smoothing function is not appropriate
687 for the Raia del Pedale and South Ferriby datasets due to their lower sampling density and
688 stratigraphic resolution.

689

690

691 **Model Description.**

692 **Summary.** The model is a simplified version of the GEOCARB and COPSE family of models (11, 13, 49-
693 51), focusing on the shorter term carbon cycle processes. Normalized forcing parameters are:
694 degassing (D) (linked to seafloor spreading rate), uplift driving erosion (E), and effects of vegetation
695 and lithology on weatherability (W). The model is perturbed by an atmospheric CO₂ emission from LIP
696 volcanism and calculates the response of the C, P and U cycles (Fig. S2). Briefly, silicate weathering
697 responds to temperature and acts as a sink of C through carbonate burial, and a source of P and U to
698 the ocean. Organic matter production is dependent on P concentrations, and acts as a sink for P and
699 C. The extent of oceanic anoxia is controlled by O₂ supply and O₂ consumption through the
700 remineralization of organic matter. Anoxic sediments act as a sink for U and C. P concentrations are
701 also influenced by the extent of anoxia due to the redox sensitive burial of Fe-bound and organic-
702 bound P. The U-isotope mass balance is then used to trace the extent of seafloor anoxia and the C-
703 isotope mass balance to trace the relative proportion of organic C burial.

704

705 **Carbon and phosphorus cycles.** The model considers both inorganic and organic parts of the carbon
706 cycle. The mass balance for the ocean-atmosphere reservoir of inorganic carbon (A) is given by:

$$707 \quad dA/dt = F_d - F_w + F_{ox} - F_{morg} - F_{torg} (+ F_{cw} - F_{cb}) + F_{LIP} \quad (2)$$

708 where the carbon fluxes are F_d = degassing at seafloor spreading centres, F_w = silicate weathering
709 (followed by carbonate burial), F_{ox} = input of organic carbon from oxidative weathering and degassing,
710 F_{morg} = marine-derived organic-carbon burial, and F_{torg} = terrestrially derived organic-carbon burial.
711 Carbonate weathering, F_{cw} is assumed to be balanced by a corresponding carbonate burial flux, F_{cb}
712 (i.e. $F_{cb}=F_{cw}$) so these terms do not affect the dynamics. F_{LIP} is an additional carbon input perturbation
713 due to LIP emplacement; $F_{LIP} = 0$ at present.

714 For present-day fluxes (in 10^{12} molC yr⁻¹) the terms $F_d = k_d = 8$, $F_w = k_w = 8$, $F_{ox} = k_{ox} = 9$, $F_{morg} = k_{morg} =$
715 4.5 , $F_{torg} = k_{torg} = 4.5$, and $F_{cb} = F_{cw} = k_c = 16$, based on COPSE (and hence GEOCARB) are assigned. The
716 present-day reservoir $A_0 = 3.2 \times 10^{18}$ molC, giving a residence time with respect to silicate weathering
717 removal of 400 thousand years (kyr), or with respect to silicate weathering plus organic-carbon burial
718 of 188 kyr.

719 The degassing flux is given by:

$$720 \quad F_d = k_d * D \quad (3)$$

721 The silicate weathering flux is taken to depend on temperature, CO₂, uplift, and the effects of
722 vegetation and lithology (e.g. basalts versus granites) on weatherability:

$$723 \quad F_w = k_w * E * W * f(\text{CO}_2) * f(T) \quad (4)$$

724 Existing models assume that weathering is predominantly biotic by the mid-Cretaceous and therefore
725 $f(\text{CO}_2)$ is the response of vegetation (rather than abiotic weathering) to CO₂ following Michaelis-
726 Menten kinetics (after GEOCARB III). Here, CO₂ concentration is given in PAL, representing multiples
727 of present atmospheric level:

$$728 \quad f(\text{CO}_2) = 2 * \text{CO}_2 / (1 + \text{CO}_2) \quad (5)$$

729 The proportion of inorganic carbon that resides in the atmosphere as CO₂ scales with total ocean-
730 atmosphere inorganic carbon inventory (A)

$$731 \quad \text{CO}_2 = (A/A_0)^2 \quad (6)$$

732 The effect of changes in CO₂ and solar luminosity on temperature (from GEOCARB III) is:

$$733 \quad \Delta T = k_{\text{CO}_2} * \ln(\text{CO}_2) - k_{\text{SL}} * (t/570) \quad (7)$$

734 where $k_{\text{CO}_2} = 4^\circ\text{C}$ (corresponding to a present-day climate sensitivity of about 2.8°C for doubling CO₂)
735 and $k_{\text{SL}} = 7.4^\circ\text{C}$ for t in millions of years ago, which gives $\sim 1.2^\circ\text{C}$ cooling at 94 Ma due to lower solar
736 luminosity.

737 The dependence of weathering on temperature neglects temperature-driven changes in runoff and
738 only considers the kinetic effect of temperature on mineral dissolution, which using an activation
739 energy appropriate for granite is given by:

$$740 \quad f(T) = \exp(0.09 * \Delta T) \quad (8)$$

741 The input of organic carbon from oxidative weathering and degassing is kept constant for simplicity
742 (although oxidative weathering depends on uplift and organic-carbon degassing depends on
743 degassing):

$$744 \quad F_{\text{ox}} = k_{\text{ox}} \quad (9)$$

745 Terrestrial-derived organic-carbon burial is assumed to be fertilised by CO₂ (consistent with the
746 weathering formulation):

$$747 \quad F_{\text{torg}} = k_{\text{torg}} * f(\text{CO}_2) \quad (10)$$

748 Marine-derived organic-carbon burial is taken to depend on ocean phosphorus concentration
749 (normalized to modern values):

$$750 \quad F_{\text{morg}} = k_{\text{morg}} * (P/P_0) \quad (11)$$

751 Thus, to determine marine-derived organic-carbon burial the phosphorus cycle needs to be
752 considered.

753 Phosphorus weathering is assumed to be tied to silicate weathering:

$$754 \quad F_{\text{Pw}} = k_{\text{Pw}} * (F_w/F_{w0}) \quad (12)$$

755 where $k_{\text{Pw}} = 72 \times 10^9 \text{ molP yr}^{-1}$.

756 The ocean P balance is given by:

$$757 \quad dP/dt = F_{\text{Pw}} - F_{\text{OrgP}} - F_{\text{FeP}} - F_{\text{CaP}} \quad (13)$$

758 where F_{OrgP} is burial associated with organic matter, F_{FeP} is burial adsorbed to iron oxides, F_{CaP} is burial
759 associated with carbonate minerals. For present day fluxes ($\times 10^9 \text{ molP yr}^{-1}$) we assign $k_{\text{OrgP}} = 18$, $k_{\text{FeP}} =$
760 18 , $k_{\text{CaP}} = 36$.

761 Organic phosphorus burial is sensitive to the anoxic fraction of the ocean, f_{anoxic} , given very different
762 $(C/P)_{\text{organic}}$ burial ratios under anoxic ($CP_{\text{anoxic}}=1000$) and oxic ($CP_{\text{oxic}}=250$) bottom waters (52):

$$763 \quad F_{\text{OrgP}} = F_{\text{morg}} * ((f_{\text{anoxic}}/CP_{\text{anoxic}}) + ((1-f_{\text{anoxic}})/CP_{\text{oxic}})) \quad (14)$$

764 Iron oxide-bound P burial is also sensitive to the anoxic fraction of the ocean:

$$765 \quad F_{\text{FeP}} = k_{\text{FeP}} * (1-f_{\text{anoxic}}) \quad (15)$$

766 Calcium-bound phosphorus burial is assumed to scale with P input to sediments from sinking organic
767 matter (with no redox dependence):

$$768 \quad F_{\text{CaP}} = k_{\text{CaP}} * (P/P_0) \quad (16)$$

769 The function defining the fraction of seafloor overlain by anoxic waters, f_{anoxic} , adopts a logistic
770 functional form dependent on the balance between normalized oxygen demand ($k_U * (P/P_0)$) and
771 normalized oxygen supply (O_2/O_{20}), where k_U represents the efficiency of nutrient utilization in
772 upwelling regions:

$$773 \quad f_{\text{anoxic}} = 1 / (1 + e^{-k_{\text{anox}} * (k_U * (P/P_0) - (O_2/O_{20}))}) \quad (17)$$

774 Here, $k_U = 0.5$ is chosen based on observations of high-latitude nutrient utilization and agreement
 775 across a range of models (9, 59, 71, 72) that a transition from an oxic to an anoxic deep ocean occurs
 776 at $O_2/O_{20} \sim 0.5$ PAL (for present nutrient levels), or at $P/P_0 \sim 2$ (for present O_2 levels). $k_{anox} = 12$ is chosen
 777 based on observations that $\sim 0.2\text{--}0.3\%$ of the seafloor is overlain by anoxic bottom waters at present
 778 (73), and model agreement that the transition from oxic to anoxic conditions is abrupt as, for example,
 779 O_2/O_{20} drops from ~ 0.6 to ~ 0.4 (9, 59, 72).

780 We make a simple consideration of the effect of increases in atmospheric O_2 by integrating any
 781 increases in organic-carbon burial above the initial steady state to give a cumulative input of O_2 ,
 782 calculating the resulting change in O_2/O_{20} (where the present $O_{20} = 3.7 \times 10^{19}$ mol), and feeding this
 783 change in O_2/O_{20} back into f_{anoxic} . We do not consider longer term negative feedbacks on O_2 because
 784 of the short interval under consideration here.

785 **Carbon-isotope mass balance.** The carbon-isotope mass balance follows the carbon mass balance:

$$786 \quad d(A \cdot \delta_A)/dt = F_d \cdot \delta_d + F_{ox} \cdot \delta_{ox} + F_{cw} \cdot \delta_{cw} + F_{LIP} \cdot \delta_{LIP} - (F_{morg} + F_{torg}) \cdot (\delta_A - \Delta) - (F_w + F_{cb}) \cdot \delta_A \quad (18)$$

787 The above equation can be simplified to:

$$788 \quad d(A \cdot \delta_A)/dt = F_{in} \cdot \delta_{in} + F_{LIP} \cdot \delta_{LIP} - F_{org}(\delta_A - \Delta) - F_{carb} \cdot \delta_A \quad (19)$$

789 where $F_{in} = F_d + F_{ox} + F_{cw}$ is the total steady state input flux of carbon of mean isotopic composition δ_{in}
 790 = -5 ‰ (as often assumed in other studies), the carbon input perturbation F_{LIP} has isotopic composition
 791 δ_{LIP} (see sensitivity analysis below), $F_{org} = F_{morg} + F_{torg}$ is the total burial flux of organic matter with
 792 fractionation Δ (taken here to be 25 ‰) and $F_{carb} = F_w + F_{cb}$ is the total burial flux of carbonates (with
 793 no assumed C-isotope fractionation).

794 Simplifying equation (19) in the same manner ($dA/dt = F_{in} + F_{LIP} - F_{org} - F_{carb}$), then applying the product
 795 rule ($d(A \cdot \delta_A)/dt = A \cdot d\delta_A/dt + \delta_A \cdot dA/dt$) and substituting and rearranging gives:

$$796 \quad d\delta_A/dt = (F_{in} \cdot (\delta_{in} - \delta_A) + F_{LIP} \cdot (\delta_{LIP} - \delta_A) - F_{org} \cdot (-\Delta))/A \quad (20)$$

797 With no perturbation ($F_{LIP} = 0$) at steady state ($d\delta_A/dt = 0$) this corresponds to the widely-used
 798 formulation:

$$799 \quad \delta_A = \delta_{in} + f \cdot \Delta \quad (21)$$

800 Where 'f' is the fraction of carbon buried in organic form i.e. $f = F_{org}/F_{in} = F_{org}/(F_{org} + F_{carb})$. For the
 801 modern fluxes suggested above $f = 9/33 = 0.273$ and $\delta_A \sim 1.8 \text{ ‰}$ if $\Delta = 25$.

802 Uranium-isotope mass balance

803 The U-cycle model assumes that the riverine input of U is driven by silicate weathering, all U sinks
804 scale linearly with U concentration, the hydrothermal sink of U also scales with degassing from
805 seafloor spreading (D), and the anoxic sink of U also scales with f_{anoxic} , the fraction of anoxic seafloor:

$$806 \quad dU/dt = F_{\text{riv}} - F_{\text{hyd}} - F_{\text{anoxic}} - F_{\text{carbonate}} - F_{\text{suboxic}} - F_{\text{oxic metals}} \quad (22)$$

$$807 \quad F_{\text{riv}} = k_{\text{riv}} * F_w / F_{w0} \quad (23)$$

$$808 \quad F_{\text{hyd}} = k_{\text{hyd}} * D * (U/U_0) \quad (24)$$

$$809 \quad F_{\text{anoxic}} = k_{\text{anoxic}} * (U/U_0) * f_{\text{anoxic}} / f_{\text{anoxic0}} \quad (25)$$

$$810 \quad F_{\text{carbonate}} + F_{\text{suboxic}} + F_{\text{oxic metals}} = F_{\text{other}} = k_{\text{other}} * (U/U_0) \quad (26)$$

811 For a modern $U=1.85 \times 10^{13}$ mol, we use estimates of $k_{\text{anoxic}} = 6.2 \times 10^6$ mol yr⁻¹, $k_{\text{hyd}} = 5.7 \times 10^6$ mol yr⁻¹,
812 $k_{\text{other}} = 36 \times 10^6$ mol yr⁻¹, and therefore $k_{\text{riv}} = 47.9 \times 10^6$ mol yr⁻¹ (16, 19). The corresponding isotope mass
813 balance is:

$$814 \quad d\delta_U/dt = (F_{\text{riv}} * (\delta_{\text{riv}} - \delta_U) - F_{\text{hyd}} * \Delta_{\text{hyd}} - F_{\text{anoxic}} * \Delta_{\text{anoxic}} - F_{\text{other}} * \Delta_{\text{other}}) / U \quad (27)$$

815 (Note here that Δ are all positive for U isotopes, in contrast to Δ for C isotope fractionation.) For the
816 modern (18, 19, 30, 54) we assume $\delta_{\text{riv}} = -0.29$, $\Delta_{\text{hyd}} = 0.2$, $\Delta_{\text{anoxic}} = 0.5$, and $\delta_U = -0.39$, requiring $\Delta_{\text{other}} =$
817 0.0156 for steady state.

818

819 Mid-Cretaceous model initialisation, U cycle assumptions and perturbation scenarios

820 **Boundary conditions:** The mid-Cretaceous had elevated rates of seafloor spreading and
821 corresponding degassing, with $D=1.5$, as used in GEOCARB/COPSE models. At the time of OAE 2, an
822 increase in hydrothermal input by 20% has been estimated to be consistent with an observed Sr-
823 isotope shift (8, 74) hence we use $D=1.8$ as the baseline value. Uplift was considerably lower in the
824 mid-Cretaceous than at present, with $E=0.7$ from GEOCARB/COPSE used as a baseline. The rise of
825 angiosperms is assumed to have finished increasing weathering rates by 100 Ma, following
826 GEOCARB/COPSE. Basalt area is estimated to be comparable to today (75). Therefore, present
827 weatherability $W = 1$ is used as a baseline. Lower solar luminosity at 94 Ma translates to 1.2°C of
828 cooling. Higher steady-state atmospheric O₂ is considered as part of our sensitivity analysis.

829 **U Isotope Fractionation Factor:** The sensitivity of changes in calculated $\delta^{238}\text{U}_{\text{sw}}$, in terms of
830 reflecting the degree of anoxia, is dependent on the assumed magnitude of fractionation during U
831 reduction between anoxic sediments and seawater (Δ_{anoxic}). Modern observational, theoretical and
832 experimental data suggest that the reduction of U(VI) to U(IV) is associated with a positive U-isotope
833 fractionation factor of between 0.4 to 1.3 ‰ (14, 19, 20, 23-25, 30, 76) where the apparent
834 enrichment factor in natural settings may be highly variable and dependent on local geochemical and
835 depositional conditions (18, 20, 30), which are hard to quantify on a global scale during Earth's history.
836 To this end, previous mass balance models for the Permo-Triassic mass extinction and OAE 2 have
837 employed various enrichment factors of +0.5 (14, 26), +0.6 (28) and +0.77 ‰ (29), which generate
838 significantly different estimates for the size of anoxic sinks, and hence seafloor area covered by anoxic
839 waters, required to drive a given negative $\delta^{238}\text{U}$ excursion. The assumption of Δ_{anoxic} remains a
840 fundamental limitation for the interpretation of ancient $\delta^{238}\text{U}$ data. Due to this uncertainty, we use
841 the observed difference in $\delta^{238}\text{U}$ between contemporaneous anoxic shales (ODP site 1261, Fig. 1)
842 where the average post-OAE value reported for ODP 1262 is 0.063‰ (14) and Eastbourne is -0.45‰,
843 giving a Δ_{anoxic} of +0.51‰, which is within the range of observational data from modern anoxic basins
844 (14, 19, 20, 23-25, 30, 76). Sensitivity analysis (Table S2) demonstrates that increasing Δ_{anoxic} to 0.6‰,
845 would increase the magnitude of the negative $\delta^{238}\text{U}_{\text{sw}}$ excursion by 0.1‰ for the same CO_2 emission
846 scenario and size of anoxic sinks. To match the observed excursions in Eastbourne when $\Delta_{\text{anoxic}} = 0.6‰$
847 then requires a decrease in the perturbation magnitude, and lower estimates of the area of seafloor
848 anoxia of 4 to 10%. If Δ_{anoxic} were assumed to be much larger this drives the steady state $\delta^{238}\text{U}_{\text{sw}}$ lower
849 than the range recorded at Eastbourne and is not deemed plausible without invoking unrealistically
850 high δ_{riv} values, or invoking a diagenetic offset for which there is no evidence. It would also require
851 smaller changes to the carbon cycle model, with a smaller CO_2 emission, temperature increase and
852 relative change in productivity, in order to drive the same magnitude of $\delta^{238}\text{U}$ excursion. The result of
853 this would be to decrease the amount of carbon removed from the ocean-atmosphere system through
854 productivity and burial under anoxic conditions, and therefore create a decrease in the magnitude of
855 the calculated positive CIE. Using a Δ_{anoxic} of +0.5‰ generates a good agreement between the model
856 outputs for temperature and estimates from proxy reconstructions, and produces a $\delta^{13}\text{C}$ excursion
857 that is only ~0.5‰ lower than observed at Eastbourne. Thus we do not consider the higher
858 fractionation factor/smaller perturbation scenario to be realistic for OAE 2. The internally consistent
859 carbon cycle model therefore helps increase the confidence in our interpretation of the area of
860 seafloor that was overlain by an anoxic water column.

861 **U riverine input:** The modern average measured riverine $\delta^{238}\text{U}$ composition (δ_{riv}) is
862 indistinguishable from bulk silicate earth (BSE) at -0.29‰ (18, 19, 21, 54, 69, 77). In the mid-

863 Cretaceous setup of the C–P–U model (see below), δ_{riv} is set to $\sim -0.19\text{‰}$ in order to generate $\delta^{238}\text{U}_{sw}$
864 values in agreement with the $\delta^{238}\text{U}_{carb}$ data from Eastbourne under different atmospheric O_2 scenarios.
865 Although this value is slightly higher than modern average δ_{riv} it is a plausible figure given the
866 uncertainty of exposed weatherable lithologies on land. Indeed, it has been suggested that δ_{riv} could
867 vary by up to a few per mil for different periods of Earth’s history (28) although estimates are limited
868 by substantial uncertainties including the paleogeology reconstructions, degree of soil development,
869 vegetation cover and changing weathering regimes. The chosen value used here for δ_{riv} is also within
870 the range back-calculated through mass-balance modelling of the modern U cycle ($\delta_{riv} = -0.18$ to -0.28
871 ‰), which encompasses the uncertainty of the size and isotopic composition of modern U sinks (19).
872 Thus, using either the modern δ_{riv} that is close to BSE, or a slightly higher δ_{riv} of -0.19‰ makes little
873 difference within the uncertainty of the current characterization of U sinks.

874 **Resulting steady state:** With these changes in boundary conditions (at $\text{O}_2 = 1$) the predicted
875 $A/A_0 = 1.63$, $\text{CO}_2 = 2.7 \text{ PAL}$ and $\Delta T = +2.7^\circ\text{C}$ (i.e. global average $T \sim 18^\circ\text{C}$), comparing well with GEOCARB
876 III and the warmer world at the time. Predicted $P/P_0 = 1.43$, which assuming present O_2 generates
877 more anoxia ($f_{anoxic} = 0.03$). $\delta^{13}\text{C} \sim +3.2 \text{‰}$ is predicted thanks to elevated terrestrial and marine
878 organic-carbon burial (consistent with data from Eastbourne and South Ferriby). These changes in
879 boundary conditions lead to a predicted increase in U river input outweighed by an increase in the
880 anoxic sink, leading to a predicted decrease in ocean U content to $0.9 \times 10^{13} \text{ mol}$ (50% of present), and
881 when $\delta_{riv} = -0.19$ and $\Delta_{anoxic} = 0.5$ this gives a predicted $\delta_U = -0.53 \text{‰}$.

882 **LIP CO_2 injection:** A carbon cycle perturbation associated with the emplacement of the
883 Caribbean-Columbian, Madagascar and/or other LIPs close to the Cenomanian/Turonian boundary
884 has been hypothesized as the trigger for OAE 2. These two LIPs are each estimated at $4.5 \times 10^6 \text{ km}^3$
885 basalt (78) equivalent to mass $1.3 \times 10^7 \text{ Gt}$ (density 2.9 g cm^{-3}) that, assuming $2.0 \text{ wt}\% \text{ CO}_2$ (79), yields
886 $2.6 \times 10^5 \text{ GtCO}_2$ or $7.1 \times 10^4 \text{ GtC}$ or $5.91 \times 10^{18} \text{ molC}$. This figure gives a combined total carbon pool that
887 could be degassed throughout the emplacement of these two LIPs of $\sim 12 \times 10^{18} \text{ molC}$. The nature and
888 corresponding rate of degassing is uncertain. We consider release events of $\sim 10\text{--}20\%$ of the total as
889 plausible (i.e. $1.2\text{--}2.4 \times 10^{18} \text{ molC}$), which over 150 kyr amounts to $0.8\text{--}1.6 \times 10^{13} \text{ molC yr}^{-1}$, up to a
890 doubling of the degassing flux. Clearly multiple perturbations of approximately this magnitude could
891 have occurred during the full interval of LIP emplacement.

892 If the prescribed CO_2 injection is assumed to be solely mantle derived with a $\delta^{13}\text{C}$ signature of
893 -5‰ , then a minor negative CIE (of $\sim 1\text{‰}$) before the positive CIE of OAE 2 is predicted (Table 2).
894 However, there is little indication of this feature in the $\delta^{13}\text{C}$ record for OAE 2, unlike other similar
895 events such as the early Aptian OAE 1a ($\sim 120 \text{ Ma}$) (1), suggesting either a slower initial rate of CO_2

896 emission for OAE 2, where the potential negative excursion was counterbalanced by the burial of
897 organic matter, or that the degassed volatiles contained a mixture of mantle-derived and assimilated
898 carbonate C from overlying crustal material. Because of this potential complication, the injected
899 carbon is assumed to have a $\delta^{13}\text{C}$ signature of 0‰ in the model runs shown in Fig. 3. Due to the
900 inherent delay in the carbon cycle between the timing of the CO_2 injection and maximum extent of
901 oceanic anoxia, the maximum $\delta^{13}\text{C}$ predicted for the subsequent positive CIE is less sensitive to this
902 assumed input CO_2 signature, with only $\sim 0.1\%$ difference between the two scenarios (Table S2).

903 **Sensitivity analyses.** A series of sensitivity analyses show that decreasing background CO_2
904 degassing (D), uplift and erosion rates (E), or the weatherability of exposed landmasses (W), leads to
905 reduced levels of anoxia during the pre-OAE 2 interval and therefore higher predicted values of
906 $\delta^{238}\text{U}_{\text{sw}}$. Such boundary conditions make the calculated changes in anoxia, [U], $\delta^{238}\text{U}_{\text{sw}}$ and $\delta^{13}\text{C}$ less
907 sensitive to the same CO_2 perturbation. Importantly, for reasonable initial boundary conditions that
908 are consistent with more complex biogeochemical models, geologically plausible magnitudes of CO_2
909 increases can drive expanded anoxia, and inferred $\delta^{238}\text{U}_{\text{sw}}$ changes in all cases. In a first set of model
910 sensitivity analyses (Table S1) we vary the model boundary conditions of D, E, W, and atmospheric O_2
911 over plausible ranges. We consider $D=1.5$ as a lower limit (no extra increase in hydrothermal activity)
912 and $D=2.1$ (symmetric perturbation) as an upper limit. E is varied over ± 0.2 , which spans most of the
913 Phanerozoic range (13). W is varied over ± 0.2 given uncertainty in the effect of angiosperms on
914 weathering and in the background area of volcanic rocks at the time (75). For atmospheric O_2 following
915 recent work (53) we only consider increases and we limit these to a level (1.2 PAL) that returns
916 modern levels of ocean anoxia (given evidence for more widespread anoxia prior to OAE2). These
917 changes in boundary conditions alter the initial steady state and the core dynamics of the model, and
918 we examine how this changes the response to a common CO_2 injection perturbation of 1.8×10^{18} molC
919 over 150 kyr (1.2×10^{13} molC/yr). In a second set of sensitivity analyses (Table S2) we take the same
920 perturbation experiment from standard boundary conditions with $\text{O}_2 = 1$ PAL and vary the assumed
921 $\delta^{13}\text{C}$ of LIP CO_2 input, and δ_{riv} and Δ_{anoxic} in the U isotope system. These changes only affect the isotopic
922 expression of the modelled C, P, U changes.

923 **SI Discussion**

924 Primary biotic and abiotic calcite and aragonite precipitates have been shown to record ambient
925 solution $\delta^{238}\text{U}$ values accurately with minimal fractionation, despite structural co-ordination changes
926 with regard to calcite precipitation (21, 25, 31, 32). These studies also suggest that there is no
927 appreciable fractionation driven by biotic ‘vital effects’, consistent with the fact that U does not serve
928 a biological function. A minor positive fractionation factor has been observed for abiotic aragonite
929 formation under very specific experimental conditions, whereby the solid phase has a more positive

930 $\delta^{238}\text{U}$ than the residual solution (32) although the applicability of these experimental results to real-
931 world conditions is still not well understood. This complication could only have potentially effected
932 Raia del Padale, which was originally comprised of mixed aragonite and calcite that is typical of
933 shallow-water platform settings. U concentrations are also vastly different between calcite and
934 aragonite, with aragonite being up to two orders of magnitude greater. This difference is clearly seen
935 in the greater U/Ca ratios of the Raia del Pedale carbonates (Fig. 1) and the mixed mineralogy of this
936 section could also explain the greater degree of scatter. Despite this the general trends in U/Ca are
937 replicated at the two sites.

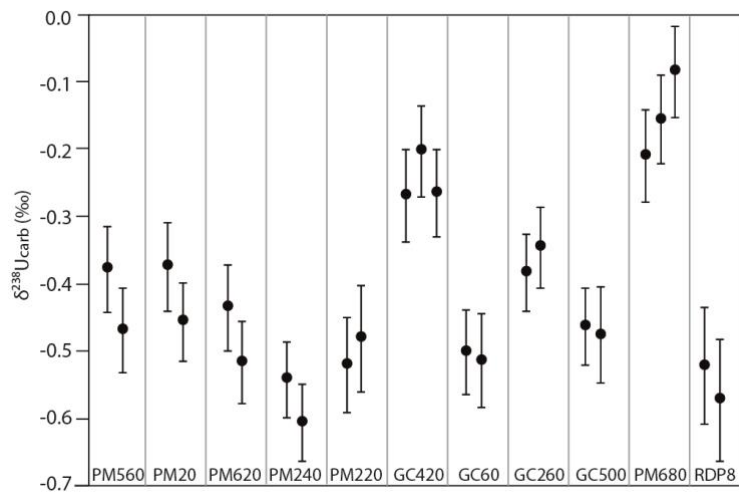
938 Diagenesis is an important consideration for understanding any geochemical dataset,
939 especially in carbonates that can undergo extensive syn- and post-depositional recrystallization. The
940 redox-sensitive nature of U also introduces additional complications in the diagenetic environment. U
941 mobility in sediments is heavily influenced by local reduction and diffusion within pore waters (18, 31,
942 80) and secondary U can be incorporated into diagenetic cements or into neomorphosed carbonates
943 (31, 81). This process has been demonstrated for individual carbonate cements (81) and for bulk
944 samples from modern carbonate cores which tend to show a positive $\delta^{238}\text{U}$ offset compared to
945 contemporary seawater (31). Negative $\delta^{238}\text{U}$ offsets have also been measured in some dolomitized
946 carbonates (21, 31). The reason for this is poorly understood but it may be due to either the influence
947 of meteoric waters with more negative $\delta^{238}\text{U}$ signatures, or alteration under hypersaline conditions
948 (21, 31).

949 The influence of pore-water reduction and syn-sedimentary recrystallization or cementation
950 should be lowest in the Eastbourne Chalk section, which is thought to have had an oxygenated water
951 column during deposition, with very low total organic carbon (TOC; less than 0.15% (82)) and no
952 appreciable sulphate reduction in pore waters (33). The impact of later stage diagenesis should be
953 minimal at Eastbourne, with the stable low-magnesium calcite mineralogy characteristic of pelagic
954 calcifiers that is more resistant to diagenetic alteration (83). This supposition is confirmed by the lack
955 of correlation between $\delta^{238}\text{U}$ with [U] or Mn/Ca and Sr/Ca (Fig. S3). Indeed, Eastbourne has been
956 widely used as a reference section for many isotopic systems because of the exemplary nature of the
957 archive. It is encouraging that major changes in $\delta^{238}\text{U}$ are not dependent on lithology, and also cross
958 major lithology transitions. In general the $\delta^{238}\text{U}_{\text{carb}}$ data during the OAE 2 interval show systematic
959 trends that correspond to hypothesised changes in other environmental parameters. Preceding the
960 CIE of OAE 2, however, is a negative $\delta^{238}\text{U}_{\text{carb}}$ excursion that corresponds to a U/Ca decrease. We do
961 not focus on this excursion, however it is plausibly related to observed cyclic black shales that precede
962 OAE 2 and thought to represent anoxia driven by orbital variability (68).

963 Carbonate samples from South Ferriby also show little geochemical evidence for
964 recrystallization (Fig. S3) although the section is generally more lithified than Eastbourne. The
965 presence of the distinct black shale (the so-called 'Black Band') above the unconformity raises some
966 concerns. Potentially, U mobilization from the organic-rich level could explain the positive $\delta^{238}\text{U}$
967 excursion seen in South Ferriby. The highest $\delta^{238}\text{U}_{\text{carb}}$ in this section also correspond to the highest
968 Mn/Ca ratios, suggesting the influence of pore water U reduction.

969 In contrast to the pelagic sites, Raia del Pedale has undergone extensive recrystallization, with
970 evidence for interaction with meteoric waters coming from depleted Sr/Ca and, in places, high Mg/Ca
971 indicative of dolomitization. In particular, the three lowest $\delta^{238}\text{U}$ values in Raia del Pedale are from
972 samples with higher Mg/Ca and anomalously high U/Ca ratios (Fig. 1, Fig. S3), consistent with an
973 observed negative isotopic shifts between modern dolomites and seawater (21, 31). These low values
974 however, form a negative excursion that appears to correlate with the Eastbourne $\delta^{238}\text{U}_{\text{carb}}$ data
975 during the more variable pre-OAE 2 interval (Fig. 1). Further investigation of pre-OAE 2 U-isotopes are
976 required to test if this excursion is primary. The anomalously high $\delta^{238}\text{U}_{\text{carb}}$ values within the OAE 2
977 interval at Raia del Pedale are more typical of signatures found in anoxic shales and may potentially
978 indicate extensive early diagenetic alteration and secondary incorporation of U from *in situ* organic-
979 matter degradation. Interestingly, Mn is not enriched in these samples, as would be expected due to
980 alteration under reducing pore waters. These high $\delta^{238}\text{U}_{\text{carb}}$ values are, however, restricted to a few
981 samples only, suggesting that these potential geochemical impacts were probably spatially limited.
982

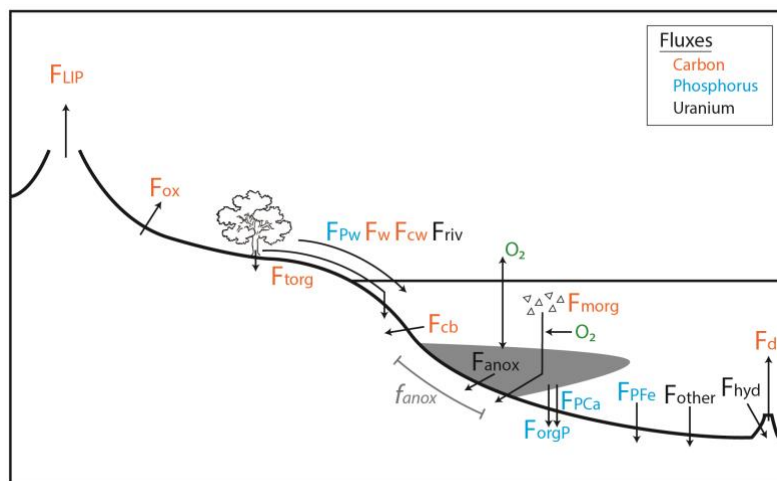
983 **Supplementary Figures**



984

985 Fig. S1: Replicate measurements performed on multiple samples between analysis sessions.

986 Uncertainty shown as 2 SE.



987

988 Fig. S2. Schematic illustrating coupled model fluxes of the C–P–U cycles with reference to the
989 variables defined in the text.

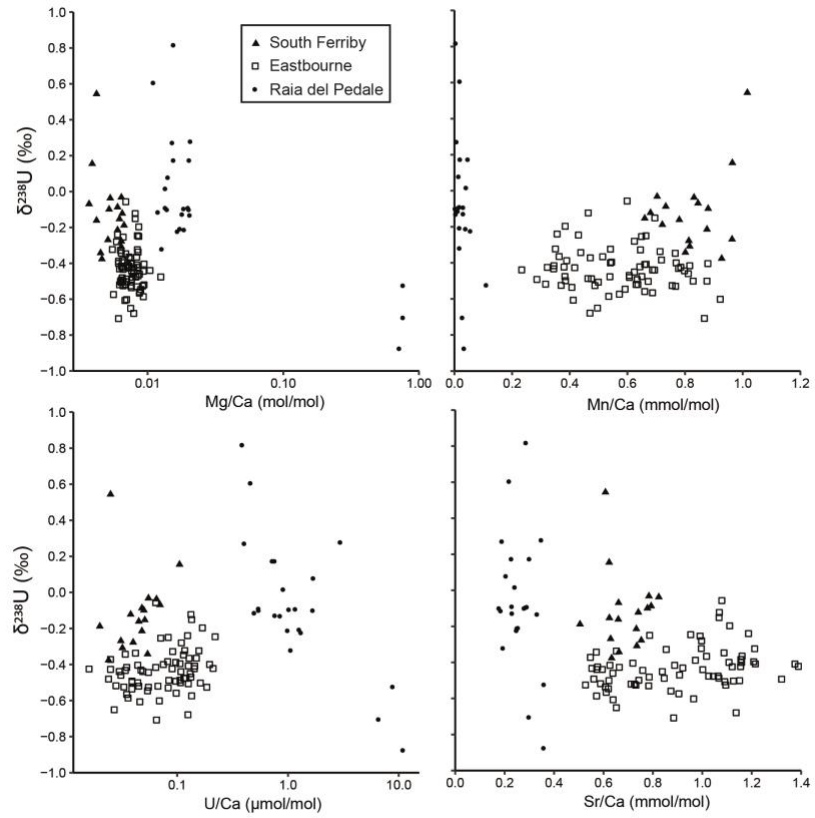
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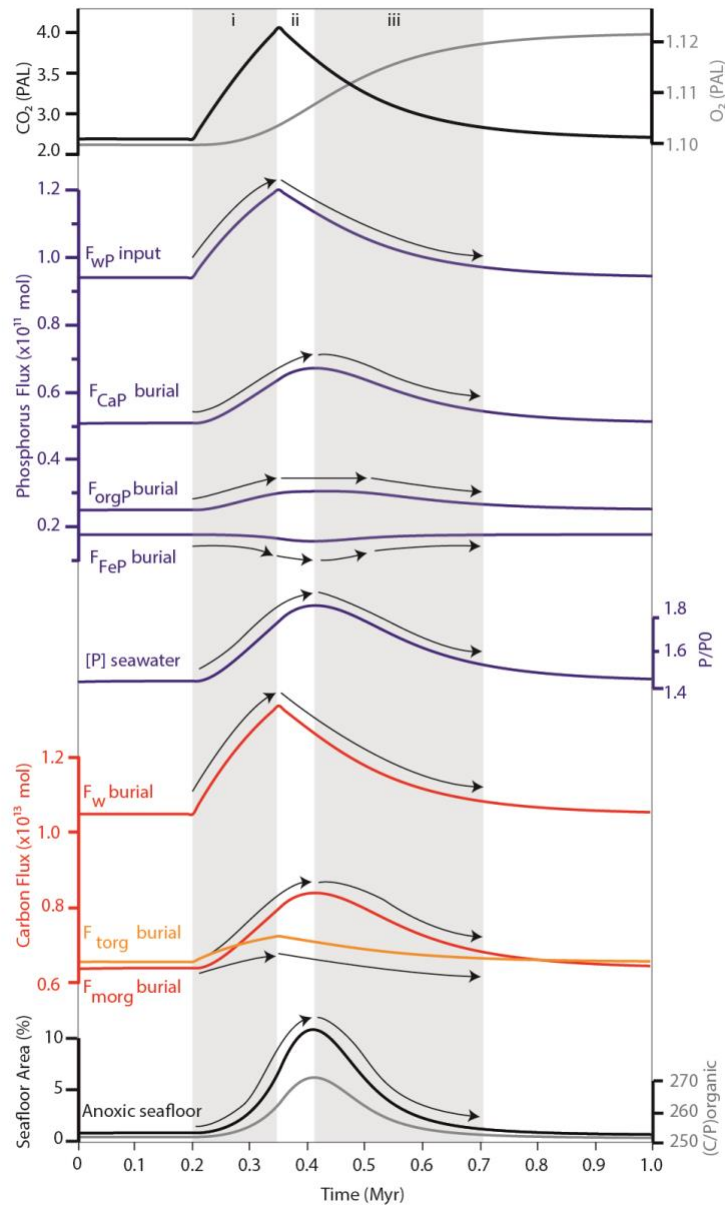
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996 Fig. S3: Cross plots of $\delta^{238}\text{U}$ against geochemical indicators of diagenesis.

997



998

999 Fig. S4: C–P–U model outputs for a single expansion of anoxia under the $O_2 = 1.1$ scenario with a CO_2
 1000 emission of 1.05×10^{13} mol C yr⁻¹ for 150 kyrs. See model description and main text for element flux
 1001 symbols. The behaviour of the C and P cycles can be divided into three stages related to the
 1002 development of anoxia: i) the onset of deoxygenation; ii) peak extent of anoxia; and iii)
 1003 reoxygenation. C and P fluxes are separated and the arrows highlight the behaviour of each flux
 1004 across the three stages. Specifically, a decoupling of [P] and P inputs (F_{wP}) in stage (ii) is related to
 1005 changes in the C/P ratio and burial of organic bound P (F_{orgP}) and Fe bound P (F_{FeP}). The minor
 1006 increase in atmospheric O_2 , due to the burial of organic carbon, is also shown. These internal P
 1007 dynamics create the delay in the maximum extent of anoxia compared to the atmospheric CO_2
 1008 increase and also facilitate reoxygenation as CO_2 decreases.

1009

Table S1. Sensitivity analysis varying the model boundary conditions (with a common 1.8×10^{18} molC perturbation over 150 kyr with $\delta_{LIP} = 0$ ‰). Boxed rows highlight boundary conditions used in the main text.

Initial conditions				Initial steady-state outputs						Maximum excursion outputs					
D	E	W	O ₂ /O ₂₀	CO ₂	P/P ₀	U/U ₀	f _{anoxic}	δ ¹³ C	δ _U	CO ₂	P/P ₀	U/U ₀	f _{anoxic}	δ ¹³ C	δ _U
1	1	1	1	1	1	1	0.003	1.82	-0.39*	1.82	1.69	0.35	0.12	3.82	-0.87*
1.8	0.7	1.0	1.0	2.68	1.43	0.50	0.031	3.23	-0.53	4.26	2.45	0.03	0.92	5.54	-1.01
1.8	0.7	1.0	1.1	2.69	1.42	0.91	0.009	3.21	-0.39	4.28	1.95	0.20	0.17	4.35	-0.88
1.8	0.7	1.0	1.2	2.69	1.41	1.18	0.003	3.20	-0.30	4.29	1.89	0.78	0.04	4.22	-0.52
1.5	0.7	1.0	1.0	2.24	1.22	0.82	0.009	2.91	-0.39	3.53	1.68	0.27	0.12	4.19	-0.89
2.1	0.7	1.0	1.0	3.12	1.67	0.20	0.120	3.57	-0.63	4.79	2.86	0.03	0.99	6.00	-0.87
1.8	0.5	1.0	1.0	4.42	1.35	0.63	0.019	3.50	-0.47	6.63	1.76	0.15	0.18	4.34	-0.85
1.8	0.9	1.0	1.0	1.92	1.49	0.40	0.046	3.02	-0.56	3.14	2.74	0.03	0.99	5.77	-1.01
1.8	0.7	0.8	1.0	3.70	1.37	0.59	0.023	3.41	-0.49	5.68	1.90	0.08	0.33	4.53	-0.93
1.8	0.7	1.2	1.0	2.10	1.47	0.43	0.041	3.08	-0.55	3.41	2.67	0.03	0.98	5.73	-1.01

*Note: U-isotope system set up for the modern with $\delta_{riv} - 0.29$. All other scenarios assume $\delta_{riv} - 0.19$

1013

Table S2. Sensitivity analysis varying the model isotopic assumptions. Using standard boundary conditions of D=1.8, E=0.7, W=1.0, O₂=1.0, and a common 1.8x10¹⁸ molC perturbation over 150 kyr.

Input assumptions			$\delta^{13}\text{C}$			$\delta^{238}\text{U}_{\text{sw}}$			
$\delta^{13}\text{C}$ LIP	$\delta^{238}\text{U}_{\text{river}}$	Δ_{anoxic}	initial	min	max	initial	min	max	Final
-5			3.23	2.26	5.41				
0			3.23	2.96	5.54				
-10			3.23	1.44	5.29				
	-0.19	0.5				-0.53	-1.01	-0.31	-0.48
	-0.29	0.5				-0.63	-1.11	-0.41	-0.58
	-0.09	0.5				-0.43	-0.91	-0.21	-0.38
	-0.19	0.4				-0.46	-0.85	-0.29	-0.43
	-0.19	0.6				-0.59	-1.17	-0.33	-0.53

1014

1015

1016 Data Tables

Eastbourne U-isotope data and trace element ratios performed on 0.6 or 1 M Sodium Acetate leachates. Li-isotope data after ref 8 and this study. Heights from base of section.									
Sample Name	Height (cm)	$\delta^{238}\text{Ucarb}$ (‰)	2SE	Mg/Ca (mol/mol)	Sr/Ca (mmol/mol)	Mn/Ca (mmol/mol)	U/Ca ($\mu\text{mol/mol}$)	$\delta^7\text{Li}$ (‰)	2SD
GC600	0	-0.559	0.065	0.009	0.873	0.660	0.098		
GC580	20	-0.477	0.057	0.008	1.045	0.539	0.115		
GC560	40	-0.366	0.057	0.008	1.017	0.514	0.128		
GC540	60	-0.197	0.085	0.009	1.111	0.384	0.169		
GC500	100	-0.470	0.071	0.009	0.908	0.606	0.133	19.1	0.6
GC460	140	-0.123	0.057	0.008	1.069	0.464	0.133		
GC440	160	-0.451	0.066	0.008	0.838	0.624	0.131		
GC420	180	-0.246	0.068	0.008	0.953	0.430	0.220		
GC400	200	-0.488	0.068	0.008	0.846	0.475	0.084	20.3	0.2
GC380	220	-0.679	0.048	0.008	1.137	0.470	0.125		
GC360	240	-0.476	0.063	0.008	1.064	0.383	0.120		
GC300	300	-0.487	0.059	0.007	1.067	0.463	0.106	19.2	0.9
GC260	340	-0.365	0.057	0.007	1.161	0.363	0.149		
GC200	400	-0.423	0.068	0.007	1.155	0.380	0.143		
GC160	440	-0.388	0.065	0.006	0.972	0.389	0.097		
GC120	480	-0.323	0.065	0.006	0.996	0.350	0.117		
GC100	500	-0.240	0.087	0.006	1.187	0.356	0.124	23.9	0.1
GC60	540	-0.507	0.069	0.006	1.091	0.499	0.126		
GC40	560							20.6	0.4
GC00	600	-0.280	0.052	0.006	0.997	0.624	0.109	11.2	0.5
PM20	620	-0.416	0.059	0.006	0.888	0.816	0.064		
PM40	640	-0.328	0.059	0.006	0.854	0.645	0.091		
PM60	660	-0.526	0.066	0.009	1.095	0.755	0.185	11.9	0.0

PM100	700							7.4	0.7
PM120	720	-0.400	0.071	0.008	1.159	0.541	0.196		
PM140	740	-0.503	0.067	0.007	1.126	0.672	0.162		
PM180	780	-0.347	0.060	0.007	1.088	0.766	0.114		
PM200	800							8.7	0.1
PM220	820	-0.501	0.070	0.006	1.151	0.876	0.107		
PM240	840	-0.574	0.057	0.006	0.905	0.572	0.135		
PM260	860	-0.602	0.068	0.007	0.965	0.922	0.069		
PM280	880	-0.708	0.060	0.006	0.884	0.867	0.065		
PM300	900	-0.408	0.065	0.007	1.215	0.665	0.126	5.9	0.4
PM340	940	-0.400	0.071	0.006	1.108	0.769	0.075	10.7	0.3
PM360	960	-0.057	0.092	0.007	1.079	0.599	0.064		
PM400	1000	-0.431	0.058	0.006	0.921	0.777	0.047	9.9	0.6
PM420	1020	-0.324	0.080	0.008	1.212	0.541	0.157		
PM440	1040	-0.396	0.064	0.007	1.208	0.545	0.106		
PM480	1080	-0.254	0.060	0.007	0.991	0.649	0.071		
PM500	1100	-0.409	0.060	0.008	1.375	0.660	0.144	15.3	0.4
PM520	1120	-0.384	0.059	0.006	1.156	0.718	0.082		
PM540	1140	-0.493	0.061	0.008	1.321	0.634	0.097		
PM560	1160	-0.424	0.063	0.007	1.121	0.785	0.090	13.1	0.4
PM580	1180	-0.422	0.082	0.008	1.389	0.643	0.211		
PM600	1200							8.3	0.5
PM620	1220	-0.476	0.061	0.012	1.026	0.650	0.117	10.2	0.4
PM640	1240	-0.440	0.082	0.010	1.002	0.694	0.108	22.8	0.4
PM680	1280	-0.151	0.068	0.008	1.068	0.695	0.135		
PM720	1320	-0.404	0.068	0.009	0.719	0.879	0.044		
PM740	1340	-0.250	0.060	0.009	0.787	0.664	0.083		
PM760	1360	-0.444	0.067	0.009	0.759	0.798	0.055	22.5	0.1

WC0	1390	-0.520	0.101	0.009	0.725	0.630	0.067	20.9	0.2
PM800	1400	-0.530	0.075	0.008	0.783	0.767	0.083		
W0.2	1410							21.8	0.2
WC200	1440							17.7	0.7
W0.6	1450	-0.546	0.071	0.009	0.620	0.592	0.054		
W0.8	1470								
W1	1490	-0.460	0.069	0.009	0.638	0.811	0.031	15.7	0.5
W1.2	1510							13.1	0.2
W1.4	1530	-0.566	0.069	0.009	0.611	0.687	0.035		
WC300	1540							14.5	0.6
W1.6	1550	-0.439	0.071	0.008	0.595	0.687	0.030		
W2	1590	-0.337	0.065	0.008	0.652	0.713	0.042	17.8	0.2
W2.4	1630	-0.586	0.075	0.009	0.572	0.535	0.036		
W2.8	1670	-0.481	0.074	0.006	0.785	0.607	0.024		
W3	1690							22.5	0.3
W3.2	1710	-0.520	0.066	0.006	0.712	0.636	0.045		
W3.6	1750	-0.501	0.077	0.007	0.622	0.828	0.044		
W4	1790	-0.526	0.084	0.007	0.731	0.487	0.055	20.7	0.6
W4.4	1830	-0.499	0.073	0.007	0.743	0.385	0.039		
WC600	1840							23.8	0.6
W4.8	1870	-0.651	0.078	0.007	0.652	0.496	0.027		
W5	1890	-0.343	0.075	0.006	0.571	0.438	0.034	20.6	0.3
W5.4	1930	-0.607	0.066	0.007	0.639	0.412	0.046		
W5.6	1970								
W6	1990	-0.536	0.059	0.008	0.608	0.407	0.039	20.1	0.4
W6.4	2030	-0.416	0.066	0.007	0.626	0.345	0.035		
W7	2090	-0.427	0.069	0.007	0.664	0.417	0.025	28.7	0.3
W7.4	2130	-0.372	0.080	0.008	0.547	0.471	0.025		

W8.4	2230	-0.525	0.068	0.007	0.527	0.370	0.033		
W8.8	2270	-0.518	0.068	0.007	0.588	0.316	0.028		
W9.8	2370	-0.433	0.059	0.007	0.553	0.344	0.034		
W10.8	2470	-0.492	0.061	0.006	0.561	0.286	0.039		
W12	2590	-0.425	0.080	0.008	0.576	0.322	0.016		
W13	2690	-0.439	0.059	0.008	0.573	0.233	0.049		

1017

1018

Eastbourne redox sensitive trace metal ratios performed on 0.5 M Acetic acid leachates as in reported in ref. 35. Heights from base of section.						
Sample Name	Height (cm)	U/Ca ($\mu\text{mol/mol}$)	Cr/Ca ($\mu\text{mol/mol}$)	Ni/Ca ($\mu\text{mol/mol}$)	Cu/Ca ($\mu\text{mol/mol}$)	V/Ca ($\mu\text{mol/mol}$)
GC560	40	0.168	2.442	2.947	2.719	0.953
GC500	100	0.094	2.412	1.371	2.840	0.692
GC440	160	0.174	2.145	1.253	1.573	1.059
GC380	220	0.118	2.115	1.372	2.205	0.766
GC320	280	0.103	2.197	1.381	1.558	0.805
GC260	340	0.123	2.784	2.754	6.432	0.831
GC140	460	0.162	2.318	1.338	3.167	0.890
GC80	520	0.129	2.993	3.446	5.424	0.841
GC200	580	0.163	2.389	1.255	3.196	0.689
PM40	640	0.111	1.556	4.805	6.998	0.888
PM100	700	0.130	1.358	4.855	4.330	0.759
PM160	760	0.088	1.599	4.339	4.842	0.814
PM220	820	0.112	2.093	3.677	5.676	1.193
PM280	880	0.095	1.399	4.259	4.122	0.668
PM340	940	0.078	1.578	3.437	2.102	0.656
PM400	1000	0.070	1.249	1.993	3.738	0.511
PM440	1040	0.063	1.116	1.053	2.310	0.544
PM500	1100	0.092	1.747	1.272	1.858	0.533
PM540	1140	0.112	2.169	3.304	7.983	0.603
PM580	1180	0.118	7.099	23.542	3.919	1.464
PM620	1220	0.093	18.572	24.022	7.000	2.591
PM640	1240	0.052	7.212	2.347	5.434	0.988
PM680	1280	0.183	24.409	24.815	21.766	4.497

PM700	1300	0.083	5.164	1.994	5.227	1.205
PM740	1340	0.086	11.833	7.012	18.659	3.679
PM760	1360	0.058	6.489	2.758	10.789	2.650
PM800	1400	0.083	1.336	1.692	2.592	1.283
W0.4	1430	0.050	3.061	2.287	5.001	0.923
WC200	1440	0.043	0.985	0.754	3.885	0.469
W0.8	1470	0.057	1.055	0.869	5.591	0.764
W1	1490	0.054	0.917	1.225	5.312	0.489
W1.4	1530	0.037	0.891	0.864	3.621	0.266
W1.6	1550	0.036	0.984	1.145	4.553	0.337
W2.2	1610	0.024	0.859	0.713	2.871	0.279
W2.8	1670	0.026	0.833	0.722	3.966	0.305
W3.4	1730	0.030	0.944	0.982	4.669	0.376
W4.2	1810	0.035	1.134	4.042	4.008	0.438
W5.2	1910	0.031	0.883	0.551	3.502	0.387
W5.6	1970	0.058	0.907	0.564	4.186	0.602
W6.4	2030	0.030	0.856	1.914	7.638	0.419
W7.2	2110	0.028	1.013	0.767	3.519	0.449
W7.6	2150	0.021	1.285	0.762	3.872	0.372
W8.2	2210	0.033	0.639	0.566	2.811	0.417
W8.8	2270	0.028	0.765	0.575	2.477	0.318
W9.4	2330	0.030	0.797	0.369	3.255	0.383
W10.6	2450	0.033	0.698	0.618	3.270	0.519
W11.2	2510	0.025	0.616	0.793	2.544	0.366
W11.8	2570	0.030	1.734	0.936	4.600	0.423
W12.4	2630	0.049	5.762	0.953		1.162

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Geochemical data for South Ferriby as used in Fig. 1 and S2. Heights from base of section							
Sample Name	Height (cm)	$\delta^{238}\text{Ucarb}$ (‰)	2SE	Mg/Ca (mol/mol)	Sr/Ca (mmol/mol)	Mn/Ca (mmol/mol)	U/Ca ($\mu\text{mol/mol}$)
SF-100	0	-0.277	0.060	0.006	0.753	0.813	0.040
SF-90	10	-0.214	0.058	0.006	0.733	0.876	0.048
SF-75	25	-0.122	0.059	0.007	0.741	0.680	0.038
SF-65	35	-0.308	0.057	0.006	0.735	0.816	0.032
SF-55	45	-0.087	0.059	0.006	0.793	0.733	0.048
SF-40	60	-0.099	0.058	0.005	0.778	0.880	0.050
SF-25	75	-0.037	0.058	0.005	0.824	0.831	0.065
SF-15	85	-0.032	0.061	0.006	0.784	0.703	0.055
SF25	125	-0.069	0.060	0.004	0.661	0.845	0.070
SF30	130	0.155	0.066	0.004	0.623	0.964	0.105
SF45	145	0.544	0.058	0.004	0.608	1.016	0.025
SF55	155	-0.161	0.056	0.004	0.660	0.780	0.045
SF60	160	-0.376	0.045	0.005	0.633	0.927	0.024
SF75	175	-0.269	0.052	0.005	0.629	0.963	0.031
SF85	185	-0.342	0.047	0.004	0.663	0.801	0.054
SF95	195	-0.153	0.050	0.006	0.624	0.660	0.051
SF150	250	-0.188	0.064	0.007	0.505	0.721	0.020

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Geochemical data for Raia del Pedale as used in Fig. 1 and S2. Heights from base of section. U/Ca presented in Fig 1 are from ref. 35							
Sample Name	Height (cm)	$\delta^{238}\text{Ucarb}$ (‰)	2SE	Mg/Ca (mol/mol)	Sr/Ca (mmol/mol)	Mn/Ca (mmol/mol)	U/Ca ($\mu\text{mol/mol}$)
RDP0	0.00	0.280	0.061	0.021	0.350	0.000	2.992
RDP3	3.00	0.081	0.067	0.014	0.207	0.016	1.710
RDP8	8.00	-0.522	0.086	0.769	0.361	0.112	8.902
RDP9	9.00	-0.702	0.070	0.768	0.300	0.029	6.639
RDP16	16.00	-0.872	0.077	0.720	0.360	0.035	11.040
RDP16	19.00	-0.206	0.076	0.017	0.256	0.019	1.272
RDP22.5	22.50	-0.221	0.092	0.017	0.250	0.057	1.323
RDP32	32.00	-0.126	0.073	0.018	0.232	0.032	0.770
RDP33.25	33.25	0.175	0.076	0.016	0.229	0.021	0.729
RDP38	38.00	-0.098	0.070	0.014	0.179	0.004	0.548
RDP39.2	39.25	0.273	0.067	0.015	0.191	0.009	0.407
RDP42	42.00	0.607	0.062	0.011	0.220	0.020	0.463
RDP44	44.00	-0.090	0.069	0.020	0.292	0.032	1.176
RDP51.65	51.65	-0.098	0.070	0.020	0.281	0.011	1.688
RDP56	56.00	0.819	0.079	0.016	0.288	0.006	0.389
RDP59.25	59.25	0.175	0.080	0.020	0.302	0.048	0.764
RDP68.25	68.25	-0.093	0.060	0.019	0.288	0.015	1.028
RDP71.75	71.75	-0.112	0.109	0.012	0.186	0.013	0.501
RDP75.5	75.50	-0.130	0.065	0.021	0.333	0.006	0.856
RDP80.5	80.50	-0.089	0.074	0.014	0.231	0.019	0.547
RDP85	85.00	-0.318	0.085	0.013	0.195	0.019	1.067
RDP89	89.00	0.019	0.062	0.014	0.243	0.042	0.913
RDP97	97.00	-0.210	0.087	0.019	0.252	0.040	1.001