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1 Unravelling the sources of carbon emissions at the onset
2 of Oceanic Anoxic Event (OAE) 1a

3 Markus Adloff^{a,*}, Sarah E. Greene^b, Ian J. Parkinson^c, B. David A. Naafs^{d,c},
4 Will Preston^c, Andy Ridgwell^{e,a}, Dan J. Lunt^a, José Manuel Castro Jiménez^f,
5 Fanny M. Monteiro^a

6 ^a*School of Geographical Sciences, University of Bristol, University Road, BS81SS, UK*

7 ^b*School of Geography, Earth and Environmental Sciences, University of Birmingham,
8 Edgbaston, B152TT, UK*

9 ^c*School of Earth Sciences, University of Bristol, Queen's Road, BS81RJ, UK*

10 ^d*School of Chemistry, University of Bristol, Cantock's Close, BS81TS, UK*

11 ^e*Department of Earth Sciences, University of California Riverside, 900 University Ave,
12 Riverside, CA 92521, USA*

13 ^f*Departamento de Geología, CEACTierra, Universidad de Jaén, Campus Las Lagunillas
14 s/n, 23071 Jaén, Spain*

15 **Abstract**

16 The early Aptian Oceanic Anoxic Event (OAE) 1a represents a major pertur-
17 bation of the Earth's climate system and in particular the carbon cycle, as
18 evidenced by widespread preservation of organic matter in marine settings and
19 a characteristic negative carbon isotopic excursion (CIE) at its onset, followed
20 by a broad positive CIE. The contemporaneous emplacement of a large igneous
21 province (LIP) is invoked as a trigger for OAE 1a (and OAEs in general), but this
22 link and the ultimate source of the carbon perturbation at the onset of OAEs
23 is still debated. In this study, we simultaneously assimilate an atmospheric
24 $p\text{CO}_2$ reconstruction along with a $\delta^{13}\text{C}$ record from the Spanish Cau section in
25 an Earth system model to obtain a novel transient reconstruction of emission
26 rates and identify the primary carbon-emitting sources across the negative CIE
27 interval at the onset of OAE 1a. We reconstruct carbon emissions of 4,300 -

*Corresponding author

Email address: markus.adloff@bristol.ac.uk (Markus Adloff)

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28 29,200 Pg from a mixture of carbon sources. This estimate is a lower bound,
29 as contemporaneous organic carbon burial is not accounted for. Carbon was
30 first released at slow rates from a ^{13}C -depleted reservoir (e.g. thermo- and/or
31 biogenic methane from sill intrusions). Towards the end of the negative CIE
32 the rate of emissions increased and they became more ^{13}C -enriched, likely from
33 a dominantly volcanic source (e.g. LIPs). New osmium isotope ($^{187}\text{Os}/^{188}\text{Os}$)
34 measurements, from the same section as the $p\text{CO}_2$ reconstruction and $\delta^{13}\text{C}$
35 data, reveal a shift to less radiogenic values coinciding with the change towards
36 mantle-derived carbon emissions as indicated by our modelling results, lending
37 further support to our interpretation. These results highlight that geologically
38 triggered carbon emissions were likely driving the OAE onset.
39 *Keywords:* Aptian, data assimilation, carbon source, osmium isotopes,
40 volcanism

41 1. Introduction

42 The early Aptian Oceanic Anoxic Event (OAE) 1a (~ 121 Ma, Olierook et al.,
43 2019) represents one of the largest perturbations of the climate system during
44 the last 200 Myr (Jenkyns, 2010). Although OAEs, including OAE 1a, were first
45 identified in the 1970s (Schlanger & Jenkyns, 1976), it is still unclear exactly how
46 the oceanic system evolved into such an extreme state. Increasing nutrient sup-
47 ply to the photic zone and, to a lesser extent, climate warming may have played
48 a crucial role (Larson & Erba, 1999; Jenkyns, 2010; Monteiro et al., 2012), driv-
49 ing increased productivity and subsurface oxygen demand, and reduced oxygen

50 solubility, respectively. The resulting reduced oceanic oxygen concentrations
51 would have led to enhanced preservation and burial of organic matter (Jenkyns,
52 2010; Monteiro et al., 2012). However, the driver of this warming and increased
53 nutrient delivery is debated.

54 While all OAEs are associated with a positive carbon isotopic excursion
55 (CIE), interpreted as reflecting an interval of enhanced burial of organic matter,
56 the onset of OAE 1a is characterized by a distinct negative CIE prior to the
57 positive CIE (Menegatti et al., 1998). Because of the broad temporal coincidence
58 of OAE 1a with the emplacement of the Ontong-Java Plateau, the negative CIE
59 could have resulted from carbon input from (submarine) volcanism initiating
60 the OAE by warming the climate and releasing nutrients to the ocean (e.g.
61 Weissert, 1989). A phase of increased volcanism during the onset of OAE 1a
62 (and other OAEs) is further highlighted by the observed perturbations of the
63 marine osmium cycle (Tejada et al., 2009; Bottini et al., 2012). However, it is
64 not certain whether volcanic emissions could have caused the negative CIE, so
65 destabilization of methane hydrates has been proposed as an alternative, more
66 ^{13}C -depleted carbon source (Beerling et al., 2002; van Breugel et al., 2007;
67 Méhay et al., 2009; Malinverno et al., 2010).

68 Constraining the carbon source for OAE 1a is crucial as it has implications
69 for the mass and rate of carbon emissions and hence extent of the perturba-
70 tion estimated from the carbon isotope record. The size of a negative CIE
71 is a function of the mass of emitted carbon and its isotopic composition (e.g.
72 Kirtland-Turner & Ridgwell, 2016), a predominantly volcanic carbon source

73 (CO₂), relatively enriched in ¹³C compared to organic carbon sources, requires
74 massive carbon emissions to match the negative CIE, resulting in high atmo-
75 spheric CO₂ concentrations (e.g. 3000 ppm *p*CO₂ increase (Bauer et al., 2017)).
76 In contrast, the estimated emissions, and hence CO₂ concentrations, are much
77 lower if the carbon originated predominantly from more ¹³C-depleted sources
78 (e.g. 600 ppm for methane emissions (Beerling et al., 2002)).

79 The nature of the dominant carbon source is important to determine the
80 role of Earth system feedbacks in triggering OAE formation in the Aptian and
81 its sensitivity to carbon injections. For instance, predominant volcanic CO₂
82 emissions suggest OAE formation was primarily a response to external forcing
83 while a dominant biogenic methane source would indicate a major role for feed-
84 backs between surface carbon reservoirs and an increased sensitivity to carbon
85 emissions. This could have broader implications for our understanding of the
86 formation of other Mesozoic OAEs since negative CIEs are also known from
87 OAE 1b and the Toarcian OAE, and the presence of a suppressed negative CIE
88 before OAE 2 is discussed (Jenkyns, 2010).

89 Recently, a continuous *p*CO₂ proxy record across OAE 1a (Naafs et al.,
90 2016) together with a $\delta^{13}\text{C}_{carb}$ record from the same site provides an oppor-
91 tunity to identify the carbon source and quantify emissions rates across OAE
92 1a. The *p*CO₂ record, based on compound-specific and bulk $\delta^{13}\text{C}_{carb}$ from the
93 hemipelagic Tethyan Cau section, shows a gradual *p*CO₂ increase during the
94 negative CIE starting around 800-1000 ppm and reaching maximum values of
95 2000 ± 1200 ppm before the recovery of the negative CIE. This estimate is within

96 the range of previously reconstructed $p\text{CO}_2$ values during the Early Cretaceous
97 of 200-2000 ppm (see Wang et al., 2014, for a compilation) and is between $p\text{CO}_2$
98 values predicted for a pure methane and a purely volcanic source. At the same
99 time as the $p\text{CO}_2$ increase, organic proxies indicate a 2-5 °C increase in lo-
100 cal sea-surface temperature (SST) during the negative CIE (Mutterlose et al.,
101 2014; Naafs & Pancost, 2016). Global temperature changes of this magnitude
102 are consistent with a doubling to quadrupling of $p\text{CO}_2$, depending on assump-
103 tions about the initial $p\text{CO}_2$ and the climate sensitivity of the Early Cretaceous.
104 At the same time, biological turn-over and carbonate platform drowning, key
105 features of OAE 1a, have been interpreted as being the results of surface ocean
106 acidification due to increased $p\text{CO}_2$ (Wissler et al., 2003; Erba et al., 2010).
107 However, it is debated whether the relative timing, the amount, and the rate of
108 emitted carbon explain the observed changes (Gibbs et al., 2011; Naafs et al.,
109 2016).

110 In this study, we assimilate the observed negative CIE and reconstructed
111 $p\text{CO}_2$ changes from the Cau section in the Earth system model cGENIE (Ridg-
112 well et al., 2007) to constrain carbon emissions across the onset of OAE 1a and
113 to understand their implications for the dominant carbon sources. We also as-
114 sess the effect of the $p\text{CO}_2$ changes on Earth's climate, as well as quantifying
115 the implications of uncertainties in $p\text{CO}_2$ change estimates and duration of the
116 OAE onset for reconstructing the carbon cycle perturbation. Finally, we com-
117 pare our simulation results to a newly-generated osmium isotope record from
118 the same section, which allows us to reconsider the relationship between the

119 negative CIE and environmental perturbations at the onset of OAE 1a.

120 **2. Methods**

121 *2.1. The Cau section*

122 The studied succession belongs to the Almadich Formation, made of a cyclic
123 succession of hemipelagic marls and marlstones of late Barremian- Aptian p.p.
124 age (Aguado et al., 1999; Castro et al., 2008). The interval considered in this
125 study is located in the upper part of the lower member of the Almadich Forma-
126 tion, of Early Aptian age (Desayesites forbesi ammonite zone, B. blowi plank-
127 tonic foraminifera, see Naafs et al., 2016). The lithologies present are dominant
128 grey marls (beds 0.5 – 2.5 m thick) alternating with marly limestones (beds
129 0.3 – 0.6 m thick), with two levels of dark marls (ca. 0.4 m thick) containing
130 small pyrite nodules. The samples used for Os analyses were taken from the
131 marly limestone levels, that have a CaCO₃ content between 36 and 70% wt and
132 TOC contents ranging from <0.7 to 2.4% wt. We assume stratigraphic continu-
133 ity and a constant sedimentation rate as there is no sedimentological evidence
134 for interruption of sedimentation (hard/soft-ground or erosive surfaces). Also
135 biostratigraphy and nannofossils associations are consistent with a continuous
136 record, at the scale of the resolution available (Aguado et al., 1999; Naafs et al.,
137 2016).

138 *2.2. Model setup and experiment design*

139 cGENIE is the carbon cycle centric version of the Grid ENabled Integrated
140 Earth system model (GENIE), which couples a 3D ocean with a 2D zonally-

141 averaged atmosphere (Edwards & Marsh, 2005), accounting for the biogeochem-
142 ical cycles of carbon, phosphorus, sulphur, oxygen and stable carbon isotopes
143 (Ridgwell et al., 2007). cGENIE can be run with a vertically-integrated sediment
144 module which builds up deep sea sediments with varying CaCO_3 content and
145 carbon isotopic composition, resolving for CaCO_3 burial and dissolution (Ridg-
146 well & Hargreaves, 2007) and with a terrestrial weathering module (Colbourn
147 et al., 2013) hence closing the geological carbon cycle. Its design also allows for
148 inversion (data assimilation) experiments (e.g. Cui et al., 2011; Gutjahr et al.,
149 2017, see below).

150 Here we ran the model on a 36x36 grid with 16 vertical levels in the ocean,
151 and included modules for weathering and marine sediments by which alkalinity
152 is added to or removed from the ocean.

153 In our model set-up, carbon emissions affect temperature, weathering rates,
154 ocean solubility of gases, and ocean circulation. Wind fields are fixed and pre-
155 scribed as boundary conditions and hence do not change in response to changing
156 atmospheric CO_2 concentration and greenhouse warming.

157 We configured cGENIE for the early Aptian using paleogeography and bathymetry
158 (see Appendix A), wind fields, and planetary albedo from the Aptian modelling
159 setup of Donnadieu et al. (2006). Temperature effects on biogeochemical pro-
160 duction were included, and remineralization profiles adapted to a warmer ocean
161 (John et al., 2014). We spun up the model in a first step (20 kyr) to equilibrate
162 to an atmospheric CO_2 concentration of 840 ppm (as reconstructed for the start
163 of the negative CIE in Naafs et al. (2016)) and atmospheric $\delta^{13}\text{C}$ value of -6

164 ‰, which is at the upper limit of measured values of the modern atmospheric
165 composition (Craig, 1953). Carbonate production is set up as in Ridgwell &
166 Schmidt (2010) to represent a warm, ice-free world. We assume that organic
167 carbon burial did not play a major role for the marine carbon cycle during the
168 interval that we simulate here (see Discussion section), but in order to equili-
169 brate oceanic $\delta^{13}\text{C}$ in the absence of organic carbon burial, we set the carbon
170 isotopic composition of the weathering flux to 6.2 ‰. Initial calcium ion supply
171 from terrestrial weathering is set to 34 Pmol yr^{-1} , resulting in an average sur-
172 face calcite saturation state (4.3) within the reconstructed range for the Early
173 Cretaceous (Ridgwell & Zeebe, 2005). The initial oceanic nutrient inventory
174 was set to $1.06 \mu\text{mol PO}_4^{3-} \text{ kg}^{-1}$ ocean water, equalling half of the preindus-
175 trial oceanic phosphate inventory. This is based on a model-data comparison
176 of pre-OAE 1a anoxia extents, following Monteiro et al. 2012 (see appendix
177 G). However, sensitivity experiments showed that the initial phosphate inven-
178 tory has little effect on our results (see fig. D2 in the appendix). During the
179 second spin up stage (500 kyr), the model was run as an open system so that
180 volcanic outgassing, riverine input of solutes to the ocean, and carbonate burial
181 in marine sediments could reach equilibrium.

182 To constrain carbon fluxes and their isotopic composition across OAE 1a
183 onset, we used a double data assimilation approach. Similar studies with single
184 assimilations used the size of negative CIEs to constrain emission rates associ-
185 ated with other major carbon cycle events such as the Paleocene-Eocene Ther-
186 mal Maximum (PETM, e.g. Cui et al., 2011; Turner & Ridgwell, 2013) making

187 assumptions about the $\delta^{13}\text{C}$ signature of the carbon source. Here however, we
188 intend to reconstruct both the mass and provenance (i.e. isotopic composition)
189 of the emitted carbon. Knowing that a given isotopic excursion can be pro-
190 duced by adding a small mass of very ^{13}C -depleted carbon or a large mass of
191 less ^{13}C -depleted carbon (e.g. Kirtland-Turner & Ridgwell, 2016) this consti-
192 tutes a problem with two unknowns (mass and composition), the solution for
193 which requires an additional constraint to the CIE size. We hence combined
194 estimates of atmospheric $\delta^{13}\text{C}$ changes with estimates of $p\text{CO}_2$ changes (Naafs
195 et al., 2016) to reconstruct carbon emissions across the onset of the OAE 1a
196 negative CIE.

197 Each time step, cGENIE adds the mass of carbon needed to reach the pre-
198 scribed $p\text{CO}_2$ target to the atmosphere, and then calculates the required isotopic
199 signature of that carbon addition to create the right CIE. We allow a maximum
200 carbon emission rate of 10 Pg yr^{-1} with a $\delta^{13}\text{C}$ not more negative than -100
201 ‰. We used the Naafs et al. (2016) $p\text{CO}_2$ estimates based on $\Delta^{13}\text{C}$, the offset
202 between compound specific $\delta^{13}\text{C}$ in organic matter and bulk carbonate $\delta^{13}\text{C}$
203 ($\delta^{13}\text{C}_{carb}$) (Popp et al., 1998). We assumed that the evolution of $\delta^{13}\text{C}_{carb}$ re-
204 flects marine and atmospheric $\delta^{13}\text{C}$ changes (see Appendix C). For this, we
205 correct sedimentary $\delta^{13}\text{C}$ measurements from the Cau section by applying a 7.7
206 ‰ offset to account for different baseline $\delta^{13}\text{C}$ values between atmosphere and
207 biogenic CaCO_3 . We smooth these data by applying a running mean over 7
208 data points (see fig. 1) to minimize the effect of short-term fluctuations. This
209 smoothed sedimentary $\delta^{13}\text{C}$ is characterized by an averaged 1.3 ‰ negative

210 excursion (maximum value 1.6 ‰), which is similar to most other measured
211 negative CIE amplitudes for OAE 1a (see Appendix B). This dual data assim-
212 ilation exercise is analogous to Gutjahr et al. (2017)’ assessment of the PETM
213 carbon cycle perturbation, except here taking pelagic $\delta^{13}\text{C}_{carb}$ as a measure of
214 atmospheric rather than surface ocean (DIC) $\delta^{13}\text{C}$ changes, and we directly
215 constrain the rate of emissions using a $p\text{CO}_2$ reconstruction rather than surface
216 pH.

217 We additionally address a number of uncertainties in the data and its in-
218 terpretation. Firstly, implications of uncertainties in the $p\text{CO}_2$ reconstruction
219 are explored by using an ensemble of 5 simulations with different $p\text{CO}_2$ tra-
220 jectories within the uncertainty envelope reconstructed by Naafs et al. (2016)
221 (fig. 1). This uncertainty envelope accounts for uncertainties in the calibration
222 of the $p\text{CO}_2$ reconstruction method to Cretaceous organisms and due to the
223 fluctuability of the ^{13}C curve in Cau outcrop data. Secondly, the duration of
224 the negative CIE at the OAE 1a onset is poorly constrained. We run every
225 $p\text{CO}_2$ scenario with a duration of 100 kyr, but since durations between 20 and
226 300 kyr have been suggested for the onset of the negative CIE interval (e.g. van
227 Breugel et al., 2007; Li et al., 2008; Malinverno et al., 2010; Hu et al., 2012),
228 we repeat some simulations with three different assumed durations: 20 kyrs, 40
229 kyrs and 300 kyrs. In total, we run 20 inverse modelling simulations, exploring 5
230 different $p\text{CO}_2$ trajectories and 4 different assumptions about durations. Addi-
231 tional sensitivity experiments were run to investigate the sensitivity to boundary
232 conditions (see Appendix).

233 *2.3. Osmium isotope analysis*

234 Osmium (Os) isotopes and Os and Rhenium (Re) concentrations were de-
235 termined at the School of Earth Sciences, University of Bristol. Samples were
236 digested in Carius tubes using techniques slightly modified from Shirey & Walker
237 (1995). 0.3-0.5 g of powder was added to the Carius tubes. As the rock con-
238 tained a significant amount of carbonate (46 m - 62 m of the Cau section consist
239 of a succession of marl and marly limestone, see Quijano et al. (2012) and Naafs
240 et al. (2016) for more information on the lithologies) the powder was decarbon-
241 ated with a small amount of concentrated HCl prior to sealing. Appropriate
242 amounts of ^{190}Os and ^{185}Re spikes were added to the Carius tubes along with
243 additional HCl and HNO_3 to make up 12 ml of inverse aqua regia. The samples
244 were then digested at 230 °C for 48 hours. This technique recovers hydrogenic
245 Os and Re without dissolving any detrital material.

246 Os and Re separation chemistry and mass spectrometry follows methods re-
247 cently described in detail elsewhere (Josso et al., 2019). For Os isotope analyses
248 the purified Os was analysed on a ThermoFisher Triton thermal ionisation mass
249 spectrometer (TIMS) in N-TIMS mode, with instrumental mass fractionation
250 corrected for using the exponential mass fractionation law and a $^{192}\text{Os}/^{188}\text{Os}$
251 ratio of 3.08271. Os concentrations were determined by isotope dilution and the
252 isotope data were spike-stripped to yield the sample $^{187}\text{Os}/^{188}\text{Os}$ ratio. Total
253 procedural blanks were determined for each dissolution batch and yielded Os
254 concentrations of 0.31 and 1.46 pg, with $^{187}\text{Os}/^{188}\text{Os}$ ratios of 0.1783 ± 0.0347
255 and 0.1745 ± 0.0557 respectively. All data were corrected for the procedural

256 blank with blank corrections usually less than 1.5 % (range 0.04 to 4 %) on the
257 concentration and usually less than 1 % (range 0.02-3.5 %) on the $^{187}\text{Os}/^{188}\text{Os}$
258 ratio depending on sample size and Os concentration. A DTM solution standard
259 was run during the analytical period and yielded a $^{187}\text{Os}/^{188}\text{Os}$ ratio of 0.17396
260 ± 39 (2sd, n=12), within error of previous determinations of the standard (e.g.
261 Birck et al., 1997). Precision on the standard and samples is similar because
262 they were both run at comparable beam intensity, with the exception of one
263 sample, which has a precision of 12 ‰.

264 Re isotopes were measured by multi-collector inductively plasma source mass
265 spectrometry (MC-ICP-MS) on a ThermoFisher Neptune using a SEM and
266 corrected for instrumental mass fractionation using the exponential mass frac-
267 tionation law and a $^{191}\text{Ir}/^{193}\text{Ir}$ ratio of 0.59418. Total procedural blanks were
268 determined for each dissolution batch and yielded Re concentrations of 5 ± 2
269 pg, with Re concentrations determined by isotope dilution and blank corrected
270 using the appropriate procedural blank. Blank corrections for the Re concen-
271 trations are less than 2 % (range 0.14-4.1 ‰). A 9 ppt solution of the NIST
272 SRM3134 Re standard, doped with Ir, was run during the analytical period and
273 yielded a $^{187}\text{Re}/^{185}\text{Re}$ ratio within error of previous determinations of the stan-
274 dard (Miller et al., 2009). One sample (Cau-21) was fully duplicated, because
275 it had elevated Re concentrations, low Os concentrations and gave a low initial
276 $^{187}\text{Os}/^{188}\text{Os}$ ratio at a key part of the section. The duplicate analyses gave
277 initial $^{187}\text{Os}/^{188}\text{Os}$ ratios just outside of analytical uncertainty, but confirming
278 that there is significant drop in initial $^{187}\text{Os}/^{188}\text{Os}$ ratio in the section at this

279 time.

280 **3. Results**

281 The model simulations provide estimates of carbon emission rates and the
282 isotopic composition of the emitted carbon for the onset of the Aptian nega-
283 tive CIE under different $p\text{CO}_2$ change scenarios (fig. 2a-b) and onset durations
284 (fig. 2c-d). In our simulations, total emissions ranged between 4,300-29,200
285 Pg C with emission rates between 0.01 and 0.7 Pg C yr^{-1} . The uncertainty
286 in our carbon emission estimates is almost equally attributable to uncertainties
287 in the duration of the event and the $p\text{CO}_2$ reconstruction (10,300 Pg C and
288 13,700 Pg C respectively). Alternative assumptions on initial atmospheric CO_2
289 concentrations, oceanic nutrient inventory and surface ocean saturation state
290 add only minor additional uncertainty to emission rates (see sensitivity stud-
291 ies in Appendix D). The model results imply isotopically light carbon input
292 (mean composition across all scenarios = -42 ‰) at the start of the negative
293 CIE (46 - 50 m) for all scenarios. Toward the peak of the negative CIE, our
294 model results indicate a shift towards the input of less ^{13}C -depleted carbon
295 with a mean composition of -6 ‰ between 55.5 - 60.0 m. Additional sensitiv-
296 ity studies (see Appendix D) demonstrate that the necessity for ^{13}C -depleted
297 carbon at the start of the negative CIE is a robust result considering uncertain-
298 ties in initial atmospheric CO_2 concentration, initial oceanic nutrient inventory,
299 initial carbonate saturation state, and different biological production schemes
300 (temperature-dependent or -independent production).

301 In our simulations the calcite saturation state (Ω) of the surface ocean only
302 decreases appreciably in experiments with carbon emission rates of at least
303 0.1 Pg C yr^{-1} (see fig. F1 in Appendix F). In scenarios with slower carbon
304 emissions, surface ocean Ω stays constant or increases slightly. The total Ω
305 change and the associated uncertainty intervals vary between locations and are
306 largest in the tropics (up to ~ -2.5 in the most extreme experiments, see fig.
307 F2 in Appendix F). Sea surface temperature (SST) changes are more variable
308 with $p\text{CO}_2$ change scenario than assumed CIE onset duration, and are largest
309 around Antarctica (see fig. F3 in Appendix F).

310 The new Os data from Cau (see fig. 3) show a low concentration of Os
311 with a slight trend towards more radiogenic values (> 1) towards the start of the
312 negative CIE (40-45 m). This is followed by increasing Os concentration during
313 the CIE, its isotopic composition shifting to less radiogenic values (< 0.4). A
314 brief excursion back to low concentrations of very radiogenic Os (> 1.2) half
315 way through the negative CIE onset (~ 55 m) is based on two data points. All
316 data can be found in Appendix E.

317 4. Discussion

318 Our inverse modelling suggests that the addition of ^{13}C -depleted carbon dur-
319 ing the start of the negative CIE (Phase A in fig. 4), followed by sustained input
320 of carbon with a largely volcanic isotopic signature (Phase B, typically between
321 -5 and -8 ‰ away from subduction zones (Javoy et al., 1986)), reconciles the
322 measured $\delta^{13}\text{C}_{carb}$ and reconstructed $p\text{CO}_2$ at Cau (Naafs et al., 2016). Our

323 method results in required ‘net’ carbon fluxes and their isotopic compositions.
324 While we cannot exclude mixes of sources (and sinks) that sum to the right
325 ‘net’ values, this provides information about the dominant carbon source. The
326 requirement for initial carbon emissions more ^{13}C -depleted than volcanic CO_2
327 across our simulations indicates that contribution from organic carbon sources
328 (typically between -10 and -30 ‰ (e.g. Meyers, 1994)) and methane, the only
329 surficial carbon with a $\delta^{13}\text{C} < -40$ ‰, was necessary to drive the CIE at the
330 OAE 1a onset (see fig. 2b,d and Appendix). Predominant methane emissions,
331 however, are only required in scenarios with minimal $p\text{CO}_2$ rise during the CIE
332 onset. All scenarios require the most ^{13}C -depleted carbon input between 48 m
333 and 52 m. The isotopic composition of the carbon input predicted by cGENIE
334 between 52 m and 58 is consistently less depleted (around or above -10 ‰ for
335 most simulations) almost regardless of uncertainty in reconstructed $p\text{CO}_2$ or
336 duration. This is consistent only with an appreciable contribution from a vol-
337 canic source. The shift in source from sedimentary organic carbon to volcanism
338 occurs relatively rapidly, within 4 m at Cau or 25 kyr assuming a negative CIE
339 onset duration of 100 kyr and a constant sedimentation rate. Importantly, this
340 difference in isotopic composition between strata below and above 52 m is a
341 robust feature in all simulations and suggests that Earth’s system transitioned
342 from predominant input from an isotopically light carbon source to a heavier
343 one. Scenarios with low $p\text{CO}_2$ increase require another pulse of biogenic CO_2
344 to reach the lowest $\delta^{13}\text{C}$ values at the end of our simulations, while the isotopic
345 signature of carbon inputs in scenarios with higher emission rates remains close

346 to mantle values.

347 A shift to a more mantle-like source is supported by published Os isotope
348 data which have been used to infer changes in volcanic activity across OAEs,
349 given that the main source of unradiogenic Os is mantle material (Turgeon &
350 Creaser, 2008; Bottini et al., 2012; Du Vivier et al., 2014). Published OAE 1a Os
351 isotope records from pelagic sections show a shift towards unradiogenic (mantle-
352 like) values during the negative CIE at the onset of OAE 1a (Bottini et al., 2012),
353 suggesting a strengthened Os flux from the mantle. While a direct comparison
354 between Os records from Cau and other sites is hampered by uncertainties in
355 age models and missing stratigraphic ties, records from all sites show similar
356 features. The consistent shift to very unradiogenic Os-isotope values in the Cau
357 record (fig. 4) is similar to that recorded in other sections and adds confidence
358 to the assumption that strengthened volcanism was the primary control on Os
359 isotope changes recorded across the negative CIE at Cau, even more so given the
360 small amount of preserved organic matter preserved and its platform setting.
361 Similar to other Tethyan sections, the Os record at Cau shows a shift back to
362 radiogenic values before stagnating at unradiogenic values (in phase B in fig. 4,
363 Jenkyns (2018)). Since this excursion is not evident in records from all OAE
364 1a sections, it might represent a local or regional change with little implications
365 for the reconstructed volcanic strength. In that case the shifts from less to
366 more mantle-like compositions of Os and carbon coincide at Cau (fig. 4). If
367 the radiogenic excursion at ~ 55 m in the Cau section reflects a global change
368 of Os fluxes, the resolution of the $p\text{CO}_2$ reconstruction is either too coarse to

369 evidence a short-term decline in volcanic carbon emissions, or the relationship
370 between carbon and Os isotopic changes during the onset of LIP emplacement
371 is more complex. At this moment, we cannot test this relationship at other
372 sites because combined records of Os isotope and inferred $p\text{CO}_2$ changes are
373 not available at the required resolution.

374 Emissions of carbon from an organic matter reservoir are needed to explain
375 carbon isotopic changes at Cau regardless of the $p\text{CO}_2$ scenario or assumed du-
376 ration of CIE onset. However the exact mechanism of its release at the onset of
377 OAE 1a is not clear. It is unlikely that depleted carbon was emitted due to feed-
378 back mechanisms in response to a warming climate (e.g. destabilized methane
379 hydrates), given the small changes in atmospheric $p\text{CO}_2$ at that time. There is,
380 however, evidence in support of geologically-triggered emissions of ^{13}C -depleted
381 organic carbon during the onset of OAE 1a. Polteau et al. (2016) showed that
382 magma intrusions into overlying sediments during the formation of the Barents
383 Sea Sill Complex in the early Aptian could have destabilized a considerable mass
384 of organic carbon and led to thermogenic methane emissions. OAE 1a happened
385 at a time of high magmatic activity, including the emplacement phases of the
386 Ontong-Java Plateau LIP (e.g. Larson & Erba, 1999). We therefore suggest
387 that at the start of the negative CIE magmatic activity increased under thin
388 crust, intruding into organic-rich marine sediments and causing small releases of
389 ^{13}C -depleted thermogenic gases. This initial phase of increased mantle activity
390 was then followed by a more eruptive phase of LIP emplacement, when large-
391 scale submarine volcanism dominated exogenic carbon emissions and delivered

392 large amounts of new mantle material to the ocean floor. This increased the
393 flux of unradiogenic Os into the ocean through hydrothermal vents as well as
394 weathering of fresh basalt and released vast amounts of mantle-sourced carbon
395 to the atmosphere. Thus, the initial $\delta^{13}\text{C}$ decrease and the excursions in Os iso-
396 topes and $\Delta^{13}\text{C}$ are consistent with a scenario in which the recorded CIE and
397 $p\text{CO}_2$ change occurred independently: The first through release of sedimentary
398 organic carbon, and the latter through volcanic CO_2 emissions.

399 Our results based on records of atmospheric $p\text{CO}_2$ and $\delta^{13}\text{C}$ are consistent
400 with published model interpretations of Os isotope profiles, which also find
401 that volcanism was the primary carbon source during OAE 1a (Bauer et al.,
402 2017). The mass of emitted carbon in our results is however different from that
403 reconstructed by Bauer et al. (2017) using Os isotopes. The biomarker-based
404 estimate suggests an increase in $p\text{CO}_2$ of 200-1700 ppm (Naafs et al., 2016),
405 whereas the model interpretation of $^{187}\text{Os}/^{188}\text{Os}$ implies that $p\text{CO}_2$ rose by
406 about 3,000 ppm (Bauer et al., 2017). We tested the implications for carbon
407 sources assuming a continuous CO_2 concentration increase of 3000 ppm, which
408 resulted in increased carbon emissions ($0.32 \text{ Pg C yr}^{-1}$ assuming a 100 kyr
409 duration, see fig. D4 in the appendix) compared to our scenarios, with isotopic
410 compositions between -20 and -3 ‰. This is still consistent with a predominantly
411 volcanic source, but reduced the need for more ^{13}C -depleted organic carbon
412 release. Temperature proxy records across OAE 1a (Mutterlose et al., 2014;
413 Naafs & Pancost, 2016) are consistent with regional temperature changes of 1-5
414 °C produced by scenarios based on the Naafs et al. (2016) $p\text{CO}_2$ reconstruction.

415 The higher (Os based) $p\text{CO}_2$ changes result in a temperature rise of 7°C globally
416 in cGENIE, equivalent to $5 - 6^\circ\text{C}$ in the Tethys which is at the upper end of
417 reconstructed temperature change.

418 Organic carbon burial is not simulated by our model, meaning that we have
419 to assume that organic carbon burial did not strengthen significantly during the
420 negative CIE onset. Although carbon contents at Cau and at other locations
421 only increase during the recovery of the negative CIE and start of the positive
422 CIE (generally referred to as segments C4-C6 Menegatti et al., 1998; Quijano
423 et al., 2012; Erba et al., 2015), it is unknown to what degree the marine carbon
424 and Os cycles were influenced by organic matter burial locally and/or globally.
425 Widespread organic carbon burial would shorten the residence time of Os and
426 carbon in the ocean and create a sink for ^{12}C , increasing carbon emission rates
427 needed to produce the observed environmental changes. The results from our
428 experiments should thus be regarded as a low-end estimate of the mass of emit-
429 ted carbon. However, our experiments indicate that the transition from more
430 ^{13}C depleted organic carbon input (potentially from sill intrusions) to mantle-
431 sourced carbon (volcanism) is robust against changes in emission rate and total
432 mass of emitted carbon and hence should be detectable despite changes in the
433 amount of organic matter burial.

434 In terms of timing, high-resolution temperature records (based on palynology
435 and $\delta^{18}\text{O}$) at the expanded pelagic section of Pusiano and the shelf section
436 at La Bédoule (Keller et al., 2011; Lorenzen et al., 2013) indicate a time lag
437 between the start of the negative CIE and the main period of warming. This

438 would be consistent with our reconstructed emissions scenarios. At pelagic
439 sections lower sediment accumulation rates and coarser sampling resolutions
440 (e.g. Cismon, Roter Sattel, Pacific DSDP site 463 and Atlantic DSDP site 398),
441 likely prevented the identification of such time lag between temperature proxies
442 and negative CIE (Menegatti et al., 1998; Ando et al., 2008; Naafs & Pancost,
443 2016).

444 Major changes in the marine nannofossil community, including dwarfism,
445 and a decline in CaCO_3 burial rates, have been observed during the negative
446 CIE and interpreted as an indicator for a decline in calcite surface ocean satu-
447 ration state due to large carbon emissions (Weissert & Erba, 2004; Erba et al.,
448 2010). The widespread occurrence of surface ocean acidification during OAE
449 1a as driver of biological turn-over is contested (Gibbs et al., 2011; Naafs et al.,
450 2016). Our simulations show that only scenarios with a large carbon source or
451 short event duration result in an appreciable Ω decline ($\Delta\Omega > 0.5$), globally
452 and locally. A decrease in saturation state could have imposed significant stress
453 onto calcifiers (Weissert & Erba, 2004). At Cau, the $p\text{CO}_2$ changes and the
454 decline in Ω (around 54 m) began well after the onset of the nannofossil decline,
455 which starts at 43 m. In our longer simulations (≥ 100 kyr) increased weath-
456 ering and continental runoff eventually counteract the CO_2 dissolution effect,
457 which leads to a recovered or even increased calcite saturation state at the end
458 of these experiments, despite ongoing carbon emissions. Further quantitative
459 comparison of the impact of carbon emissions onto the surface ocean carbonate
460 chemistry in our simulations is hampered by remaining uncertainties. For ex-

461 ample, the carbonate system of the pre-OAE 1a Aptian surface ocean is poorly
462 constrained. Particularly the mass of emitted carbon needed to cause apprecia-
463 ble Ω changes and the position of the carbonate compensation depth depend
464 largely on these initial conditions. The slow carbon emission rates predicted by
465 our model are also a function of the duration of the negative CIE onset. Unless
466 the onset took 20 kyr (Malinverno et al., 2010, the smallest current estimate)
467 or less, it is unlikely that carbon emissions could have outpaced the weather-
468 ing feedbacks that stabilize the surface ocean saturation state (Hönisch et al.,
469 2012). As such, estimated onset durations and the temporal decoupling between
470 changes in calcite saturation state and nannoconid crisis make it unlikely that
471 this major perturbation of the marine nanno-calcifier community was a direct
472 result of carbon emissions driving the negative CIE.

473 5. Conclusion

474 We used transient model simulations based on combined $p\text{CO}_2$ and $\delta^{13}\text{C}$
475 records to identify dominant carbon sources and quantify associated emission
476 rates. The simulations show that carbon emission rates of 0.043-1.46 Pg C
477 yr^{-1} reconcile the negative CIE and $p\text{CO}_2$ estimates, with the exact rate de-
478 pending on the assumed duration of OAE onset. Assuming a baseline $p\text{CO}_2$
479 of 850 ppm this is consistent with estimates of moderate temperature change
480 from diverse proxy data. Our results suggest that an initial input of predomi-
481 nantly ^{13}C -depleted carbon was responsible for the negative CIE associated with
482 OAE 1a, but had little effect on atmospheric $p\text{CO}_2$. This carbon could have

483 originated from sill intrusions into marine organic rich sediments, potentially
484 releasing thermogenic and/or biogenic methane. Following this initial phase,
485 volcanic (less ^{13}C -depleted) carbon emissions sustained the stagnant part of the
486 negative CIE and fuelled most of the recorded atmospheric $p\text{CO}_2$ increase. Our
487 results indicate that organic carbon emissions in form of methane and/or CO_2 ,
488 despite shaping the $\delta^{13}\text{C}$ curve, did not increase the atmospheric carbon reser-
489 voir appreciably. Importantly, this demonstrates that the scale and nature of
490 the Earth system perturbation at the onset of OAE 1a cannot be determined
491 based on the existence of a negative CIE or its size alone. This may also be
492 the case for other negative CIEs in the geological record. Our results further
493 imply the possibility of a time lag between the carbon isotopic excursion and
494 the major phase of carbon emissions, but highly resolved records from other
495 sections are necessary to corroborate this finding. Uncertainties in the amount
496 of atmospheric $p\text{CO}_2$ change and duration of the OAE onset create propor-
497 tionally large uncertainties on carbon emission rates and their initial isotopic
498 composition. Increased accuracy in duration and $p\text{CO}_2$ change estimates are
499 needed to narrow down our estimates of carbon flux magnitudes and origins.
500 However, the shift from more to less ^{13}C -depleted carbon emissions is robust
501 against these uncertainties, as are the pure volcanic-like carbon fluxes towards
502 the peak of the negative CIE. We conclude that mantle activity increased in the
503 early Aptian, heating organic-rich marine sediments and releasing thermogenic
504 methane before sustained mantle eruptions led to OAE formation.

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520 **Code Availability**

521 *muffin overview*

522 The code for the cGENIE.muffin model is hosted on GitHub. The current
523 version can be obtained by cloning:

524 <https://github.com/derpycode/cgenie.muffin>

525 A manual, detailing code installation, basic model configuration, plus an exten-
526 sive series of tutorials covering various aspects of muffin capability, experimental

527 design, and results output and processing, is provided. A PDF of the manual
528 can be downloaded here:

529 <http://www.seao2.info/cgenie/docs/muffin.pdf>

530 The latex source and most up-to-date built PDF file can be obtained by cloning:

531 <https://github.com/derpycode/muffindoc>

532 *Instructions summary*

533 The muffin manual contains instructions for obtaining, installing, and testing
534 the code, plus how to run experiments. Specifically:

535 **Section 1.1** – provides a basic over-view of the software environment required
536 for installing and running muffin.

537 **Section 1.2.2** – provides a basic over-view of cloning and testing the code.

538 **Section 15.4** – provides a detailed guide to cloning the code and configuring a
539 Ubuntu (18.04) software environment including netCDF library installa-
540 tion, plus running a basic test.

541 **Section 15.6** – provides a detailed guide to cloning the code and configuring a
542 macOS software environment including netCDF library installation, plus
543 running a basic test.

544 **Section 1.3** – provides a basic guide to running experiments (also see 1.6 and
545 1.7).

546 **Section 1.4** – provides a basic introduction to model output (much more detail
547 is given in Section 12).

548 *Model experiments*

549 Configuration files for the specific experiments presented in the paper can be
550 found in the directory:

551 `cgenie.muffin\genie-userconfigs\MS\adloffetal.2019`

552 Details of the different experiments, plus the command line needed to run each
553 one, are given in `readme.txt`.

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739 **Figures**

a) Reconstructed atmospheric carbon b) Model forcing for stage C3
 across OAE 1a

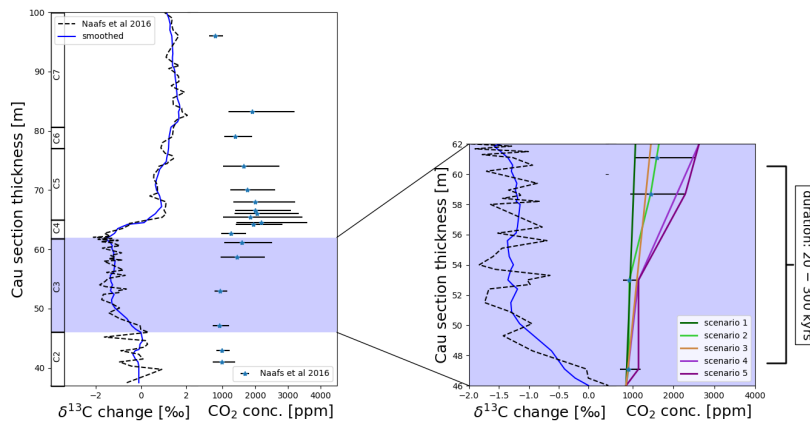
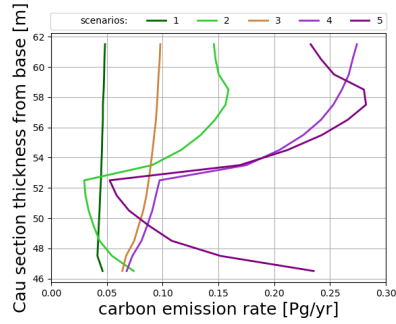
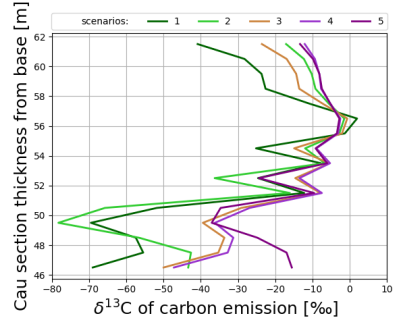


Figure 1: Reconstructed properties of the surficial carbon reservoir during OAE 1a (segments after Naafs et al. (2016)) and $p\text{CO}_2$ forcings used in this study. a) Reconstruction of bulk carbonate $\delta^{13}\text{C}$ variation and $p\text{CO}_2$ across OAE 1a based on proxy data from Cau (Naafs et al. 2016a): b) Zoom into the onset of the negative CIE, showing $\delta^{13}\text{C}$ changes and the 5 $p\text{CO}_2$ scenarios we tested in cGENIE. $\delta^{13}\text{C}$ values are given as offsets from the highest value in the smoothed $\delta^{13}\text{C}_{carb}$ data series prior to the start of OAE 1a.

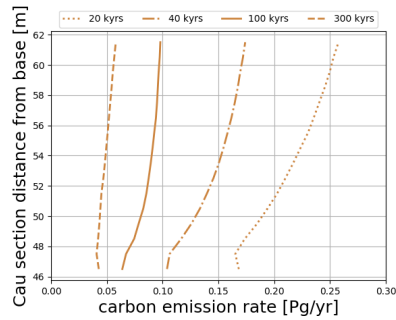
a) carbon mass fluxes across stage C3



b) $\delta^{13}\text{C}$ of carbon fluxes



c) carbon masses with different C3 duration estimates



d) $\delta^{13}\text{C}$ with different duration estimates

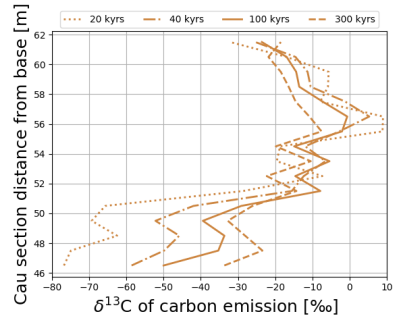


Figure 2: Carbon fluxes and their isotopic signatures across the onset of the negative CIE as calculated by cGENIE. (a) and (b) show differences due to different $p\text{CO}_2$ change scenarios (see fig. 1), all assuming a CIE onset duration of 100 kyr. (c) and (d) show the influence of a different CIE onset duration (20-300 kyr) exemplarily for scenario 3. The results are averaged over 1m depth equivalent.

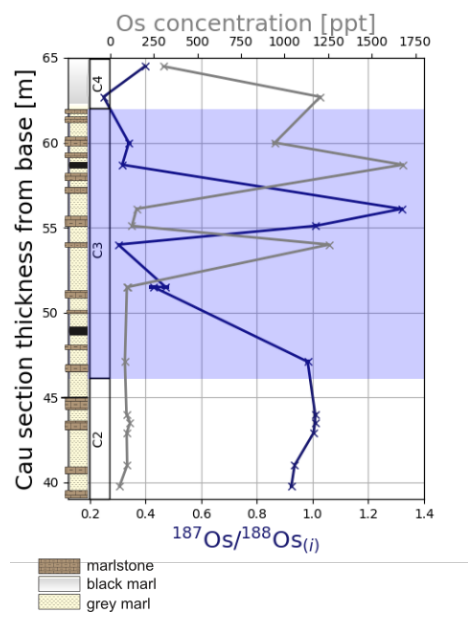


Figure 3: Os concentration and $^{187}\text{Os}/^{188}\text{Os}_{(i)}$ between meters 39 and 65 of the Cau section. The onset of the negative CIE is marked by blue shading. Uncertainties on data are smaller than the symbols.

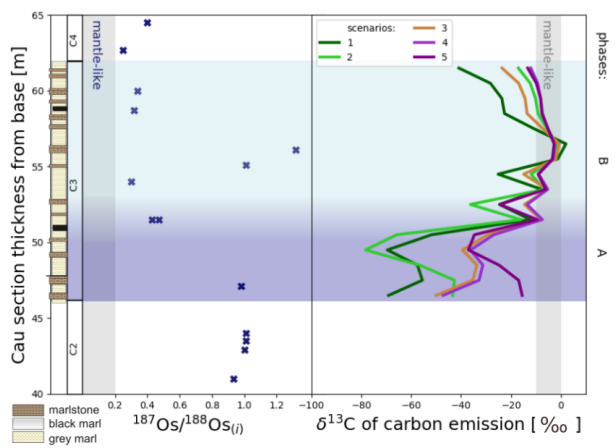


Figure 4: Osmium isotopic changes (left) and reconstructed $\delta^{13}\text{C}$ signature of emissions based on carbon isotopes (right) at Cau. The two shaded sections show phases of least (A) and most prominent (B) volcanic activity.