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Introduction to PTRS volume

Placing our current “hyperthermal” in the context of rapid climate change in our geological past

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“...there are known knowns. These are things we know that we know. There are known unknowns. That is to say, there are things that we know we don't know. But there are also unknown unknowns. There are things we don't know we don't know.” Donald Rumsfeld 12th Feb 2002.

Introduction

It is widely recognised that anthropogenic climate change and ocean acidification resulting from the emission of vast quantities of CO₂ and other greenhouse gases, pose a considerable threat to ecosystems and modern society. Global temperatures are already warmer today than at any time in at least the last two thousand years [1], and unabated use of fossil fuel will cause continued warming and sea-level rise potentially for millennia as the climate system slowly adjusts to the enhanced greenhouse effect [2]. Exactly how the climate will respond to this anthropogenic forcing is currently uncertain because our understanding of the climate system is incomplete. There are the things, however, we know we know. For instance, that as we increase the concentration of CO₂ in the atmosphere the climate will warm [3]. Then there are the things we know we don't know, such as the exact value of climate sensitivity and the extent to which it depends on background climate state (e.g. [4]). Then there are the things we know nothing about – the unknown unknowns – which have the potential to take future climate into unimagined directions. Much of the research into predicting future warming involves the use of complex numerical models, climate models, which encapsulate the state of the art understanding of the modern climate system. While these tools can inform on the “known unknowns”, albeit imperfectly as these are often too uncertain to parameterise directly or are emergent properties of the model that are hard to test against observations, they are completely blind to the unknown unknowns as these are by definition not quantitatively represented in any model. Consequently, we urgently need to find alternative tools other than models to investigate them.

The climate has often changed in the past, and these ancient climate events naturally included the response of the climate system to all the feedbacks in operation - even if we currently don't know anything about what these mechanisms might be. Studying the behaviour of the system during these natural climate cycles is therefore one of the only ways to perform a true test of the climate models and whether the system they encapsulate behaves like the real climate system during large CO₂ emission events. Although there are few, if any, actual analogues of anthropogenic climate change in the geological record, there are a number of examples of abrupt climate change that are particularly useful in testing the limits of our best climate models. These include the hyperthermal events of the early Cenozoic. What these events are, what caused them, what effects they had, and what they can teach us about our climate system, are the subject of this special volume.

Background

Human-driven warming is superimposed on the relatively cold pre-industrial “ice-house” climate state, with ice sheets several kilometres thick on both poles. Such ice-house climate states are relatively rare over the last 4.5 billion years, the most recent interval with similar levels of continental glaciation as today was the Carboniferous ~300 million years ago [5]. In contrast, for much of Earth's history, “greenhouse” climate states predominate. These are globally warm intervals characterised by a lack of polar ice caps. The gradual cycling between these two climate states occurs on 100 million-year timescales associated with the growth and destruction of super continents, i.e. at rates much slower than anthropogenic climate change [6]. Superimposed on this grand and slow icehouse-greenhouse climate cycling are a number of geologically abrupt

events – the hyperthermals. They largely frequent the geological record during greenhouse climate states and, although each hyperthermal is unique, they share a number of common features (Table 1). These include:

- (i) rapid warming with a relatively abrupt onset duration of 1,000-100,000 years.
- (ii) a total duration of >0.1 and <2 million years, with the P-T boundary being exceptional and lasting for around 5 million years.
- (iii) a negative carbon isotope excursion of typically less than 4 ‰ near the start of the event indicative of the addition of vast quantities of isotopic light carbon. Again the P-T boundary is larger than the other hyperthermals in this regard.
- (iv) a reduction in oceanic oxygen content sometimes leading to ocean anoxia and/or euxinia. Organic carbon burial typically increases during the event and black shales are often deposited.
- (v) a doubling of atmospheric CO₂ and ocean acidification of around 0.3 to 0.4 pH units. CO₂ rise during the P-T boundary and Palaeocene-Eocene Thermal Maximum (PETM) likely exceeds a doubling.
- (vi) The hydrological cycle intensifies with wet regions generally getting wetter and dry regions drier.
- (vii) an increase of continental erosion/weathering rates.
- (viii) relatively large biotic responses in the first half of the Phanerozoic (Paleozoic and early Mesozoic mass extinctions often associated with hyperthermals), and muted or mixed responses in the latter half of the Phanerozoic.

Although these phenomenon are apparent in a number of the hyperthermals, not every feature is associated with every event and our understanding of each individual event is often far from complete. Nonetheless, the common occurrence of these phenomenon suggests there is a common response of the Earth System to rapid carbon addition. Indeed, the Earth's climate today seems to be responding in such a similar fashion to anthropogenic greenhouse gas emissions (e.g. [7]) that it was recently questioned how we might distinguish between the fingerprint of an ancient industrialised civilisation and a naturally occurring geological hyperthermal event [8].

What use are hyperthermals?

What makes hyperthermals particularly useful is that they represent natural examples of abrupt climate change. However the vagaries of the geological record mean that determining the duration of their onset using traditional geological time keeping is particularly challenging, yet it is the duration of the initial warming that is of particular relevance for our future. Indeed, the imperfections of the geological record mean durations of the majority of the events in Table 1 are likely to be overestimated [6]. In this volume, Kirtland-Turner [9] presents a number of novel ways to better constrain the onset duration of the PETM – the most recent of the hyperthermals (Table 1). Although not necessarily the most impactful hyperthermal, the PETM is key test-bed for understanding the Earth System response to rapid carbon injection because it is the most recent hyperthermal. Kirtland-Turner [9] shows that a number of observations are most consistent with an onset duration for the PETM of around a few thousand years (an order of magnitude slower than our current anthropogenic “hyperthermal”). Although work in this volume on the New Jersey Margin suggests this remains contentious [10], despite previously published estimates of a very rapid onset on the timescale of decades being attributed to an artefact of sediment recovery (e.g. [11]).

New constraints on the magnitude of carbon addition during the PETM have recently arisen due to the development of boron-based proxies that track the state of the ocean carbonate system (e.g. [12,13]). Babila et al. [14] use these novel proxies to show that ocean pH declined by around 0.3 pH units globally during the PETM. These types of observations, coupled with carbon cycle model results [9,15] allow the total carbon added to the Earth system over the entire PETM (170 thousand years; [16]) to be constrained at around 10,000 Pg of carbon [9,15]. During the initial millennial-scale onset modelling by Kirtland-Turner [9] and others (see [9] for references) suggests rates of carbon addition of up to 1-2 Pg C yr⁻¹. While these vast quantities of carbon exceed what humanity have emitted thus far (545 Pg C from 1870 to 2014; www.co2.earth), they are

similar to what we would emit if we used all conventional and non-conventional fossil fuel resources (9,000-14,000 Gt C; [17]). In addition, although the rates of carbon addition during the PETM were on average some 10x slower than modern rates, they are comparable to the average anthropogenic rate over the last 150 years or so (4 Pg C yr⁻¹). These recent studies therefore suggest that the PETM is perhaps a closer analogue to present and future climate change than has been assumed previously.

This close analogy places the findings of Kiehl et al. [18], Schmidt et al. [19] and Benton [20] into sharp focus. The climate rapidly warmed globally around 4-6 °C during the PETM [21,22] and as summarised in Kiehl et al. [18] this dramatically influenced the hydrological cycle – with dry regions such as the interior of the American continent becoming drier and wet regions like the East Asian region, get wetter. By configuring a high-resolution climate model to the PETM, Kiehl et al [18] were able to also show that accompanying the warming across the PETM, the seasonal climate and hydrological cycle became more extreme. These findings are very similar to what is expected in years to come as our climate continues to warm [18].

The PETM, unlike other notable hyperthermals (Table 1), is not associated with significant extinctions, although there is a significant terrestrial mammalian turnover [23] with many mammalian species notably dwarfing across the event [24]. These changes were triggered in part by massive species migration driven by the poleward shifts in the biogeographic boundaries of nearly all terrestrial taxa. Similar biogeographic and population shifts occur in marine plankton as well [25,26]. Schmidt et al. [19] present a study of benthic foraminiferal morphology change across the PETM revealed by micro-CT scanning. Benthic foraminifera are one of the few marine organisms to suffer significant extinction across the PETM [27]. Schmidt et al. [19] show that the impact of the event in terms of benthic foram shell morphology is water depth dependent, suggesting an important role for ocean acidification and ocean anoxia in driving the extinction of these species, as stress-related dwarfing and a potential change in reproductive strategy is evident at the deepest deep ocean sites.

In contrast, the hyperthermal at the P-T boundary was associated with the most severe terrestrial and oceanic mass extinction of the last 541 million years (the Phanerozoic) where 96% of species became extinct – earning the event its alternative name “the great dying” (Table 1). The smorgasbord of kill-mechanisms responsible for such an enormous level of extinction at the P-T boundary are reviewed here by Benton [20]. In the oceans, poster species of the Palaeozoic such as rugose coral, crinoids and trilobites were most probably wiped out by ocean anoxia and euxinia and lethal levels of warming, all serving to squeeze the habitable zone in the ocean out of existence. Whereas on land, plants and animals struggled to cope with the heat-shock as a consequence of mean annual temperatures reaching >35-40 °C and were likely tipped over the edge by the combined effects of warming, acid rain and aridity. Notably, Benton [10] suggests these similar kill-mechanisms were operative during the other hyperthermal events listed in Table 1, albeit at a reduced magnitude.

As summarised by Kump [28], in terms of carbon isotope excursion, the P-T boundary hyperthermal and the PETM share many similarities (Figure 1 of Kump, [28]). Yet the warming after the P-T boundary was more extreme and extended for longer than PETM. Kump [28] examines a possible cause for this contrasting behaviour and in so doing highlights another key value of hyperthermals – they can provide valuable insights into the nature and behaviour of carbon cycle feedbacks, particularly those acting as negative feedbacks (i.e., aiding Earth System recovery from rapid carbon injection), or as positive feedbacks (i.e., adding additional carbon to the surficial reservoir). Arguably these feedbacks represent the “unknown unknowns” – the potential wild cards of climate change.

The way the Earth System recovers from carbon addition in most carbon cycle models is principally via two mechanisms (see for example [29]): (i) on 1000-10,000 year timescales the carbon added during the hyperthermal is dissolved in the ocean lowering pH, which is subsequently partially neutralised through reaction with CaCO₃ in deep sea sediments; (ii) enhanced temperatures increase silicate weathering which draws down CO₂ on 10,000 years to 1 million year timescales. There is an emerging realisation however,

revealed in part by studying hyperthermals, that there is a third important feedback that operates in response to rapid warming related to enhanced organic carbon burial (e.g. [30]). Warm water is able to store less oxygen than cold water and enhanced silicate weathering brings an increased nutrient flux fertilising the ocean with nutrients. Both phenomenon enhance organic carbon burial – a significant sink of carbon dioxide [30]. Here Gibbs et al. [31] pick apart the mechanisms of this process further by examining the size of coccolithophorid cells across the PETM. They conclude that in response to the warming across the event, cell size decreases at shelf sites. Such smaller cells have a higher quota of organic carbon compared to inorganic carbon, which would enhance the export of carbon from the surface waters and encourage its burial and long-term storage in sediments. The hyperthermal events of the Cretaceous are associated with enhanced organic carbon burial and characterised by such thick deposits of organic-rich black shale they are frequently known by the alternative moniker “Ocean Anoxic Events” (OAEs; Table 1). Jenkyns [32] shows that organic carbon burial and CO₂ draw down by the deposition of black shales during both OAE-1a and OAE-2 (Table 1) periodically exceeded the CO₂ addition to the system. This not only drove a reversal of the negative $\delta^{13}\text{C}$ excursion associated with the start of the OAEs as the light isotope was preferentially incorporated in the buried organic matter, but also caused globally distributed rapid cooling to near pre-excursion temperatures and ocean re-oxygenation (e.g. the Plenian cold reversal; Jenkyns [32]).

In the majority of models of the long-term impact of anthropogenic CO₂ emissions, a dynamic organic carbon feedback is currently not included (e.g. see [33] for a review) – yet it clearly is responsible for rapid recovery of the system during hyperthermals (e.g. Jenkyns [32]). This may therefore have significant, and as yet unexplored, implications for the potential longevity of anthropogenic climate change. On a less positive note, it is commonly assumed that the capacity of silicate weathering to draw down CO₂ from any rapid carbon addition is effectively limitless. However, as Kump [28] notes, silicate weathering is currently at least partly supply limited [34]. Thus if the levels of CO₂ addition to the system outstrips the supply of fresh material into the weathering zone by erosion, the efficiency of silicate weathering as a negative feedback is compromised. Current estimates place the magnitude of carbon addition at the P-T boundary to be 30,000 to 40,000 Pg C and Kump [28] proposes at the P-T boundary the arrangement of the continents into the super continent Pangea, led to expansive continental interiors free from tectonism, reducing the capacity of the silicate weathering to draw down CO₂ contributing to protracted nature of the extreme global warmth that occurred during this event ([28]; Table 1). Given the importance of elevated temperature in the kill mechanisms for the P-T extinction [20], this implies that the failure of a key Earth System feedback at this time was at least in part responsible for the extinction of 96% of species on the planet. This not only underlines the importance of the carbon cycle feedbacks in ensuring Earth's habitability, it also places urgent emphasis on determining the silicate weathering capacity of the current Earth System.

The causes of hyperthermals

Ancient industrialised civilisations withstanding [8], a number of triggers have been proposed for the hyperthermals events shown in Table 1, with the PETM once again being the test bed for many ideas. For example, early hypotheses, driven by the co-occurrence of a pronounced negative $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ anomaly at the PETM, suggested that large volumes of methane hydrates were destabilised adding significant quantities of isotopically light carbon to the surficial carbon reservoir [35]. Such a destabilisation could have an external trigger (e.g. volcanism) or could have been the result of ocean warming perhaps in response to orbitally driven changes in ocean circulation [36,37,38]. Other ideas involve the addition of large volumes of isotopically depleted permafrost carbon, again in response to orbital forcing [38]. Such positive feedbacks are the type of “unknown unknown” that could, if they were to become active in the near future, significantly contribute to global warming in the next 100-1000 years.

There are a number of relatively short orbitally paced warming events throughout the late Palaeocene and early Eocene (e.g. [39]), where the added carbon is likely sourced from within the surficial reservoir perhaps as permafrost, dissolved organic carbon and/or methane. The PETM however stands out, both in terms of

magnitude but also in terms of timing by not occurring at the same phase of the orbital cycle as the other smaller events [39,40]. This suggests orbital forcing alone is not likely to be responsible for triggering the PETM.

The emplacement of Large Igneous Provinces are commonly associated with hyperthermals, for example: the Siberian Traps and the P-T boundary, the Central American Magmatic Province (CAMP) with the end Triassic, the Karoo-Ferrar with the Toarcian OAE and the North Atlantic Igneous Province (NAIP) with the PETM, amongst others [41]. Many of these LIPs are not only associated with vast outpourings of volcanic rock at the surface, but they are also frequently associated with magma intruding into sedimentary rocks in the subsurface. Gases such as CO₂ and SO₂ are released directly from the surface volcanism, but as Svenson et al. [42] discusses, the sedimentary rocks intruded by sills and dykes undergo significant metamorphism releasing vast quantities of a variety of other gas (CO₂, CH₄, CH₃Cl, CH₃Br). Importantly, the emplacement of such intrusives is geological fast and can dominate the greenhouse gas budget, indeed, just ~1% of the total area of the Tunguska Basin effected by sill intrusion could generate 1000 Gt CO₂ (272 Pg C; Svenson et al. [42]). The recent high precision dating of the sill emplacement in Siberia and its coincidence with the negative carbon isotope excursion at the P-T extinction event [43] strengthens this link. Similar linkages between sill emplacement and the hyperthermals have also been proposed for the PETM [15,36] and other hyperthermals [43].

Why these intrusive phases occur during the eruption of a large igneous province is widely debated. One suggestion is that the PETM is coincident with the transition from “rift to drift” when mid-ocean ridge basalt-like flows are able to interact with carbon-rich basin fill sediments for the first time [44]. Schaller & Fung [10] present a different, potentially controversial idea based on new evidence from the rapidly accumulating sections of the New Jersey Margin. Schaller & Fung [10] suggest that, in addition to the NAIP, there was also a tektite generating impact coincident with the PETM. The chemical composition of the tektites point to carbonate source rocks, tentatively identify the Marquez Dome as the crater - a 12.7 km wide impact structure in central Texas dated at 58 ± 2 Ma that overlaps with both new ages for the tektites (55 ± 3 Ma; [10]) and the PETM itself (55.9 Ma; [45]). The PETM, like the K-Pg boundary [46], therefore appears to be a time of significant volcanic activity and occurred near synchronously with an impact event. However, if the Marquez Dome is the crater for the PETM impact, it is significantly smaller than the Chicxulub crater at the K-Pg boundary (12.7 vs 150 km diameter). Schaller & Fung [10] suggest that while the impactor, and the vapourised CaCO₃ at the impact site, may have contributed carbon to the surficial reservoir during the onset of the PETM, the relatively long duration of the hyperthermal necessitates additional carbon-sources. Drawing on recent modelling [47] and through analogy with the Deccan Traps [46], Schaller & Fung [10] propose that rather than being directly responsible for the carbon addition to the system at this time, the impact induced an acceleration of NAIP volcanism providing the remainder of the carbon [15]. Hyperthermals are certainly extraordinary events and are relatively rare in the geological record, however more work is no doubt required to determine if their occurrence requires the coincidence of an impact event during LIP emplacement. Nonetheless, given known cratering rates and the typical million-year duration of flood volcanism the coincidence of impacts and LIP emplacement is perhaps not that unlikely [48].

What can we learn about our future?

The hyperthermals of the geological record are not direct analogies of the Anthropocene. They tend to be of a longer duration than is likely for the future, even if we exploit all available fossil fuel [47] and, although they remain poorly constrained, their onset duration is likely slower (1000s rather than 100s of years; [9] and Table 1). It is however evident from the many decades of research that the hyperthermal events can tell us unique information about how the Earth’s climate system behaves following a period of abrupt carbon addition. For instance, they highlight that extensive warming is an inevitable consequence of significant carbon addition and this is likely to be associated with ocean acidification and ocean anoxia, and increase in extremes of precipitation and seasonal temperatures (Table 1 and references therein). In terms of feedbacks, the

likelihood of significant positive feedbacks amplifying any external carbon addition is still uncertain, however it is becoming clear that organic carbon burial is an important, and as yet, underappreciated negative feedback that plays a key part in the recovery of the system.

We have already emitted approximately 545 Pg of C, with a current rate of $\sim 10 \text{ Pg C yr}^{-1}$. Although this rate apparently stabilised in recent years, 2017 AD saw global carbon emissions grow once more (www.co2.earth). These modern rates of carbon emission likely dwarf the rate seen during the onset of the PETM by a factor of 10 or so [9]. If humanity's fossil fuel use is not tackled rapidly through the development of a low carbon economy we face the possibility of emitting as much carbon as was released during the PETM but in a fraction of the time (0.5 vs. 50-100 thousand years; [9,15]). The magnitude of atmospheric CO₂ change (and hence the magnitude of warming, anoxia and ocean acidification) that occurs following any carbon addition to the Earth system is function of rate, due to the timescales of a number of key negative feedbacks [50]. Why the Palaeozoic hyperthermals are associated with significantly greater extinction rate is currently not known. However, a consensus is emerging that it is the extreme heat and anoxia that are the likely "kill mechanisms" [20]. Given that the rate of carbon addition during our "anthropogenic hyperthermal" eclipses that of the PETM, at the very least we are likely looking at a potential future with a more severe impact on life on Earth than any climate change event of the last 56 million years [51]. Exactly how severe however remains perhaps one of the most pressing of the "unknown unknowns".

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Table 1. A summary of the most significant hyperthermals in the last 300 million years

Hyperthermal	Age (Ma)	Approximate onset duration (kyr)	Approximate total duration of warmth/ $\delta^{13}\text{C}$ excursion (kyr)	Magnitude of marine neg. $\delta^{13}\text{C}$ (%- excursion (if present)	Low latitude SST warming (ΔK)	Ocean Anoxia/Euxinia	Approximate pCO ₂ change (from pre event to peak) in ppm	Approximate surface ocean acidification (pH units)	Extinction intensity (% marine species)	References
Paleocene-Eocene Thermal Maximum	55.9	0.1 - 3	170	3-4	3-4	Y	800 to 2200	0.3	1-20	[9,14,15,16,52,22,45,52,53,54]
OAE 2	<i>~93</i>	<i>~100</i>	1322	-	1.5-2	Y	370 to 500	-	20	[55,56,57,58,75]
OAE 1a	<i>~120</i>	1-100	1145	3-4	2-4	Y	1000 to 2000	-	22	[32,55,59,60,61,75]
early Toarcian OAE	<i>~183</i>	<i>~150</i>	300	5-6	2-5	Y	350 to 1200	-	15-20	[55,62,63,64,65,66,76]
end-Triassic	201.6	<i>~85</i>	1200	<i>~1.5</i>	3-4 <i>terrestrial</i>	Y	1120 to 2240 2000 to 4400	-	80	[67,68,69,70,71,72,77]
Permian-Triassic	251.9	10-20	>5000	5-6	10-12	Y	1400 to 4200 2800 to 7000	0.4	95	[13,43,73,74,77]

Italics denote a higher level of uncertainty

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