Clim. Past Discuss.,
https://doi.org/10.5194/cp-2019-118-AC1, 2020

Climate of the Past
Discussions

CPD

Interactive comment

Interactive comment on "Cretaceous Oceanic Anoxic Events prolonged by phosphorus cycle feedbacks" *by* Sebastian Beil et al.

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Received and published: 1 January 2020

Reply to Interactive comment on "Cretaceous Oceanic Anoxic Events prolonged by phosphorus cycle feedbacks" by Sebastian Beil et al. Hugh Jenkyns (Referee) RC1 hughj@earth.ox.ac.uk

Received and published: 17 October 2019

This paper aims to illustrate the potential role of P-cycle feedbacks in prolonging OAE 1a and OAE 2. The paper contains useful data and some excellent diagrams but is rather densely written, skips over some important problems, switches tenses a lot when describing geological phenomena, and ignores some relevant literature. The fundamental point that the low P/TOC ratios in the OAE sediments points definitively to

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phosphorus recycling (and nutrient re-supply to planktonic biota) during these events tends is easily lost. The issue of P recycling during OAEs has, of course, been made previously, including from a modelling perspective (e.g. Mort papers; Nederbragt et al). The value of the account lies in the fact that the sections described are stratigraphically very expanded and give superb detail as to changes in the carbon cycle before, during and after OAEs.

First of all, we would like to sincerely thank Hugh Jenkyns for his insightful and constructive feedback. Following his comments, we revised and streamlined the manuscript to improve readability and highlight the main findings of our study. We also addressed important questions, which were not previously touched upon, and we included missing essential references (listed at the end of this rebuttal).

Abstract and beyond: the statement that the evolution of the carbon-isotope curve of the two OAEs, as classically defined, shows remarkable similarities needs to be qualified.

We will discuss in more detail the similarities and dissimilarities in the evolution of the carbon-isotope curves of the two OAEs in the revised manuscript. We plan to modify the abstract and to insert the following text into section 4.2.2 of the discussion:
The similarities in the general shape of the δ13C excursion of OAE1a and OAE2 (precursor, onset, peak and plateau phase) suggest similar forcing and response mechanisms. However, there are also remarkable differences in the amplitude and duration of individual phases: in particular, the higher amplitude and extended duration of the precursor phase (negative δ13C excursion preceding the onset of the positive δ13C excursion) and the exceptionally long duration of the plateau phase of OAE1a, which in most classic localities is not associated with the deposition of organic carbon rich black shales ("Selli level"). The different durations of the precursor and plateau phases of OAE1a and OAE2 may have been linked to the magnitude and duration of the triggering volcanic exhalations. In addition, different orbital configurations may have influenced long-term marine organic carbon burial on a global scale. Furthermore, obliquity-forced

intensification of monsoonal systems may have resulted in periods of enhanced tropical weathering associated with nutrient supply to the ocean and wind driven equatorial upwelling, which promoted carbon sequestration during the plateau and recovery phases. Periods of low 41 kyr variability in orbital obliquity (obliquity nodes), which occur every 1.2 Myr and are commonly associated with global cooling episodes in Cenozoic warm climate records (e.g., Pälike et al., 2006), may have triggered interruptions or termination of globally enhanced carbon burial.

The defining characteristic of OAE 2 is the overarching positive excursion; for OAE 1a it's the negative excursion. Many OAE 2 sequences (e.g. Eastbourne, UK) show no negative excursion, although its absence is probably due to the presence of the sub-plenus erosion surface in the case of the English section. More needs to be made of all this because the apparently more stratigraphically complete Tarfaya record of OAE 2 clearly offers a unique perspective. The New Zealand record of Gangl et al. (EPSL, 518, 172–182) and Japanese record of Nemoto and Hasegawa (Palaeo-cubed, 309, 271–280) may also show this negative excursion but it is certainly not everywhere apparent.

To address this comment, we propose to insert a new paragraph in section 4.2.1 of the discussion as follows:
The negative carbon isotope excursion at the onset of OAE2 is absent from many classic OAE2 sections in Europe and US Western Interior Basin due to the stratigraphic incompleteness of these records. The missing negative excursion is commonly associated with a short-term hiatus at the onset of the positive excursion, often expressed as a sharp lithological contact (base of the Bonarelli horizon in some of the Umbrian Scaglia sections (Italy) (Jenkyns et al., 2007; Batenburg et al., 2016) and sub-plenus erosion surface in the Eastbourne Section (UK) (Paul et al., 1999; Gale et al., 2005)). However, negative spikes preceding the onset of the positive δ13C excursion were documented in high resolution data sets, even in the relatively condensed Umbrian sections (e.g., Furlo section, Jenkyns et al., 2007) and in more

expanded shelf sections at Wunstorf, northern Germany (Voigt et al., 2008) and Oued Mellegue, Tunisia (Nederbragt and Fiorentino, 1999). Records from expanded sections in Mexico (Elrick et al., 2009) and Japan (Nemoto and Hasegawa, 2011) also clearly exhibit the negative excursion. Recently, a high resolution δ13C record from the South Pacific Ocean and cyclostratigraphic age model based on magnetic susceptibility measurements indicated that the duration of the negative excursion was âĹij50 kyr (Gangl et al., 2019) and allowed correlation of the onset of the negative isotope shift with the beginning of LIP activity at 94.44 $\pm$0.14 Ma (Du Vivier et al., 2015).

As regards OAE 1a, as illustrated in Fig. 4, the main positive excursion extends higher than the C6 segment (1.e. post OAE 1a - unless C6 is extended higher in the section). Do we need a total redefinition of OAE 1a, as implied here? If so, all of this needs to be made clear as perhaps we have been biased by the records of the Cismon and Piobbico cores. But there is a problem: where are the abundant black shales that correspond to the C6 and C7 relatively heavy carbon-isotope segments, given that the original OAE definition is rooted in the quasi-coeval organic-rich record on a global basis?

We agree that this issue needs to be further discussed. We plan to expand the discussion on the discrepancy between the stratigraphic extension of OAE1a black shale occurrences and the positive carbon isotope excursion (section 4.2.1), as follows:
The restriction of the black shale facies (Selli level) to the onset phase of the global carbon isotope curve in many classical OAE1a localities requires an additional carbon sink to account for the globally high δ13C values in segments C6 and C7, if we assume the same mechanism of organic carbon burial is responsible for global positive δ13C excursions. Typical black shales with high TOC values corresponding to the Selli level of OAE1a are not encountered in the stratigraphic succession at La Bédoule.

However, the sedimentary record of black shales and ocean anoxia is patchy across OAE1a; typical examples such as the Resolution Guyot record (Jenkyns, 1995) and the classical records from Cismon and Piobbico (Erba, 1992; Erba et al., 1999) are probably not fully representative of global organic matter accumulation across OAE1a. Preliminary TOC data from more complete sedimentary successions across OAE1a in southern Spain (Cau section and the recently drilled Cau core, Naafs et al, 2016; Ruiz-Ortiz et al., 2016) provide evidence for enhanced organic carbon accumulation during carbon isotope segment C7 (Naafs et al., 2016, Supplementary Figure S4). These sections exhibit unusually high sedimentation rates and a carbon isotope record, which matches that of La Bédoule (Naafs et al., 2016, Supplementary Figure S1), underlining the completeness of both records. In the northern South Atlantic (e.g., DSDP Site 364 offshore Angola), an expanded sediment sequence showing evidence of cyclic episodes of intense anoxia/euxinia was deposited at neritic depths in a large, restricted basin (Behrooz et al., 2018; Kochhann et al., 2014), which may have contributed to organic carbon burial during C6 and C7. There is also evidence of black shale facies extending beyond isotope segment C6 in northeastern Tunisia (Elkhazri et al., 2013), albeit with TOC values in the upper Bedoulian black limestone facies (corresponding to segment C7) significantly lower (< 0.5 percent) than in the black limestones deposited within the C4 segment (4.5 percent).
Another possible candidate for enhanced organic carbon burial during the plateau stage of the δ13C excursion are shallow marine to brackish water sediments in the high Arctic (i.e., Axel Heiberg Island, Canadian Artic Archipelago), where organic carbon rich sedimentation (TOC >5 percent) persisted during most of OAE1a (Herrle et al., 2015). However, the most likely location for burial of larger amounts of organic carbon during OAE1a is the central and east Pacific Ocean. Incompletely cored sequences on the Shatsky Rise indicate deposition of organic matter rich sediments during OAE1a under dysoxic-anoxic conditions at ODP Sites 1207 and 1213 that possibly extended into the upper part of the δ13C excursion (Dumitrescu and Brassell, 2005, 2006; Dumitrescu et al., 2006). Moreover, large areas of lower Cretaceous crust

and sediments along the margins of the Pacific Ocean, which are likely candidates for organic matter-rich sedimentation, have been subducted, leaving less than 15 percent of the lower Aptian seafloor accessible today (Hay, 2007). Thus, we consider the record of global organic carbon burial documented by the positive δ13C excursion to be more representative for the duration of OAE1a than the stratigraphic extent of local black shales. However, this view strongly depends on the interpretation of globally elevated δ13C values as the result of enhanced organic carbon burial. We cannot fully exclude an influence of shallow marine and terrestrial carbonate cycles in maintaining elevated δ13C values in the marine dissolved inorganic carbon reservoir (Weissert et al., 1998). For example, an increase in the proportion of carbonate weathering, relative to organic carbon and silicate weathering, could have maintained long lasting positive excursions in marine δ13C without substantially enhanced burial of organic carbon (Kump and Arthur, 1999).

Line 21: not clear which events are being referred to with 'respectively'

We deleted this word and revised the sentence. The corrected sentence now reads: Based on analysis of cyclic sediment variations, we estimate the duration of individual phases within OAE1a and OAE 2. We identify: (1) a precursor phase (negative excursion) lasting 430 kyr for OAE1a and 130 kyr for OAE 2, (2) an onset phase of 390 and 70 kyr, (3) a peak phase of 600 and 90 kyr, (4) a plateau phase of 1400 and 200 kyr and (5) a recovery phase of 630 and 440 kyr.

Lines 23–25: nutrients may have been supplied by basalt–seawater interaction, probably involving LIPs. (Mentioned later in the text but not here)

We now mention this possible nutrient source as follows:
The extended durations of the peak, plateau and recovery phases imply fundamental changes in global nutrient cycles either (1) by submarine basalt-sea water interactions, (2) through excess nutrient inputs to the oceans by increasing continental weathering

and river discharge or (3) through nutrient-recycling from the marine sediment reservoir.

Line 55: cite original paper by Scholle and Arthur (1980)

We added this reference:
The ensuing positive δ13C excursion is generally attributed to enhanced burial rates of 12C enriched organic carbon in marine organic-rich shales and/or in terrestrial peat and coal deposits (e.g., Scholle and Arthur, 1980; Jenkyns, 1980; Schlanger et al., 1987; Arthur et al., 1988).

Line 66: Are these Mort papers the appropriate references for discussion of transgression? See Jenkyns (1980)

We included the primary citation of Jenkyns (1980) and we extended the sentence to include the additional source of increased terrestrial weathering, as follows:
These include fertilization by nutrient input in the ocean system in association with the activity of large igneous provinces (LIPs) (e.g., Schlanger et al., 1981; Larson, 1991; Trabucho-Alexandre et al., 2010), sea level controlled remobilization of nutrients from flooded low altitude land areas associated with major marine transgressions (e.g., Jenkyns, 1980; Mort et al., 2008), increased phosphorus input resulting from intensified weathering on land (e.g., Larson and Erba, 1999; Poulton et al., 2015), or release of phosphorus as a main limiting nutrient from sediments into the water column under anoxic bottom water conditions (e.g., Ingall and Jahnke, 1994; Slomp and Van Cappellen, 2007).

Line 92: rewrite as: 'A variety of phosphorus species are discriminated against in these sediments.

We revised the sentence as suggested.

Line 98: change 'In contrast' to 'By contrast'

We revised the sentence as suggested.

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Discussion paper

Line 131: hyphenate 'intermediate-resolution' to read as written here

Changed

Line 164: do you mean nannofossils and planktonic foraminifera? 'Shells' rather implies macrofossils.

We revised the sentence as follows:
By contrast, calcium (Ca) is assumed to be of marine origin, mainly originating from calcareous nanno- and microplankton.

Line 189: hyphenate 'metal-free' to read as written here

Changed

Line 271: change 'In contrast' to 'By contrast'

Changed

Line 280: state in which segments of the OAE 1a record the cooling events have been identified. Do they conform to those illustrated in Jenkyns, 2018 (Phil .Trans Roy. Soc.) from multiple localities, namely: C3, C4 and C6? Which cooling events in the OAE 2 record correspond with the Plenus Cold Event? Are these multiple events registered anywhere else? Do they relate to the fact that Tarfaya was a palaeo-upwelling site with upward movements of cooler water or are they global? The largest positive oxygen-isotope shift (Fig. 2) seems to predate the rise in carbon isotopes: i.e. before major global carbon burial was registered, which is not as stated in the text (line 284).

The major cooling events that occurred during C4 and C6 correspond to the global events illustrated by Jenkyns (2018). A minor cooling of probable regional character (Jenkyns, 2018) is also evident during stage C3. We propose to expand section 3.2 and to add relevant references as follows:
The δ18O curves share common trends, despite cyclic lithological changes in the upper part of the sedimentary record of OAE1a in LB3. Transient cooling events, identified by

δ18O increases in LB3/LB1 and SN°4, occur during the early phases of both OAE1a and OAE2 (Figs. 2, 3 and S11).
A first prominent cooling event prior to the onset of OAE2 in Core SN°4 (Kuhnt et al., 2017), which is not identified at other localities, was probably associated with local upwelling of cooler deep water masses in the Tarfaya Basin. Cooling during OAE2 occurred in three main steps starting within the onset phase of the positive carbon isotope excursion. The most intense cooling, associated with the Plenus Cold Event, occurred during the peak phase of the excursion (in the trough between the δ13C peaks a and b). The Plenus Cold Event is globally recorded (e.g., Forster et al., 2007; Sinninghe Damsté et al., 2010; Jarvis et al., 2011; Jenkyns et al., 2017) and coincided with invasion of boreal species in the European Chalk Sea (Gale and Christensen, 1996; Voigt et al., 2003), extinction of the planktic foraminifer Rotalipora cushmani (e.g., Kuhnt et al., 2017) and re-oxygenation of bottom water masses (e.g., Eicher and Worstell, 1970; Kuhnt et al., 2005; Friedrich et al., 2006). OAE1a shows a similar response of global temperatures to enhanced organic carbon burial (Kuhnt et al., 2011; Jenkyns, 2018): the main δ18O increase during the latter part of segment C4 in the δ13C curve of Meneghatti et al. (1998) also occurs during the onset phase. A further cooling event within segment C6 follows transient warming during the peak phase. Both cooling events were recognized by Jenkyns (2018) as global events in the northeastern Atlantic Ocean (Naafs and Pancost, 2016), Italy (Bottini et al., 2015), Turkey (Hu et al., 2012) and in the Pacific Ocean (Dumitrescu et al., 2006). Similarities in the δ18O records across both OAEs imply a similar response of the ocean-climate system to lowered atmospheric pCO2-levels due to excess carbon drawdown associated with burial of vast amount of organic material on a global scale.

Line 318: it would be worth looking at the C-segment durations given by Scott , 2016: (Barremian–Aptian–Albian carbon isotope segments as chronostratigraphic signals: numerical age calibration and durations. Stratigraphy, 13, 21–47) to see how they compare with your data.

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Discussion paper

We have compared our newly reconstructed durations with those from Scott (2016). This comparison is included in Table 2.

Table 2. Durations and sedimentation rates for Aptian C-stages of Menegatti et al. (1998) and comparison with durations of Malinverno et al. (2010) and Scott (2016).

C-stage	duration (kyr)	durations (kyr) (Malinverno et al., 2010)	durations (kyr) (Scott, 2016)
C8	625		
C7	1398	1590	990
C6	315	349	110
C5	281	510	210
C4	388	239	160
C3	434	46.7	80

Line 329: hyphen not necessary in 'orbitally tuned'

Removed

Lines 343–345 and Fig. 2 and Fig. 5: it might be useful to label the features on the OAE 2 carbon-isotope profile (a,b,c,d), as illustrated by Voigt et al., 2017, EPSL. 53, 196–210.

We included the nomenclature of Voigt et al. (2007) in the text and figures to facilitate comparison with global records.

Line 465: 'prevail' - this is present tense and is but one example where past tense should be used for geological narrative. There are many instances of this error in the text. It's also important to maintain clarity when moving from description of an isotope curve to inferences about the environment.

We checked and corrected the manuscript appropriately.

Line 500: compare with the durations given by Scott (see above)

The detailed comparison of the durations of the specific C-stages in section 4.2.1 now includes the durations of Scott (2016). See text below and table 2 above.
Our estimated durations of OAE1a isotope stages agree with those of Malinverno et al. (2010) for the C6 to C8 stages (Table 2), but deviate from the reconstruction of Scott et al. (2016). Minor differences with the estimates of Malinverno et al. (2010) are caused by differing definitions of boundaries between isotope stages in the Cismon core and the LB3/LB1 composite record and by different calculations of orbital periods. The plateau phase (C7) lasted for 1400 kyr in the LB3/LB1 record, in agreement with estimates of 1590 kyr by Malinverno et al. (2010) and of 990 kyr by Scott (2016). The duration of 315 kyr for stage C6 agrees with the 349 kyr estimate by Malinverno et al. (2010), but substantially differ from the 110 kyr proposed by Scott (2016). There are larger deviations from the estimates of Malinverno et al. (2010) and Scott (2016) for stages C5 and C4 in the LB3/LB1 record. Isotope stage C5 has a shorter duration of 280 kyr compared to 510 kyr, estimated by Malinverno et al. (2010), but agrees with the 210 kyr duration of Scott (2016). By contrast, the main increase of the positive carbon isotope excursion, corresponding to C4 (Menegatti et al., 1998), has a duration of 390 kyr in LB3/LB1, which is âĹij60 percent longer than the estimate of 239 kyr by Malinverno et al. (2010) and âĹij140 percent longer than the 160 kyr duration of Scott (2016).

Line 504: change 'In contrast' to 'By contrast'

Changed

Line 552: change 'In contrast' to 'By contrast'

Changed

Lines 579: Mention needs to be made of the key paper by Handoh and Lenton, 2003 (Global biogeochemical Cycles, 17, 1092, who also discuss the cycling of phosphorus to maintain productivity during OAEs. This paper draws on the important papers of Föllmi (Geology, 1995, 23, 503-506; Earth-Science Reviews, 1996, 40, 55–124) that discuss the long-term stratigraphy of phosphorus in the stratigraphic record.

We added these references in section 4.4. The revised sentence now reads:
Phosphorus is the primary limiting nutrient controlling marine biological productivity on longer (geological) timescales (e.g., Holland, 1978; Broecker and Peng, 1982; Smith, 1984; Codispoti, 1989) with the potential to control the occurrence of high productivity events (e.g., Föllmi 1996; Handoh and Lenton, 2003).

Line 581: say how synthesized from atmospheric nitrogen. This will involve a brief discussion on cyanobacteria and papers by Kuypers et al. (Geology, 2004), and others

A short explanation with appropriate references has been added to section 4.4 as follows:
By contrast to nitrate, which can be synthesized from atmospheric nitrogen primarily by cyanobacterial N2 fixation under anoxic conditions (e.g., Rigby and Batts, 1986; Rau et al., 1987; Kuypers et al., 2004), the phosphorus supply to the ocean is restricted by riverine terrestrial input (Ruttenberg, 2003).

Page 615: is 'largest' the right word? Most significant?

We revised the text as follows:
New high-resolution stable isotope and XRF-scanner data were integrated with published records from Cores LB3 and LB1 in the South Provence Basin (Lorenzen et al., 2013; Moullade et al., 2015) and from Core SN°4 in the Tarfaya Basin (Kuhnt et al., 2017; Beil et al., 2018) to contrast the temporal evolution of two of the most significant Oceanic Anoxic Events: OAE1a and OAE2.

Line 622: given that the durations of the carbon-isotope plateau phases are so different, is their causality different as well? We know that the plateau phase of OAE 2 corresponds with maximum organic-carbon burial, at least in the Tethys–Atlantic region - but there is no such evidence for OAE 1a (except possibly Shatsky Rise). So what is going on?

We addressed this fundamental question by rewriting and expanding sub-section 4.2.1: see above reply to comment 3 concerning the definition of OAE 1a and the abundance of black shales that correspond to the C6 and C7 relatively heavy carbon-isotope segments.

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Discussion paper

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