Response to editor's and reviewers' comments

We thank the Editor and the two referees for your very constructive and detailed comments. We believe that your suggestions significantly improved the quality of the paper. We have reworked the discussion as well as addressed each specific comment of the Editor and those of the referees approved by the Editor. Below, we report our response to each comment in red, (line numbers refer to the marked-up version of the manuscript).

EDITOR'S COMMENTS

The overarching problem behind this manuscript (and the referee comments and unfortunately many published papers) lies with a premise stated on page 1797, namely that the short-lived warming events were DRIVEN by the release of 13C depleted carbon. There remains ZERO evidence that this is the case, at least to my knowledge (although I am open to edification). Clearly, the signals for warming and carbon input are somehow coupled (as nicely shown in this manuscript). However, all information to date, admittedly still sparse and debatable, suggests that warming DROVE in part the carbon input, at least that recorded by the negative d13C excursions. Until someone provides solid evidence that the carbon inputs CAUSED the associated warming, this sort of writing and discussion of data absolutely should be avoided. Indeed, because of this, the discussion (e.g., Page 1803) becomes inconsistent and awkward. Think: how does one invoke orbital changes, presumably in solar insolation, that cause carbon release to drive all temperature change? It almost has to be a coupled system where the carbon release was a POSITIVE FEEDBACK TO WARMING BUT NOT A FORCING OF WARMING.

This is crucial, because it leads to two very important concepts. First, the change in temperature should always exceed that expected from mass balance calculations for the carbon input (as emphasized 18 years ago for the PETM, Dickens et al., 1997); second, and more interesting to the current manuscript, the gradient between d18O and d13C should increase if the time between successive events is significantly less than the recharge rate of the carbon source from which the carbon derived. Both these effects appear in the data of the manuscript, but because the background through presentation sets things opposite, all becomes messy, as one would expect with a wrong framework.

The very first things to change are the axes of Figure 5. It should be d18O on the x-axis (driver) and d13C on the y-axis (response).

We appreciate this comment and we certainly agree that the initial warming triggered the carbon release, representing a positive feedback to warming, as very clearly explained by the Editor. But we also have to agree that this in turn would have amplified the extreme greenhouse warming, explaining the magnitude of warming during the hyperthermals. The most important point we are trying to deliver in the manuscript is that temperature and carbon release co-vary during hyperthermal events that characterize the (pre-) onset phase of the EECO in a similar fashion as discussed by Stap et al. (2010). The phasing between the oxygen and carbon records is another critical issue that we refrained from discussing in this paper. Since we are analyzing the data for each individual event, it would be very difficult to disentangle the initial temperature forcing from the enhanced greenhouse warming due to the carbon release, with the latter possibly dominant in this equation during the core of the event. For this reason and for consistency with the work by Stap et al. (2010), to which we follow up, we prefer to leave the axis in fig. 5 unchanged, as it would otherwise create confusion to the reader.

The next things to change are to provide a proper set-up to the problem and to the discussion.

We have restructured and rewritten part of the discussion (see below) focusing on the covariance between oxygen and carbon. It is now divided into four sections focusing on: 1) isotope covariance; 2) the "paired" events ETM2 -H2 and I1-I2; 3) thresholds and orbital forcing (including a discussion on the EECO): 4) Site 1262 vs 1263.

Additional Editor Comments

(A) The manuscript has EECO in the title, and briefly defines this interval as 52-50 Ma. As pointed out by Slotnick et al., 2012, the use of EECO, especially with these dates, causes confusion. In fact, the manuscript, with this definition, discusses an interval that occurred 2.4 to 0.4 Myr before EECO so the title makes no sense!

The root problem lies in how EECO first came to widespread usage (Zachos et al., 2001), where it was given an age but on an outdated time scale and where some key information was not aligned correctly in the depth domain. However, as one can see in the records of Zachos et al. (2001), the relative age for the start of EECO is clear – it begins approximately at the low in d13C, which we now know is the K/X event and at nominally 52.8 Ma.

In records from the Clarence Valley, New Zealand, there is a marked change in lithology that begins at the J-event, presumably when background temperatures finally passed some threshold. Slotnick et al. (2012) thus suggested that the easiest way to think about EECO is that is began after a "step" coinciding with the J-event, but is characterized by variance (and many CIEs and temperature variations) throughout the next 3 or so million years. Indeed, the data in this paper is really good for discussing EECO – the long-term rise in deep-ocean temperature appears to stop at the J-event. This would be absolutely consistent with the way Slotnick et al. presented things.

Anyway, the presentation of EECO should be fixed and explained, so it does not perpetuate confusion and the title makes sense.

We agree that the definition of the EECO should be fixed in the near future when complete high-resolution benthic isotope record (in particular oxygen records) throughout the entire EECO interval will become available. We deleted the reference to the outdated timeline from the introduction (line 42) to avoid confusion and added a sentence about the EECO and its supposed duration in the introduction (lines 44-46). We included a paragraph discussing the EECO and our definition of its onset, consistent with the way presented by Slotnick et al. (2012). Longer records are necessary to fix the total duration of the EECO but considering the J – X interval as the onset phase of the EECO explains the choice of the title. We hope that this clarifies the ambiguities carried by the definition of the EECO so far (lines 251-258).

The writing in Section 5.1 is not good for multiple reasons, all related to my primary criticism. First, in the original modeling, the “carbon capacitor” does not have to go to zero, if the long-term temperature rise stops; rather, the answer depends on the starting carbon mass, and the “recharge rate” (i.e., the fluxes in and out of the capacitor, Dickens, EPSL 2003). However, the EECO world would be in a situation where large CIEs could not happen.

We have changed the introduction of this paragraph to focus on the covariance of carbon and oxygen that represent the focus of our paper. We have expanded on this issue in Section 5.3.

Second, it is by no means clear that any of the post K/X CIEs were hyperthermals in the current, albeit poorly defined, sense. In fact, if one looks carefully at the Site 1258 record, one will see that, unlike the H-1, H-2, I-1, I-2 and K/X, many of the later CIEs are not associated with a MS spike as expected for dissolution. Of course, there is also the problem that the K/X event is not really recorded in the Site 1258 data.

The paper of Westerhold and Röhl (2009) clearly shows a correlation between $CaCO_3$ and Fe (area) counts of Site 1258 (their figure 2). Accordingly, Fe (area) count values of approximately 600 correspond with $CaCO_3$ values of ~10 wt%. Clearly in the interval above X also the Fe (area) counts reach values in this order of magnitude, which point to severe $CaCO_3$ dissolution. The reason why the MS record does not show these fluctuations might also have to do with its resolution and averaging over a larger area.

Third, a change in the d18O vs d13C “gradient” should be what happens if one accepts that temperature is driving carbon input (at least in part) and there is a carbon capacitor. In fact, the change in gradient should be related, albeit in a complicated manner, to the recharge rate. Basically, for a given temperature change, and over a time frame shorter than the recharge rate, the amount of carbon released has to be less the second time.

This section really should be rewritten and rethought.

This is now discussed in Sections 5.2.

-- Page 1796 --

Lines 24-3: This sentence is awkward in writing and concept. The warming trend is not only in deep marine records, but also in high latitude surface water records (e.g., the cited Bijl et al., 2009 paper); it also culminated in the EECO. Should be something like “… trend that culminated in an extended period of warmth …”

This has been changed into: “From the late Paleocene (~58 Ma) to the early Eocene (~50 Ma), Earth’s surface experienced a long-term warming trend that culminated in an extended period of extreme warmth, called the Early Eocene Climatic Optimum (EECO; Zachos et al., 2001, 2008; Bijl et al., 2009; Westerhold and Röhl, 2009) characterized by the warmest temperatures reached during the Cenozoic.” Lines 40-43

-- Page 1797 --

Line 8: Should be something like “… conditions and rates of change (REF).”

Changed accordingly (line 52)

Line 12: Should be “rose”.

Changed accordingly (line 56)

-- Page 1800 --

Line 6: Probably should be “… the benthic foraminifera values (d18Oc) … “ to be crystal clear.

Changed accordingly (line 169)

Line 10: It would be good to briefly introduce two usually omitted issues (and perhaps even discuss them later). These are the assumption on d18Osw, and that whatever fractionation that causes an offset from equilibrium is constant in the time domain.

This paragraph was expanded clarifying these two issues.

We have added: “The temperature scale is computed assuming an ice-free sea water (δ18Osw) value of -1.2‰ (VPDB). This value is calculated correcting the estimated deep-sea δ^{18}Osw value of -0.98‰ (SMOW) relative to PDB scales by subtracting 0.27‰ (Hut, 1987). The N. truempyi δ^{18}O was adjusted for disequilibrium vital effects by adding 0.35‰ (Shackleton et al., 1984; Shackleton and Hall, 1997), assuming that the isotopic disequilibrium for this species remained constant through time. ” (lines 172-177)

Figure 1: This can be improved, mostly because it conveys limited information relevant to the text. Why not add the locations of other sites that have relevant information across this time interval and are discussed in the text.

Figure 1 was updated with the addition of the other sites mentioned in the text.

Figure 3: The Stap data is not obvious in some of the records.

Thanks for the remark; the Stap et al. (2009) data has now been highlighted clearly in Figure 3 (now Fig. 4) for all records.

Line 1819 – should be “at” not “in”

Changed accordingly (line 772)

There is an important figure missing: the data in the depth domain. There are far too many records in far too many papers that jump to the time domain without clearly showing how things look in the depth domain first. For example, the record at Site 1258 is either spliced wrong or sampled wrong, as there is no significant change in d13C across K/X.

We agree that this is a very important figure. This figure has been added and labelled as Fig. 3. The numbering of the remaining figures was updated accordingly.

Reviewer #1, Prof. Lee Kump

"p. 1797 line 21 and elsewhere: ..ly modifiers don't take hyphens"

Changed accordingly

"p. 1799 line 3-5: Site 1263 suffered some dissolution and developed a clay layer during the PETM, so it didn't remain well above the lysocline throughout the Paleogene."

This sentence has been removed, as it was inaccurate.

"p. 1803 line 26: Need to be clear what the Oligocene-Miocene mechanisms were; expand this sentence."

This sentence has been expanded in the revised version (lines 463-466).

"p. 1806, line 3: I don't know what the authors mean by "scaled biotic response." Perhaps they could state this more clearly."

We have expanded this sentence to clarify the concept of "scaled biotic response", (lines 389-398).

"p. 1806, line 16: The ocean doesn't necessarily get colder with depth, it gets denser. So the authors can't make the conclusion here that the shallower water is saltier (and hence higher d18O). Site 1263 waters could be colder but fresher than site 1262 waters, or saltier but warmer; you can't use the relative depths to differentiate those two possibilities for why the forams at 1263 have heavier isotopic values.

This section has been rewritten (section 5.4)

"p. 1807, line 3: Organic carbon isn't being released into the ocean-atmosphere system; CO2 generated from the oxidation of organic matter might be. . . be more precise with the language."

Thank you for the remark. The sentence has been rephrased. (Line 500)

Reviewer #2, Dr. Philip Sexton

"p. 1801, line 1 – for the tuning process, what is the justification for aligning maximum a* values with maximum eccentricity? (e.g. why not maximum a* values with minimum eccentricity, or some other phase of the cycle?)"

As explained in our response to the Dr. Sexton, we interpret colour reflectance (a*) as a proxy for carbonate dissolution that we relate to eccentricity maxima, (line 199).

"p. 1803, line 20 – should 'specular' read 'speculative'? (I presume it shouldn't read 'spectacular'?)"

This word was removed

“p. 1803, line 24 – I would reference the Kirtland Turner et al. (2014) paper at the end of the following sentence “showing that episodes of carbon release continued throughout the EECO and the onset of the cooling trend” because at the moment it’s ambiguous as to who made that finding.

The reference was added (line 457).

“p. 1803, line 25 – expand on what these mechanisms are, as this relates to the later discussion where the authors discuss methane and organic carbon as sources.

This section has been expanded (lines 463-466).

“p. 1804, line 22 – ‘statically’ = ‘statistically’?

Changed accordingly

“p. 1805, line 15 – “Evidently, the a* values, representative of redness and hence carbonate dissolution”. This is an assumption. The potential controls on % CaCO3 are dissolution, but also dilution and CaCO3 productivity. How can the authors rule out at least a partial contribution from dilution by clays or a reduction in top-down CaCO3 delivery from biological productivity?

As expressed in our reply to the referee, this assumption derives from the results obtained for for the Walvis Ridge sites in the cases of ETM2 and H2 (Stap et al., 2009). The total CaCO3 content was determined to be 96% when all material is preserved and 93% on average for the studied interval, with Site 1263 showing the highest sedimentation rates. We confidently exclude any significant dilution or variation in biological productivity (lines 384-387).

Frequency, magnitude and character of hyperthermal events at the onset of the Early Eocene Climatic Optimum

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Abstract

Recent studies have shown that the Early Eocene Climatic Optimum (EECO) was preceded by a series of short-lived global warming events, known as hyperthermals. Here we present high-resolution benthic stable carbon and oxygen isotope records from ODP Sites 1262 and 1263 (Walvis Ridge, SE Atlantic) between ~54 and ~52 million years ago, tightly constraining the character, timing, and magnitude of six prominent hyperthermal events. These events, that include Eocene Thermal Maximum (ETM) 2 and 3, are studied in relation to orbital forcing and long-term trends. Our findings reveal an almost linear relationship between $\delta^{13}C$ and $\delta^{18}O$ for all these hyperthermals, indicating that the eccentricity-paced co-variance between deep-sea temperature changes and extreme perturbations in the exogenic carbon pool persisted during these events towards the onset of the EECO, in accord with previous observations for the Paleocene Eocene Thermal Maximum (PETM) and ETM2. The covariance of $\delta^{13}C$ and $\delta^{18}O$ during H2 and I2, which are the second pulses of the "paired" hyperthermal events ETM2-H2 and I1-I2, deviates with respect to the other events. We hypothesize that this could relate to a relatively higher contribution of an

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isotopically heavier source of carbon, such as peat or permafrost, and/or to climate feedbacks/local changes in circulation. Finally, the $\delta^{18}O$ records of the two sites show a systematic offset with on average 0.2‰ heavier values for the shallower Site 1263, which we link to a slightly heavier isotope composition of the intermediate water mass reaching the northeastern flank of the Walvis Ridge compared to that of the deeper northwestern water mass at Site 1262.

1 Introduction

The early Paleogene was characterized by a highly dynamic climatic system both on long- (>10^6 years) and short- (< 10^4 years) time scales. From the late Paleocene (~58 Ma) to the early Eocene (~50 Ma), Earth's surface experienced a long-term warming trend that culminated in an extended period of extreme warmth, called the Early Eocene Climatic Optimum (EECO, Ma; Zachos et al., 2001, 2008; Bijl et al., 2009; Westerhold and Röhl, 2009). During the EECO, global temperatures reached a long-term maximum lasting about 2 Myrs, characterized by the warmest temperatures of the Cenozoic (Zachos et al., 2008). Superimposed on the long-term warming trend were a series of short-lived global warming (hyperthermal) events, accompanied by the release of ^{13}C-depleted carbon into the ocean-atmosphere carbon reservoirs (Zachos, 2005; Lourens et al., 2005; Nicolo et al., 2007; Littler et al., 2014; Kirtland Turner et al., 2014). These events are of particular interest as they represent useful analogs for the current global warming, despite differences in background climatic conditions and rates of change (e.g., Zachos et al., 2008; Hönisch et al., 2012; Zeebe and Zachos, 2013).

The Paleocene Eocene Thermal Maximum (PETM or ETM1, ~56 Ma), lasting less than 200 kyr, was the most extreme of these episodes. During the PETM global temperature rose by 5–8°C, and massive amounts of carbon were released as evidenced by a significant negative carbon isotope excursion (CIE) of >3‰ in the ocean/atmosphere carbon pools, and widespread dissolution of seafloor carbonate (Kennett and Stott, 1991; Dickens et al., 1995; Thomas and Shackleton, 1996; Zachos, 2005; Sluijs et al., 2007; Zachos et al., 2008; McInerney and Wing, 2011). A series of similar events are recorded in carbonate records from marine and continental deposits from the early Paleogene, as expressed by negative excursions in $\delta^{13}C$ and $\delta^{18}O$ often accompanied by dissolution horizons (e.g., Cramer et al., 2003; Lourens et

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al., 2005; Agnini et al., 2009; Galeotti et al., 2010; Stap et al., 2010; Zachos et al., 2010; Abels et al., 2012; Slotnick et al., 2012; Kirtland Turner et al., 2014; Littler et al., 2014; Abels et al., 2015). Orbitally tuned records for this geological interval provide evidence that the early Eocene hyperthermal events were paced by variations in the Earth's orbit, specifically in the long and short eccentricity cycles. (e.g., Cramer et al., 2003; Lourens et al., 2005; Littler et al., 2014; Zachos et al., 2010; Sexton et al., 2011).

Several different carbon sources have been proposed to explain the negative CIE, including: (1) the release of methane by thermal dissociation of gas hydrates on the continental slopes (Dickens et al., 1995); (2) the burning of peat and coal deposits (Kurtz et al., 2003); and (3) the release of carbon from thawing of permafrost soils at high latitudes as a feedback or as a direct response to orbital forcing (Deconto et al., 2012); while (4) a redistribution of ^{13}C-depleted carbon within oceans has been proposed as mechanism for hyperthermals in the early to middle Eocene interval (Sexton et al., 2011).

Despite the uncertainty in carbon source and triggering mechanism of the hyperthermal events, a common reservoir has been theorized to explain the consistent covariance in benthic foraminiferal $\delta^{13}C$ and $\delta^{18}O$ across both the PETM and ETM2, indicating that changes in the exogenic carbon pool were similarly related to warming during these events (Stap et al., 2010). The aim of this paper is to test this relationship by constraining the relative timing and magnitude of changes in deep ocean temperatures and carbon isotope excursions for a series of carbon isotope excursions that succeed ETM2, initially identified by Cramer et al., (2003) in the composite bulk carbonate $\delta^{13}C$ record from several deep-sea sites (ODP Sites 690 and 1051; DSDP Site 550 and 577). For this purpose, we generated high-resolution carbon and oxygen stable isotope records of the benthic foraminiferal species *Nuttalides truempyi* from ODP Sites 1262 and 1263 (Walvis Ridge) encompassing the interval from the ETM2 (Stap et al., 2010) to the ETM3 (Röhl et al. 2005), providing the first complete high-resolution benthic stable isotope records for the early Eocene events leading to the onset of the EECO.

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2 Materials and Methods

2.1 Site location and sampling

ODP Sites 1262 and 1263 represent the deepest and shallowest end-member of a 2-km depth transect recovered during ODP Leg 208. Site 1263 is located just below the crest of the northeast flank of Walvis Ridge, in the southeastern Atlantic, at a water depth of 2717 m, whereas Site 1262 was drilled near the base of the northwestern flank of Walvis Ridge at a water depth of 4759 m (Fig. 1). The estimated paleodepths of Sites 1262 and 1263 at ~56 Ma were ~3600 m and 1500 m, respectively (Zachos et al., 2004). The material recovered at the two sites provided an expanded sequence of early Paleogene sediments, yielding a complete section mainly composed of calcareous nannofossil ooze, chalk and marls. The composite depth scale for Site 1263 was constructed using the magnetic susceptibility (MS) and sediment lightness (L*) from the four holes (Zachos et al., 2004).

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Samples were collected at the Bremen Core Repository from Holes A, B and C for Site 1263, and Holes A and B for Site 1262, according to the shipboard meters composite depth section (mcd) (Zachos et al., 2004). A 28-m thick interval of Site 1263 was sampled at a resolution of 5 cm from ~268 to ~296 mcd, and a ~6-m interval of Site 1262 was sampled at a resolution of 3 cm from ~103 to ~109 mcd (Fig. 3). Prior to the analyses, samples were freeze dried, washed and sieved to obtain fractions larger than 38, 63 and 150 µm at University of California, Santa Cruz and Utrecht University.

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2.2 Stable isotopes

Multi-specimen samples of *N. truempyi* were picked from the >150 µm fraction. The stable isotope values of picked specimens (average of 6–8 foraminiferal calcite tests) from Site 1263 were carried out at Utrecht University, using a CARBO-KIEL automated carbonate preparation device linked on-line to a Thermo-Finnigan MAT253 mass spectrometer. Calibrations to the international standard (NBS-19) and to the in-house standard (Naxos marble) show an analytical precision of 0.03‰ and 0.08‰ for $\delta^{13}C$ and $\delta^{18}O$, respectively. The stable isotope values of picked specimens

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from Site 1262 were analyzed on a KIEL IV carbonate preparation device linked on-line to a Thermo-Finnigan MAT253 mass spectrometer, at the UCSC Stable Isotope Laboratory, Santa Cruz. Calibrations to the in-house standard Carrara marble (CM05) and international standards (NBS-18 and NBS-19) yield an analytical precision of 0.05‰ and 0.08‰, for $\delta^{13}C$ and $\delta^{18}O$, respectively. All values are reported in standard delta notation relative to VPDB (Vienna Pee Dee Belemnite). Outliers were defined by adding or subtracting an upper and lower boundary of 2σ from a 13-points moving average, following the method by Liebrand et al. (2011). Published benthic isotope data of the same foraminiferal species for the ETM2 (or H1/Elmo event) and H2 were included in this study to obtain a longer continuous record of Site 1263 and 1262 (Stap et al., 2010) and for I1-I2 of Site 1262 (Littler et al., 2014)

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2.3 Paleotemperature reconstructions

Paleotemperatures were obtained from the benthic foraminiferal $\delta^{18}O$ values by applying the equation of Bemis et al. (1998):

$$T(°C) = 16.9 - 4.38\ (\delta^{18}O_c - \delta^{18}O_{sw}) + 0.10\ (\delta^{18}O_c - \delta^{18}O_{sw})^2 \qquad (1)$$

The temperature scale is computed assuming an ice-free sea water ($\delta^{18}O_{sw}$) value of -1.2‰ (VPDB). This value is calculated correcting the estimated deep sea $\delta^{18}O_{sw}$ value of -0.98‰ (SMOW) relative to PDB scales by subtracting 0.27‰ (Hut, 1987). The *N. truempyi* $\delta^{18}O$ was adjusted for disequilibrium vital effects by adding 0.35‰ (Shackleton et al., 1984; Shackleton and Hall, 1997), assuming that the isotopic disequilibrium for this species remained constant through time.

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3 Age model

Given the typical low resolution age control afforded by magneto- and bio-stratigraphy, and the availability of a robust cycle (i.e., orbital) based chronology for the Leg 208 sites (Westerhold et al., 2007), we developed an eccentricity-tuned age model for the studied interval using the red over green color ratio (a*) records of ODP Sites 1263 and 1262 (Fig. 2). For tuning, we applied first spectral analysis in the depth domain using standard Blackman-Tukey and Gaussian filtering techniques as provided by the AnalySeries program (Paillard et al., 1996). Site 1262, the deepest site at Walvis Ridge, was chosen as the backbone for our tuning. The a* record of this site clearly revealed a ~3-m period, interpreted as reflecting the climatic imprint of the 405-kyr eccentricity cycle (Lourens et al., 2005). Subsequently, we filtered this component and tuned it directly to the extracted 405-kyr eccentricity component of the La2010d orbital solution (Laskar et al., 2011) with maximum a* values, interpreted to represent maximum carbonate dissolution, corresponding to maximum eccentricity values (Table 1). A similar approach was carried out for the a* record of Site 1263 to evaluate the continuity of the successions and robustness of the filtered output (Fig. 2). Finally, the tuned age model of Site 1262 was transferred to Site 1263 by correlating >50 characteristic features in the a* records of both sites as tie points (Fig.2 and Table 2).

Different tuning options have been debated in the last 10 years, resulting in an age for the PETM ranging between ~55.5 and ~56.3 My (Lourens et al., 2005; Westerhold et al., 2008; Hilgen et al., 2010, Dinarès-Turell et al., 2014). Here we report on two tuning options (Fig. 2), assigning an age of 53.69 ±0.02 (option 1) or of 54.09 ±0.02 (option 2) to ETM2 (Westerhold et al., 2007). According to both options, ETM2 predates the 405-kyr maximum falling at an increasing limb, in agreement with observations of Westerhold et al. (2007), but in contrast with the earlier interpretation by Lourens et al. (2005), who aligned this event to a maximum in the 405-kyr cycle. Recent literature revising the Paleocene cyclostratigraphic interpretation (Dinarès-Turell et al., 2014; Hilgen et al., 2015) have shown that the Paleocene holds 25, rather than 24 405-kyr eccentricity cycles. In addition, new U/Pb ages have become available which support an age of ~66.0 Ma for the K/Pg boundary (Kuiper et al., 2008; Renne et al., 2013). These developments point to an age of ~54.0 Ma for ETM2 and therefore we plot our results anchoring the age of ETM2 to option 2 (Fig. 4). Evolutionary wavelet spectra were obtained in the time domain using the wavelet script of Torrence and Compo (http://paos.colorado.edu/research/wavelets). Prior to the analysis, carbon and oxygen records were resampled at 2.5 kyrs, detrended and normalized.

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4 Results

Our new benthic $\delta^{13}C$ and $\delta^{18}O$ records show six major negative excursions between 54 and 52 Ma (Fig. 4). They correspond to the ETM2, H2, I1, I2, J, and ETM3/X/K events, formerly recognized in deep-sea $\delta^{13}C$ bulk carbonate records and land-based

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marine and continental sections (Abels et al., 2012; Agnini et al., 2009; Cramer et al., 2003; Kirtland Turner et al., 2014; Littler et al., 2014; Lourens et al., 2005; Slotnick et al., 2012; Abels et al., 2015)

The general long-term trend in our ~2 Myr long records indicates a minor increase between 54.2 Ma and 53.2 Ma followed by an average decrease of ~0.3 ‰ in absolute values of both $\delta^{13}C$ and $\delta^{18}O$ baseline values following J (~53.1 Ma), with minor cycles evident between the six main events in both records. Following J, both records maintain rather stable values up to ETM3 (Fig. 4). These changes are negligible compared to the Paleocene-Eocene long–term warming trend and long-term negative trend in carbon isotope values. However, the onset of more generally negative $\delta^{13}C$ values, coinciding with J, has also been observed in the deep-sea bulk carbonate record at Site 1262 (Zachos et al., 2010) and in the land-based section at Mead Stream by Slotnick et al. (2012), who suggested that the pronounced change in lithology beginning with J could be used as a chronostratigraphic marker for the onset of the EECO.

Evidence for the onset of warmer temperatures leading to the EECO is evident at ~53 Ma in the benthic $\delta^{18}O$ records at both Sites 1262 and 1263 (Fig. 4). Baseline average $\delta^{18}O$ values prior to ETM2, signifying the response of the unperturbed oceanic system, represent a mean deep-sea temperature of ~12°C, which post-J increases by >0.5°C. Despite variability, our data shows that this increase in background temperature continued upwards across ETM3. Here we suggest that the onset of the EECO can be identified in our records with the onset of the general low in benthic isotope values initiated with J (~53 Ma) and thus including ETM3 within the EECO. Although longer high-resolution benthic $\delta^{18}O$ records are needed to establish the total duration of the EECO, this could represent a first step towards a formal definition of the warmest interval of the Cenozoic, avoiding ambiguity caused by changes in the time scale.

On the short-term scale, our new data across the events following ETM2 and H2 indicate a rise in temperature of ~2 °C and ~1.5 °C during I1 and I2, respectively. The J-event was associated with a temperature increase of >1°C superimposed on the further average decrease in baseline $\delta^{18}O$ value. The ETM3 is expressed in both the shallowest and deepest site at Walvis Ridge by similar isotopic excursions, with a CIE

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of ~0.8‰ and a negative shift in the $\delta^{18}O$ record of ~0.5‰, corresponding to a warming in the deep ocean of 2–2.5°C, comparable to values observed during the ETM2 (Stap et al., 2010).

Evolutionary wavelet analyses for $\delta^{13}C$ and $\delta^{18}O$ records of Site 1263 show spectral power concentrated at distinct frequencies, corresponding to the long 405-kyr and short ~100-kyr eccentricity cycles (Fig. 5). The isotope records reveal coherent patterns, with the highest spectral power concentrated during the ETM2–H2 and I1–I2. The ~100-kyr signal in $\delta^{13}C$, which is very prominent in the first 1 Myr of the record, weakens after J. The imprint of precession and/or obliquity forcing is very weak/absent throughout the entire record. As a result of our tuning approach, minima in $\delta^{13}C$ are approximately in phase with maxima in the 405-kyr and ~100-kyr eccentricity cycles, following previous work (e.g., Cramer et al. 2003; Lourens et al., 2005; Zachos et al., 2010; Stap et al., 2010).

5 Discussion

5.1 Isotope covariance

Our high-resolution benthic isotope records provide a direct constraint on the relationship between the temperature-related signal carried by the benthic foraminiferal $\delta^{18}O$ and the CIEs during the events leading to the EECO. The six events recognised in the benthic records vary in terms of both the magnitude of the CIEs and the inferred temperature changes. The most intense perturbations are associated with the ETM2, I1 and ETM3, whereas H2 and I2, which lag the larger events by one 100-kyr eccentricity cycle, are less prominent (Fig. 4). One important question then is whether all these events of varying magnitude are accompanied by the same source of light carbon released into the ocean atmosphere system and climatic response. Following Stap et al. (2010), we have assessed this by comparing the slopes of the regression line between the carbon and oxygen isotopes of the individual events (Fig. 6). These cross-plots clearly show that all events exhibit significant and coherent linear correlation in both sites with slopes ranging between 0.5 and 0.7 (Figg. 6-7), indicating a consistent relationship for all events between changes in deep-sea temperatures and carbon release. We conclude that this significant covariance between the benthic $\delta^{13}C$ and $\delta^{18}O$ records suggests a strong non-linear response to orbital forcing of global temperatures and the release of

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isotopically light carbon (e.g. methane gas and/or CO_2) into the ocean-atmosphere system during eccentricity maxima, driving subsequent carbonate dissolution and enhanced greenhouse warming, as has been observed in the older part of the record at Site 1262 (Stap et al., 2010; Littler et al., 2014). This conclusion is further underlined by the consistent scaling of CIE magnitudes between our deep-sea data and soil nodule records of the Bighorn basin for these events, which strengthens the hypothesis of a similar isotopic composition of the carbon source for the early Eocene hyperthermal events (Abels et al., 2015).

5.2 The "paired" hyperthermal events

The slopes of the regression lines for H2 and I2 appear slightly steeper than those of ETM2, I1, J and ETM3 (Fig. 6). To statistically test this (dis)similarity, we applied a student t-test to pairs of slopes, comparing all the events against each other using both a pooled and an unpooled error variance. The results show that the null hypothesis (the slopes being similar, α=0.05) is satisfied in the case of ETM2, I1, J and ETM3. The tests on the steeper slopes of H2 and I2 generally display values of $p \leq 0.05$ when tested against the other events, but values of $p \geq 0.05$ when tested against each other. This implies that the smaller events, H2 and I2, are statistically similar to each other but differ slightly from the other perturbations. Even though this statistical approach might be subject to limitations derived from the range of data points chosen for each event, it clearly shows that the slopes for H2 and I2 deviate from the average values given by the other events. Moreover, the statistical deviation of the slopes of H2 and I2 is clearer when comparing them with the average slope for all events of the two sites, since they fall outside the (99.99%) confidence limits (Fig. 7). The average slope between $\delta^{13}C$ and $\delta^{18}O$ of 0.6 for both sites is also in accord with previous observations for the onset/recovery of PETM, ETM2 and H2 by Stap et al., (2010).

The "paired" hyperthermal events, ETM2–H2 and I1–I2 thus reveal slightly different $\delta^{13}C$ vs. $\delta^{18}O$ relationships between their first (ETM2 and I1) and secondary (H2 and I2) pulses. Assuming that these signals are globally representative, this could imply that the second of the two pulses had a relatively larger contribution of an isotopically heavier carbon source than the first pulse. Such a mechanism could hint to a methane-related dominant carbon source (e.g. methane hydrates) during the initial phase of the paired hyperthermal events that is mostly depleted, so that other relatively heavier

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carbon isotope sources (e.g. wetlands, peat) become progressively more important during the successive event. Warming of intermediate water during ETM2 and I1, as was previously been suggested for the PETM and ETM2 (Jennions et al., 2015; Lunt et al., 2010), could have destabilized methane clathrates leading to their dissociation and the subsequent increased warming and large CIE. Mechanisms related to the depletion and subsequent recharge time of the inferred methane clathrate reservoir between ETM2 and H2, and I1 and I2, could explain why the second event had both a smaller magnitude and possibly a smaller relative contribution of methanogenic carbon. The smaller magnitude of the two secondary carbon pulses, regardless of the isotopic composition of their source, seems feasible because the a* values, interpreted as representative of the degree of carbonate dissolution, were significantly lower than during their preceding counterparts (Fig. 2). In other words, the degree of carbonate dissolution associated with the shoaling of the calcite compensation depth (CCD) and lysocline appears to be less severe than during the first pulses. In this respect it is worth noting that H2 and I2 also behave differently from the "larger" events in terms of biotic disruption. During the PETM, ETM2 and I1, rates of variability in planktonic communities indicate that the biotic response was proportionate to the magnitude of carbon injections, and biotic disruption linearly declined with the decreasing size of CIEs (Gibbs et al., 2012; Jennions et al., 2015). However, H2 and I2 do not show evidence of above-background variance, suggesting that during these events the system apparently failed to cross the environmental "threshold" necessary to generate a detectable marine biotic disruption (D'haenens et al., 2012; Gibbs et al., 2012). This all suggests that a change in the climate feedbacks and/or an incomplete recovery of the buffering capacity of the ocean system after the first perturbation could have played a significant role in amplifying the temperature response during the secondary pulse. On the other hand, we cannot dismiss the possibility that local circulation changes and/or partial dissolution slightly altered the anomalies in $\delta^{18}O$ and $\delta^{13}C$ during H2 and I2 at Walvis Ridge. Further research is hence needed to ratify the (global) significance of this finding.

5.3 Thresholds and orbital pacing

The transition towards the EECO is marked by a general decrease of both benthic carbon and oxygen isotopic values of ~0.3‰ at Site 1263, indicative of both long-

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term warming and progressive oxidation of organic matter releasing CO_2 into the ocean-atmosphere system. It has been theorized that the timing and magnitude of the hyperthermals would respond to the crossing of a thermal threshold, more frequently reached in phases of orbital-driven temperature increase (Lourens et al., 2005; Lunt et al., 2011). In addition, the carbon reservoir or capacitor (Dickens, 2003), regardless of its nature and as a result of the long-term temperature increase between the late Paleocene–early Eocene, would be largely depleted by the peak of the EECO, leading to an interval free of hyperthermals. In turn, a series of orbitally paced global warming events of decreasing frequency and increased size are expected to occur during the post-EECO cooling phase when the carbon reservoir would have been progressive refilled (e.g., Kirtland Turner et al., 2014). This hypothesis has been questioned with data from a composite bulk stable isotope record of Site 1258 showing that a series of negative isotope excursions continued throughout the EECO. This evidence suggests that episodes of carbon release persisted during the peak of warmth and the onset of the cooling trend (Kirtland Turner et al., 2014). Kirtland Turner and co-authors (2014) suggest that the mechanisms operating in the early Eocene climate were not necessarily exceptional but actually similar to those invoked for the Oligocene and Miocene when cyclic variations in the carbon cycle were also clearly paced by orbital forcing, particularly in the eccentricity bands (Holbourn et al., 2007; Pälike et al., 2006; Zachos et al., 2001b). Although the carbon and oxygen isotope records of the Oligocene-Miocene and early Eocene are certainly paced by eccentricity, their appearance in terms of punctuation are clearly different. In particular, a relatively sudden release (storage) of large amounts of light carbon (e.g. methane hydrates) into the ocean atmosphere system seems the only way to explain the unusual magnitude of the CIEs recorded at Walvis Ridge given the rate/magnitude of warming, as well as the carbonate dissolution and changes in benthic assemblages associated with those events (Jennions et al., 2015; Stap et al., 2009).

5.4 Site 1263 vs. Site 1262

Comparison between the benthic $\delta^{13}C$ and $\delta^{18}O$ records of Sites 1263 and 1262 reveals an almost identical pattern, although $\delta^{18}O$ values of Site 1263 are consistently ~0.2‰ heavier than those of Site 1262 (Fig. 3 and 4). A similar (reversed) pattern has been previously observed by Stap et al. (2009) in the case of ETM2, and attributed to

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differential dissolution from the shallowest to the deepest site. Conversely, selective dissolution seems unlikely to justify the persistent offset in $\delta^{18}O$ values observed throughout the new post-ETM2 record presented herein. We posit that this offset may be linked to a different average isotopic composition of the water masses at those sites. Accordingly, the intermediate water mass reaching Site 1263 were more ^{18}O-enriched than the deeper waters at Site 1262. The existence of a discrete intermediate water body in the early Eocene South Atlantic is supported by recent benthic foraminiferal assemblage, sedimentological evidence and Earth system modeling data across ETM2, which suggests that warming in the intermediate waters bathing Site 1263 led to differential patterns in sedimentary and ecological data between this site and the deeper Site 1262 (Jennions et al., 2015).

6 Conclusions

New high-resolution benthic stable isotope records from ODP Sites 1262 and 1263 provide a detailed framework to explore the (transient) nature of early Eocene hyperthermal events, during the onset of the EECO. Our results further confirm the link between large-scale carbon release and climate response to orbital forcing, in particular to short- and long- eccentricity cycles. The transition towards the EECO is marked by a general decrease of both benthic carbon and oxygen isotopic values of ~0.3‰ at Site 1263, indicative of both long-term warming and progressive oxidation of organic matter releasing CO_2 into the ocean-atmosphere system. Consistent covariance between benthic carbon and oxygen isotopes demonstrates that global temperatures and changes in the exogenic carbon pool were similarly coupled during each of the studied hyperthermal events. In this regard, we found that the second pulses of the paired hyperthermal events (i.e. H2 and I2) point to a slightly different behaviour. Whether this implies a larger role for a carbon reservoir characterized by a heavier isotopic signature remains debatable and, hence, allows for further consideration of other operational processes such as local circulation changes, partial dissolution, or different climate feedbacks. Finally we found a constant offset in oxygen isotopic values between Site 1263 and 1262, with the isotopic composition of the shallower waters at Site 1263 consistently heavier than at Site 1262, suggesting presence of a discrete water body at intermediate depths of the Walvis Ridge transect.

Acknowledgements

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TABLE 1: Age-depth tie points based on the tuning of the filtered 3 m- period extracted from Site 1262 color reflectance record and the long eccentricity cycle extracted from the Laskar solution La2010d (Laskar et al., 2011).

Site 1262 3-m period filter	**Long eccentricity cycle (kyrs)** ***Laskar 2010d*** **(Option 1)**	**Long eccentricity cycle (kyrs)** ***Laskar 2010d*** **(Option 2)**
102.750	51800	52206
104.231	52003	52410
105.711	52206	52614
107.167	52410	52816
108.648	52614	53017
110.129	52816	53216
111.635	53017	53415
113.193	53216	53615
114.750	53415	53815
116.359	53615	54016
117.865	53815	54218

TABLE 2: Color reflectance tie points from ODP Site 1263 and Site 1262 and interpolated ages obtained from the astronomically tuned age model.

Samples	Site 1263 Depth (mbsf)	Site 1263 Depth (mcd)	Samples	Site 1262 Depth (mbsf)	Site 1262 Depth (mcd)	Interpolated Age (Ma) Option 1	Interpolated Age (Ma) Option 2
1263A-26H-4, 147.5	228.575	265.425	1262B-11H-4, 137.5	92.275	101.855	51.610	52.014
1263A-26H-5, 50	229.1	265.95	1262B-11H-5, 42.5	92.825	102.405	51.727	52.132
1263A-26H-5, 90	229.5	266.35	1262B-11H-5, 102.5	93.425	103.005	51.835	52.241
1263A-26H-5, 115	229.75	266.6	1262B-11H-5, 137.5	93.775	103.355	51.883	52.289
1263A-26H-6, 147.5	231.575	268.425	1262B-11H-6, 45	94.35	103.93	51.962	52.369
1263A-26H-7, 30	231.9	268.75	1262A-10H-2, 120	88.2	104.31	52.014	52.421
1263B-22H-5, 100	230.9	269.23	1262A-10H-2, 145	88.45	104.56	52.048	52.455
1263B-22H-5, 125	231.15	269.48	1262A-10H-3, 20	88.7	104.81	52.082	52.490
1263B-22H-6, 142.5	232.825	271.155	1262A-10H-3, 60	89.1	105.21	52.137	52.545
1263B-22H-7, 45	233.35	271.68	1262A-10H-3, 87.5	89.375	105.485	52.175	52.583
1263A-27H-1, 65	232.75	272.78	1262A-10H-4, 2.5	90.025	106.135	52.265	52.673
1263A-27H-2, 7.5	233.675	273.705	1262A-10H-4, 27.5	90.275	106.385	52.300	52.707
1263A-27H-2, 17.5	233.775	273.805	1262A-10H-4, 37.5	90.375	106.485	52.314	52.721
1263A-27H-2, 25	233.85	273.88	1262A-10H-4, 45	90.45	106.56	52.325	52.732
1263A-27H-2, 125	234.85	274.88	1262A-10H-4, 77.5	90.775	106.885	52.370	52.777
1263A-27H-2, 145	235.05	275.08	1262B-12H-1, 70	96.6	107.24	52.420	52.826
1263A-27H-3, 40	235.5	275.53	1262B-12H-1, 85	96.75	107.39	52.441	52.846
1263A-27H-3, 67.5	235.775	275.805	1262B-12H-1, 100	96.9	107.54	52.461	52.867
1263A-27H-3, 100	236.1	276.13	1262B-12H-1, 110	97	107.64	52.475	52.880
1263A-27H-3, 135	236.45	276.48	1262B-12H-1, 120	97.1	107.74	52.489	52.894
1263A-27H-4, 77.5	237.375	277.405	1262B-12H-2, 5	97.45	108.09	52.537	52.941
1263A-27H-4, 100	237.6	277.63	1262B-12H-2, 22.5	97.625	108.265	52.561	52.965
1263A-27H-4, 137.5	237.975	278.005	1262B-12H-2, 60	98	108.64	52.613	53.016
1263A-27H-5, 70	238.8	278.83	1262B-12H-2, 110	98.5	109.14	52.681	53.083
1263C-9H-4, 105	240.45	280.24	1262B-12H-2, 135	98.75	109.39	52.715	53.117
1263C-9H-5, 15	241.05	280.84	1262B-12H-3, 12.5	99.025	109.665	52.753	53.154
1263C-9H-5, 100	241.9	281.69	1262B-12H-3, 40	99.3	109.94	52.790	53.191
1263C-9H-6, 2.5	242.425	282.215	1262B-12H-3, 57.5	99.475	110.115	52.814	53.214
1263C-9H-6, 15	242.55	282.34	1262B-12H-3, 65	99.55	110.19	52.824	53.224
1263C-9H-6, 32.5	242.725	282.515	1262B-12H-3, 85	99.75	110.39	52.851	53.251
1263C-9H-6, 82.5	243.225	283.015	1262B-12H-4, 10	100.5	111.14	52.951	53.350
1263A-28H-1, 40	242	284.52	1262B-12H-4, 65	101.05	111.69	53.024	53.422
1263A-28H-1, 95	242.55	285.07	1262B-12H-4, 122.5	101.625	112.265	53.097	53.496
1263A-28H-1, 115	242.75	285.27	1262A-11H-1, 137.5	96.375	112.425	53.118	53.516
1263A-28H-2, 40	243.5	286.02	1262A-11H-2, 2.5	96.525	112.575	53.137	53.536
1263A-28H-2, 70	243.8	286.32	1262A-11H-2, 12.5	96.625	112.675	53.150	53.549
1263A-28H-2, 107.5	244.175	286.695	1262A-11H-2, 50	97	113.05	53.198	53.597
1263A-28H-3, 5	244.65	287.17	1262A-11H-2, 67.5	97.175	113.225	53.220	53.619
1263A-28H-3, 27.5	244.875	287.395	1262A-11H-2, 80	97.3	113.35	53.236	53.635
1263A-28H-3, 32.5	244.925	287.445	1262A-11H-2, 85	97.35	113.4	53.243	53.642

1263A-28H-3, 65	245.25	287.77	1262A-11H-2, 97.5	97.475	113.525	53.258	53.658
1263A-28H-3, 70	245.3	287.82	1262A-11H-2, 105	97.55	113.6	53.268	53.667
1263B-24H-2, 147.5	245.875	288.165	1262A-11H-2, 132.5	97.825	113.875	53.303	53.703
1263B-24H-3, 67.5	246.575	288.865	1262A-11H-2, 147.5	97.975	114.025	53.322	53.722
1263B-24H-4, 135	248.75	291.04	1262A-11H-3, 95	98.95	115	53.446	53.846
1263B-24H-5, 47.5	249.375	291.665	1262A-11H-3, 145	99.45	115.5	53.508	53.909
1263B-24H-6, 20	250.6	292.89	1262A-11H-4, 52.5	100.025	116.075	53.580	53.981
1263C-10H-5, 65	251.05	292.93	1262A-11H-4, 57.5	100.075	116.125	53.586	53.987
1263C-10H-5, 82.5	251.225	293.105	1262A-11H-4, 72.5	100.225	116.275	53.605	54.006
1263C-10H-5, 110	251.5	293.38	1262A-11H-4, 87.5	100.375	116.425	53.624	54.025
1263C-10H-7, 1	252.91	294.79	1262A-11H-4, 135	100.85	116.9	53.687	54.089
1263C-10H-7, 5	252.95	294.83	1262A-11H-5, 5	101.05	117.1	53.713	54.089
1263C-10H-7, 10	253	294.88	1262A-11H-5, 10	101.1	117.15	53.720	54.122

Figure Captions

Figure 1: Paleogeographic reconstruction for the early Eocene (~54 Ma) showing the approximate position of Sites 1263 and 1262 (Walvis Ridge), (map provided by Ocean Drilling Stratigraphic Network, ODSN; http://www.odsn.de/odsn/services/paleomap/paleomap.html, modified). Also shown the locations of ODP Sites 690 and 1051 and DSDP Sites 550 and 577 (Cramer et al. 2003) and Mead Stream (Slotnick et al. 2012).

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Figure 2: The floating orbitally tuned age model was constructed based on the red over green color ratio (a*) records of ODP Sites 1263 and 1262. The extracted ~3-m period from Site 1262 was used to tune the record to the extracted 405-kyr eccentricity component of the La2010d orbital solution (Laskar et al., 2011), with maximum a* values corresponding to maximum eccentricity values. Interpolated ages were transferred then to Site 1263 by using age-depth tie points (black dots). Uncertainties in dating proxies prevent an absolute age for this time interval, anchored to the lack of an absolute age for the PETM. Therefore, different tuning options are available within an 800 kyr window (Westerhold et al., 2008). Two possible options are shown.

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Figure 3: Benthic *N. truempyi* $\delta^{13}C$ and $\delta^{18}O$ records from Site 1263 and Site 1262, plotted versus depth. Highlighted intervals represent the position of the early Eocene hyperthermal events.

Figure 4: Benthic *N. truempyi* $\delta^{13}C$ and $\delta^{18}O$ records from Site 1263 and Site 1262, plotted versus Age (Ma), (starting from option 2 for the age of ETM2- Westerhold et al., 2008). Highlighted intervals represent the position of the early Eocene hyperthermal events.

Figure 5: Evolutionary wavelet analyses for $\delta^{13}C$ and $\delta^{18}O$ were performed using a Morlet mother wavelet of an order of 6. The shaded area represents the 95% significance level. Spectral power above the confidence level is concentrated at distinct frequencies, corresponding to the long 405-kyr and short eccentricity 100-kyr cycles. Highlighted intervals represent the position of the early Eocene hyperthermal events.

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Figure 6: Relationship between the oxygen and carbon isotope values of *N. truempyi* during ETM2, H2, I1, I2, J and ETM3/X at Site 1263 and Site 1262. Data for ETM2

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and H2 from Stap et al., (2010) and for I1, I2 and J at Site 1262 from Littler et al., (2014). Note that, because of intense dissolution at Site 1263, ETM2 data were chosen from Site 1265. For all the events, throughout the entire event (onset+recovery phases), changes in the exogenic carbon pool are linearly related to warming. Linear regression equations refer to Site 1263 (top) and Site 1262 (bottom), respectively.

Figure 7: Slope of each event plotted together with the average slope (from all the events). The red dashed line indicates the 99% confidence interval.

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