Response to Referees and Revised Manuscript

Herein we provide a detailed point-by-point response to all referee comments and specify all changes in the revised manuscript. The response to the Referees is structured as: (1) comments from Referees, (2) author's response, (3) author's changes in manuscript. We also include a marked-up manuscript version showing the changes made (using track changes in Word). This version should be combined with your response file so that the Editor can clearly identify what changes have been made.

Response to Referees

Referee 1

Comment 1: A new multi-proxy record of the PETM from a deep sea core in high southern paleolatitudes is very welcome and adds to our understanding of this very large and still enigmatic climate perturbation. The authors should be congratulated for finding the PETM in such an old drill-core. There can be little doubt that the data should be published and the study is suitable for Climate of the Past. However the data set is moderately complex and (I think) difficult to interpret unambiguously, hence the challenge is to distil the salient points as clearly as possible for the wider community interested in this event, and for comparing records. Hopefully the review process can help with that. Some of my comments are relatively minor and can probably be dealt with quite easily but I have two more substantive issues to deal with first.

Response 1: Agreed, thank-you.

Comment 2: The relation between the bulk d13C record and the so-called ’onset’ of the PETM, and what happens near the base of the event. The PETM is by definition a warming event hence its onset (for me) is, perforce, the bit where it gets hotter, i.e. where the d18O goes negative (and is also seen in the Mg:Ca excursion in this study). This is the interval 457.25-457.29 m which corresponds to a 4cm chunk of core. However even the ’intermediate’ bulk d18O and whole foram values from within this chunk may be affected by differential diagenesis, as the paper rightly points out, and some bioturbation mixing is also likely given the burrows, so it is not clear that the real ’onset’ of the thermal maximum is sampled at all in this drill core. The paper mentions almost in passing that the portion directly below the PE boundary (457.3-457.58) has sparse and poorly preserved foraminifera (p. 256 line 3-4) but offers no explanation. This suggests the possibility, even likelihood, of dissolution burn-down caused by a transient rise in the carbonate compensation depth during the real onset, also the time of the benthic extinction, and what the core records would therefore be renewed sedimentation onto a dissolution surface / semi-hiatus.

Most previous records of the PETM have the onset of warming coincident with a large and abrupt negative carbon isotope excursion (CIE), so that the onset of the CIE is pretty much coincident with the onset of the PETM - the coupled excursions mentioned in the introduction. But this record is quite different, basically un-coupled, and I have to admit to being puzzled by the run of intermediate bulk d13C values through most of what appears to be the body of the PETM, even while bulk d18O is becoming slightly more positive. This, for me, is the enigmatic aspect of the record and the discussion (Section 3.3) does not wrestle it down sufficiently.

Part of the problem (I contend) is that the authors refer to the gradually decreasing bulk d13C values as the 'onset of the PETM' (in title, abstract line 8; discussion in Section3.3) which only confuses matters. The d18O and Mg:Ca show that this interval is in fact the 'body' of the thermal maximum. It may be the onset of the CIE at this location, I grant, but I doubt whether it correlates with the onset of the CIE everywhere else. Why the bulk d13C values are a run of intermediates is difficult to explain, especially given the foram d13C values around 457.22 m appear to be more or less 'fully negative' and similar to those above from around 467.03 m.

I do not offer an explanation for this conundrum, all I can think of is some far-fetched ideas, but the authors need to tackle it head on, and perhaps take the opportunity of clarifying what they mean by onset of the PETM / CIE / event and making sure the text is clear throughout. Incidentally, the claim that the 'onset' lasts 50-100 kyr in other sites (p. 258, line 4) contrasts with what I think most people would think of as around 10-20 kyr or less, with the 'body' being about 150-200 kyr (the Mead Stream section is also peculiar).

Also incidentally, if the lower carbonate content is because of reduced carbonate flux then the PETM here would also represent more time than an interpolated or extrapolated age model, possible equivalent to much of the body of the PETM (see also my comment on cycles below).

Response 2: We admit that we have used the term "onset" too loosely in this article. Strictly speaking we argue that the recovery interval is missing, and so what is preserved is the onset plus at least part of the core of the PETM. We will revise use of "onset" throughout the text. However, we still contend that the bulk d13C record is consistent with a relatively complete PETM. There is certainly an interval below the PETM where carbonate content is lower than background for the Paleocene, but the d13C is very stable through the upper 50 cm of Paleocene sediment, with no evidence for "burn down" affecting the d13C record and hence no basis to infer a hiatus at the base of the PETM.

The stepped decrease in d13C that occurs between the true onset at 457.27 mbsf and the core of the PETM at 457.1 mbsf is not that unusual. As well as Mead Stream, it has been widely discussed for ODP Site 690 (Bains et al., 1999; Stoll, 2005; Zachos et al. 2005; Sluijs et al. 2007), and has led to speculation that the PETM carbon release may have occurred in pulses (Sluijs et al., 2007). A point that we emphasise in the revised MS is that the three steps in the CIE at DSDP 277 correspond reasonably well with the three maxima in Mag Sus and Fe. Therefore, we can be reasonably confident that we have CIE steps A-C of Zachos et al. (2005) that span cycles 1-3 of Röhl et al. (2007), and therefore our PETM interval probably spans the interval from the base of cycle 1 to the base of cycle 4 or the first ~66 kyr of the PETM (Röhl et al. 2007). Stoll (2005) has shown that the bulk carbonate d13C is a good representation of the coccolith record at ODP 690, which also has a 3-stepped CIE very comparable to Site 277 and Mead Stream. We note that Site 690 is considered to be one of the most expanded records of the PETM and was used to calibrate deeper water records in the South Atlantic where burn-down effects are greater (Zachos et al., 2005).

Differences between the bulk carbonate and single foram signals for the PETM are also well known at ODP 690. As at DSDP 277, Acarinina records a strong CIE at the onset of the PETM. Stoll (2005) has an extensive discussion on this topic but no definitive answer. All we need to say here is that Site 277 is not unusual in having a stepped bulk carbonate CIE. This is why we focus more on the much more pronounced bulk carbonate oxygen isotope excursion (OIE).

We tend to agree with the referee's comment that the fact the PETM is a warming event is often overlooked. Nevertheless, since first identified, the PETM has consistently been defined by the carbon isotope excursion (CIE) that provides a less ambiguous way of correlating the event between geographic regions and of subdividing components of the event than the variable oxygen isotope excursion or other climate proxies (e.g. Bains et al., 1999; Zachos et al., 2005; Sluijs et al., 2007; McInerney and Wing, 2011; Dunkley Jones et al, 2013). The P/E boundary is defined by the CIE (Aubry et al. 2007), not the associated temperature shift or other climate-related phenomena. There are many locations world-wide where the location of the PETM is fixed only by the CIE, with very little known about the accompanying changes in temperature. It is also worth making the point that in the case of modern global warming, associated climate change exhibits considerable geographic variation whereas the steady increase in atmospheric CO2 is uniform across regions. This is nicely demonstrated by the compilation undertaken by Dunkley Jones et al. (2013).

*Changes to MS (with reference to underlined comments): We reduce reference to "onset" in the MS and clarify where appropriate that we are referring to the onset and body of the CIE (up to the CIE minimum, **Fig. 7**; **Lines 392-393**) but note that the PETM is defined by the CIE, not by warming (**Line 107**). We argue that neither bioturbation nor burndown can explain the "uncoupled" isotope record or the pre-PETM interval (**Lines 465-472**). We show that the isotope records for Site 277 and Mead are not that unusual but very similar to Site 690 (**Lines 473-499**). We reassess the age based on correlation with precession cycles (**Lines 432-435**).*

Comment 3: Foram preservation and diagenesis: The authors indulge in quite an extended discussion about possible diagenetic effects on foram tests and d18O and Mg:Ca. They assert, reasonably I suspect, that the preservation is better in the PETM than below and above because of inhibited recrystallization related to the higher clay content. I have often wondered whether such an effect might be at play in some PETM records which go from chalk to clay (if not cheese) - and indeed other such events such as the Mediterranean sapropels. To demonstrate this effect would be very neat. However the reader cannot evaluate the likelihood of this because no SEM evidence is presented. This paper would benefit hugely from a comparative textural study of the foram walls in internal section under high powered SEM so that the extent of recrystallization can be visually assessed relative to published criteria (which we have produced for just this kind of material - Pearson, P.N., and Burgess, C.E. 2008. Foraminifer shell preservation and diagenesis: comparison of high latitude Eocene sites. Pp. 59-72 in Austin, W.E.N. and James, R. H. (eds.), Biogeochemical Controls on Palaeoceanographic Proxies. Geological Society Special Publication 303: 59-72. Doi: 10.1144/SP303.5). This work has been done ('stringent [SEM] screening criteria were applied to exclude zones within test walls that show evidence of diagenetic alteration' - p.252, l. 5-6) so it really should be shown. Actually I will challenge the authors: given that the carbonate content is fairly high even within the PETM, I predict that all forams at this site will show clear evidence of diagenetic alteration on a micron scale throughout their entire walls (although the effect maybe variable) - I do not believe you can exclude recrystallization at this site - so go prove me wrong by illustrating the typical biogenic microgranular texture of a well-preserved foram!

It would also be nice to see what is causing the reputed poorer preservation of the forams immediately below the PETM - are they dissolved? In short, I think we should be moving away from near-meaningless phrases like 'carbonate microfossils are moderately well preserved' (p. 256, line 3) and ’relatively well preserved tests’ (p. 258, line 23) to something much better documented, but for that we need the SEM images.

*Response 3: We are forced to admit that upon review of our SEMs, all the forams in this record are affected by diagenesis, with clear evidence of overgrowths and recrystallisation. We cannot confidently use visual examination as a basis for stating that the forams in the PETM are any better preserved than below or above the PETM. The “stringent” criteria referred to in the text relates to geochemical criteria. Specifically in relation to diagenesis, this means we have not used specimens or parts of LA-ablation profiles with Sr/Ca values lower than 0.8 or higher than 1.6 in the determination of paleotemperatures from Mg/Ca ratios. Following the recommendation of Referee 2, we have constructed cross-plots to show how these criteria are applied (**new Fig. 3**).*

These cross-plots explore the relationship between Mg/Ca ratios, silicate contamination and diagenesis. The full dataset consists of average elemental ratios across selected integrated laser ablation profiles (visually screened for the effects of silicate contamination and diagenesis). This dataset has been first plotted as cross-plots of Al/Ca and Mg/Ca for Cibicides and Acarinina (the only two genera that have records spanning the entire studied interval). The data have been colour coded to differentiate samples from within the PETM and the pre-PETM interval extending down to 462.2 m. A significant linear correlation is observed for both genera, confirming that silicate contamination results in anomalously high Mg/Ca values where Al/Ca >4 (3.4 and 3.7 being the calculated screening limits with method explained in the MS). For Cibicides, silicate contamination is seen to be more common within the PETM than in the pre-PETM interval or elsewhere in the record. This is consistent with other evidence for an increase in clay within the PETM. This pattern is also evident for Acarinina but Al/Ca values are generally higher and more pre-PETM samples show evidence of silicate contamination. This may reflect differences in morphology, with the smoother surface of Cibicides providing fewer opportunities for clay particles to become fixed to the test wall.

The second set of cross plots consists of data that has first been screened for silicate contamination using the screening limit of Al/Ca > 3.4 or 3.7. The Sr/Ca ratio is used as a proxy for diagenesis, with the typical ratio in Eocene foraminifera being 1 to 1.4 (Creech et al., 2010) and the values < 0.8 or > 1.6 are considered to reflect dissolution or recrystallization (Eggins et al., 2003; see also Kozdon et al., 2013). Linear regression lines are shown for the three groups and the correlation co-efficient is added where it is significant. For Cibicides, most values lie within the screening limits but there are some significant differences in the clusters of PETM and pre-PETM samples. First, we note that the overall trend for all the samples is roughly horizontal, suggesting that the Sr/Ca ratio is generally stable around ~1 for varying Mg/Ca. However, we observe that there is a weak negative trend and stronger and steeper negative trend for PETM and pre-PETM samples, respectively. This suggests that diagenesis has a significant effect on pre-PETM Mg/Ca values for Cibicides, and may also have a small effect on PETM values. The net effect would be to introduce a warm bias to pre-PETM temperatures, and possibly also to some PETM temperatures.

For Acarinina, a clearer relationship is observed between Sr/Ca and Mg/Ca. The effects of diagenesis appear more evident, overall, probably reflecting the thinner test and greater surface area relative to the smooth walled Cibicides. The overall trend is consistent with predictions and observation form other studies: a negative correlation indicates that as diagenesis progresses Sr/Ca decreases and

Mg/Ca increases (Eggins et al. 2003; Kozdon et al. 2013). The slope of the trend lines indicate that the pre-PETM interval is most affected by diagenesis, and the PETM is the least affected. As for Cibicides, this implies that temperatures based on the Mg/Ca are significantly overestimated in the pre-PETM interval and also, but to a lesser extent, overestimated in the PETM.

In summary, we have shown how screening has been applied to exclude samples with significant levels of silicate contamination and diagenesis. We have shown that diagenesis tends to be higher in the pre-PETM interval than in the PETM. We have shown that Acarinina is more affected by both silicate contamination and diagenesis than Cibicides. And we have shown how diagenetic effects may affect temperature estimates (even for samples within the screening limits), probably overestimating pre-PETM temperatures relative to the PETM.

Therefore, although we lack visual evidence to support our hypothesis that the extreme OIE at the "onset" of the PETM in this record is an artefact of preservation changes across the P/E boundary, we do have geochemical evidence in the form of Sr/Ca ratios. In addition we observe and increase in the d18O gradient between bulk carbonate and benthic foraminifera: from <0.6 in all but one sample below the PETM to >1.1 in 2 of 3 samples within the PETM. A low gradient is also observed above the PETM (0.23 – 0.42).

We have added a supplementary figure to show the general distinctions between moderate and poor preservation, based on SEM images and LA-ICPMS profiles, but acknowledge that others may term these preservation states, poor and very poor.

*Changes to MS: it is beyond the scope of the current study to include detailed study of wall texture ("stringent criteria" referred to geochemical screening), but we have included SEM images of whole tests (**Fig. S1**) and accept that preservation is generally poor to moderate (**Lines 353, 449-454**).*

Comment 4: p. 246, line 3. Reword? These references are compilations, not original studies, but in any case all early Paleogene pCO2 reconstructions are very uncertain - and yet the statement seems to take it as fact.

*Response 4: Insert "is inferred to have" (**Line 45**).*

Comment 5: p. 250, line 9. *Acarinina* is spelt wrong. Were species combined in these analyses? We know *Acarinina* species can differ quite a bit in habitat, so it would be better to stick to species where possible as in the trace elements.

*Response 5: Move the text from Mg/Ca methods to this location (**Lines 151-166**), and revise to say: "Individual specimens from five foraminiferal genera were used for stable isotope analysis and elemental geochemistry. Specimens were selected for analysis based on visual assessment of their preservation under a stereo microscope. Wherever possible, analyses were performed on Morozovella aequa, Acarinina coalingensis, Subbotina patagonica, S. roesnasensis, and Cibicides proprius/praemundus. The following species were substituted when these species were not available: Morozovella subbotinae, M. acuta, M. apanthesma, Acarinina soldadoensis, A subsphaerica, A. esnaensis, A. nitida and Cibicidides tholus. The stable isotope signature of Acarinina soldadoensis, A subsphaerica, A. nitida and all species of Morozovella indicates they were mixed layer dwellers (Olsson et al., 1999; Quillévéré and Norris, 2003), and therefore are appropriate indicators of near* *surface conditions. Subbotina patagonica is inferred to have had a deep planktonic habitat (Pearson et al., 2006), within the thermocline. There is no data on the habitat of S. roesnasensis."*

Comment 6: p. 256, line 12. The missing NP7 and NP8 are not shown on the log on Figure 2, where the zones are combined. Hence the text says something important that is absent from the fig.

Response 6: The zones are missing, so they are not shown on Fig. 2. An unconformity is shown to lie between NP6 and NP9, which is consistent with the text.

Comment 7: p. 256. Can you say something about planktonic forams in the PETM? Did you find any excursion taxa or influx of warm forms?

*Response 7: Yes, this important text to be added (**Lines 338-341**). "The genus Morozovella has its lowest occurrence at the base of the PETM and greatest diversity within the PETM. Two species of Morozovella are restricted to the PETM: M. aequa and M. velascoensis. The latter species has rarely been found outside the PETM in the SW Pacific but M. aequa ranges into the middle late Eocene on mainland New Zealand (Hornibrook et al. 1989)."*

Comment 8: p. 257, line 9. Can you prove the Fe record is cyclical by power spectra? Are these cycles likely obliquity? If so you have virtually all the PETM here, not the 'onset'!

*Response 8: There are two few cycles present for spectral analysis or to conclude that there are four obliquity cycles. However, we have commented earlier on the likelihood that that they correspond to the precession cycles described by Röhl et al. (2007). Reword statement of duration of onset and body based on correlation with Site 690 (**Lines 359-361, 396-403**).*

Comment 9: p. 257, line 18. This positive shift in d18O of 0.4 per mil is not obvious to me on Fig 2. I wonder if this is a burn-down dissolution effect, also the enigmatic peak in Fe and mag. susc. which you say the cause is unknown.

*Response 9: The shift is there, but not obvious how this could be due to burn down. Also as discussed there is no equivalent shift in d13C. We interpret it as suggesting pre-PETM cooling (also noted at Site 690). The enigmatic shift in FE and mag sus is well below the PETM (**Lines 377-385**, slightly reworded; **Lines 433-440**, more commentary)*

Comment 10: p. 260. "Benthic foraminifera tests are dense [true] and thus less prone to recrystallization [not true?]" This has been claimed many times, and I blame myself partly for it, as we all hoped it would be true, but whenever we have actually studied the tests they are in fact just as recrystallized as the plankton. Again, you could substantiate this with SEMs or omit the statement / claim.

*Response 10: Unnecessary sentence deleted (**Line 533**)*

Comment 11: p. 262. Is there any paleontological evidence for a change in fossil assemblage that perhaps might be related to the subtropical gyre?

*Response 11: Incoming of Morozovella and diversity of the genus to be noted here (**Line 633-634**).*

Comment 12: In summary this is a fascinating new record from the high southern latitude but the text needs to be clearer regarding what is meant by the onset of the event; and images from the textural-diagenetic SEM study that has been done would improve the case for understanding what has been observed regarding foram test preservation and quantifying the temperature excursion.

Response 12: Endeavoured to address these issues in further discussion of the effects of diagenesis but have not included detailed SEMs of test walls. Whole test SEMs and Element/Ca profiles serve to (new Fig. 3) illustrate issues with silicate contamination and diagenesis.

Referee 2

Comment 1: A sediment core drilled in 1973 on the western margin of the Campbell Plateau by the Deep Sea Drilling Project (DSDP Site 277) was re-examined by Hollis et al. Within the past 40 years, a large number of studies was published based on material from this core, however, Hollis et al. are the first to discover that the PETM is preserved in a 34 cm-thick interval within these sediments. The authors used an innovative multi-proxy, multi-instrument approach (in situ measurements of trace elements in individual planktic and benthic foraminiferal shells by LA-ICP MS, d18O and d13C analyses in bulk carbonate and foraminiferal shells, XRF scans of core sections, measurements of carbonate content and magnetic susceptibility) to delineate the PETM recorded in these sediments and to compile a robust paleorecord by combining data created by this multi-proxy approach to assess the degree of alteration of foraminiferal shells and exclude Mg/Ca measurements from samples that underwent significant diagenesis.

While this study certainly warrants publication and is suitable for the journal “Climate of the Past”, I highly encourage the authors to address the potential impact of diagenesis on the Mg/Ca, d18O, and d13C values in more detail. According to the authors (page 250, lines 6 ff.), “samples were selected based on light microscope assessment of preservation, which was subsequently confirmed by SEM”. In my experience, this approach of sample screening is not sufficient. Within the past years, I analyzed hundreds of foraminiferal shells in high magnification by SEM and screened more than one thousand polished chamber wall cross-sections in epoxy mounts using SE, BSE, and CL detectors in order to locate suitable, well preserved domains for SIMS analyses. Thereby, I realized that it is practically impossible to estimate the actual preservational state of ‘frosty’ foraminiferal shells from their outer appearance. Two shells, that appear to feature the same preservational state, may turn out to be altered to completely different degrees when examining their polished chamber wall cross sections. Thus, my concern is that differential diagenesis within the studied core section may have been overseen. For example, it was shown in previous studies that foraminiferal shells at the PETM onset may be altered to a larger degree due to the effect of ‘burn-down’ (e.g. Walker and Kasting, 1992; Kozdon et al., 2013). Hollis et al. emphasize a different trend in PETM-warming at DSDP Site 277 compared to other sites, but I am not convinced that the data shown in this study exclude the possibility that part of this observation could also be caused by differential diagenesis within the core. However, the laser ablation data provide a great wealth of information that can be used to assess foraminiferal diagenesis in more detail, and I am surprised that the authors didn’t explore all the benefits of this in situ approach. Based on the method description and the data presented in Fig. 3, the Element/Calcium ratios of the three laser ablation analyses (=profiles through chamber wall) for each shell were simply averaged. If the (averaged) Al/Ca and Sr/Ca ratios exceeded certain threshold values, the shell was considered to be significantly affected by alumina-silicate contamination and/or diagenesis, consequently, the Mg/Ca ratio was not used for climate reconstruction. In my opinion, there is much more to explore. While the laser is 'drilling' through the foraminiferal chamber wall (with an ablation rate of 0.2-0.3 µm/s), the Element/Ca ratios are measured in real time (with a few seconds delay, as the ablated material needs to pass through tubes and the spray chamber before being ionized in the plasma). Therefore, it is possible to compute elemental concentration profiles through foraminiferal chamber walls at micrometer resolution. These laser ablation profiles provide detailed insights in the degree of diagenesis and/or recrystallization, as demonstrated in previous studies (e.g. Hathorne et al., 2003; Regenberg et al., 2007; Pena et al., 2008; van Raden et al., 2011). In processing these LA depth-profiles, it should be possible to determine if diagenesis and/or alumina-silicate contamination affects only on the outer and inner surface of the shells, or if these 'unwanted' phases penetrate deeper into the chamber wall. These information can also be used to locate domains that are less affected by diagenesis than the remaining shell (likely the inner part of the chamber wall). Would it be possible to compile – for comparison – an alternative Mg/Ca record using exclusively data from less altered portions of the shell? These are just suggestion, but I think they are worthwhile to explore. There are certainly more information in the data set than shown/discussed by the authors. Is it also possible to show some representative laser ablation profiles in the supplementary material?

Response 1: The referee asks that we give greater consideration to the potential impacts of diagenesis on Mg/Ca, d18O and d13C. We have addressed this in our response to Referee 1, who made a similar general comment. It is important to note that the aim of the foraminiferal isotope analysis was to provide constraints on the bulk isotope and Mg/Ca data. Based on the relatively poor preservation of the material, we fully expected to find the data somewhat compromised by diagenesis. We recognise that there may well be scope for further detailed work on the isotopic character of the foraminifera in this record of the type that the referee has pioneered, but this is outside the aims of the current study.

*For LA-ICPMS analysis, we have improved the description of methods to clarify that we followed exactly the approach the referee recommends (**Lines 204-229**). After visual examination and cleaning, specimens were ablated. The LA depth profiles were then carefully screened to identify and exclude zones of contamination, diagenesis, or elemental anomalies of uncertain origin (**Fig. S1**). After this process, average values were determined for the selected segments on the profiles, not the entire profiles (**Data plotted in new Fig. 3**). Further screening was then performed on these average values to derive the Mg/Ca values used for temperature reconstructions (**New Fig. 4 and S2**). We agree that further study should be undertaken on the profiles to directly quantify diagenetic variation through this interval. However, this is outside the scope of the current study.*

Other comments:

- Hollis et al. analyzed nine element by LA-ICP MS (Mg, Al, Si, Ca, Ti, Mn, Zn, Sr, and Ba), however, only Al/Ca, Sr/Ca and Mg/Ca ratios were used to assess the degree of diagenesis. What about Mn/Ca and Ba/Ca? The authors themselves cite studies using Mn/Ca and Ba/Ca ratios to evaluate contamination or diagenesis of foraminiferal shells. As the mechanisms of diagenesis are very complex and site-specific, it may be possible that the data are not conclusive, however, results from other Element/Ca ratios should be at least briefly mentioned.

We could add Mn/Ca and Ba/Ca plots to Fig. 3 but they don't add much to the record as they show very similar trends to Al/Ca. We have included them in ***Fig. S2****. Further discussion of other elements is outside the scope of this paper. Note that we also have a full suite of XRF data for the PETM interval that is not discussed here.*

Furthermore, the authors emphasize a positive linear correlation between measured Al/Ca and Mg/Ca ratios. This linear correlation is difficult to identify from the data shown in Fig. 3. Therefore, I suggest to include an Al/Ca:Mg/Ca cross plot in the supplementary material - Page 259, lines 23-26:

We found the suggestion of adding cross-plots most helpful. Although they were part of our working files, we have not interrogated them sufficiently and are very pleased to see how well they appear to resolve questions relating to diagenesis (new Fig. 3, ***Lines 230-241, 503-527****)*

The observation that the Mg concentration is not being reset during shell recrystallisation is important and confirms in situ measurements of Mg/Ca ratios in diagenetic overgrowth that approach the values of biogenic foraminiferal calcite (Kozdon et al., 2013). However, these findings contradict the results of inorganic precipitation experiments showing about an order of magnitude more Mg in diagenetic than in biogenic calcite (e.g. Mucci and Morse, 1983; Oomori et al., 1987), and I encourage the authors to take this opportunity and emphasize the difference between diagenesis in the sediment column and the results from inorganic precipitation experiment. Some scientists still believe that non-elevated Mg/Ca ratios are an indicator for good preservation.

*We agree that our results are in good agreement with Kozdon et al. (2013) , with very similar relationships between Sr/Ca and Mg/Ca, and an order of magnitude lower than the changes observed in laboratory experiments (****Lines 515-517****, compare our* ***new Fig. 3*** *with Kozdon et al. 2013, Fig. 6)*

Page 259, line 22: it shall read "Fig. 4d".

Corrected

Fig. 4d: is it possible to add a horizontal axis for d18O and Mg/Ca?

We think this is unnecessary because Mg/Ca is shown in Fig. 4c and d18O in Fig. 2 but yes, it's possible (at least for d18O).

~~Onset of the~~The Paleocene-Eocene Thermal Maximum at DSDP Site 277, ~~in the southern Pacific Ocean (DSDP Site 277,~~ Campbell Plateau, southern Pacific Ocean~~)~~

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ABSTRACT

Re-examination of ~~a~~ sediment cores ~~collected by the~~from Deep Sea Drilling Project (DSDP) Site 277~~)~~ on the western margin of the Campbell Plateau~~, Southwest Pacific Ocean~~ (paleolatitude of ~65°S)~~,~~ has identified an intact Paleocene-Eocene (P-E) boundary overlain by a 34 cm-thick record of the ~~initial phase of the~~ Paleocene-Eocene Thermal Maximum (PETM) within nannofossil chalk. The upper part of the PETM is truncated, either due to drilling disturbance or a sedimentary hiatus. An intact record of the onset of the PETM is indicated by a gradual decrease in δ^{13}C values over 20 cm, followed by a 14 cm interval in which δ^{13}C is 2‰ lighter than uppermost Paleocene values. After accounting for effects of diagenetic alteration, we use δ^{18}O and Mg/Ca values from foraminiferal tests to determine that intermediate and surface waters warmed by ~5-6° at the onset of the PETM prior to the full development of the negative δ^{13}C excursion. After this initial warming, sea temperatures were relatively stable through the PETM, but declined abruptly across the ~~unconformity~~ horizon that truncates the event at this site. Mg/Ca analysis of foraminiferal tests indicate peak intermediate and surface water temperatures of ~19°C and ~32°C, respectively. These

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Formatted: Header distance from edge: 1.25 cm, Footer distance from edge: 1.25 cm temperatures may be influenced by residual diagenetic factors, changes in ocean circulation, ~~enhanced poleward ocean heat transport during the PETM~~ and surface water values may also be biased towards warm season temperatures.

1. Introduction

Stable isotope analysis of foraminiferal tests from sediments cored at DSDP Site 277 (Shackleton and Kennett, 1975) provided the first paleotemperature record for the Paleogene of the Southern Ocean and laid the foundation for many subsequent studies of the regional paleoclimate and paleoceanography (e.g., Kennett 1977, 1980; Kennett and Shackleton, 1976; Hornibrook, 1992; Nelson and Cook, 2001). Over the last decade, there has been renewed interest in the early Paleogene (66 to 35 Ma) climate history of the Southern Ocean, partly driven by a societal imperative to understand how the Antarctic ice sheet will respond to anthropogenic global warming (e.g., Joughin et al., 2014). The early Paleogene was the last time that Earth is inferred to have experienced greenhouse gas levels in excess of ~600 ppm CO_2 (Zachos et al., 2008; Beerling and Royer, 2011), and therefore provides insight into a climate state that civilization may experience in coming centuries. One event in particular has been touted as a geological analogue for greenhouse gas-driven global warming: the Paleocene-Eocene Thermal Maximum (PETM, ~56 Ma). This event was a short-lived (~220 kyrs) perturbation to the climate and carbon cycle in which global temperatures rose by 4–5°C within a few thousand years (Sluijs et al., 2007; McInerney and Wing, 2011; Dunkley-Jones et al., 2013; Schmidt, 2014), with warming of up to 8°C in higher latitudes and some coastal settings (Thomas et al., 2002; Sluijs et al., 2006, 2011; Zachos et al., 2006; Hollis et al., 2012; Frieling et al., 2014). Multiple lines of evidence suggest that this warming may have been driven by a rapid injection of greenhouse gases, possibly sourced from submarine gas hydrates, as evidenced by coupled negative excursions in oxygen and carbon isotopes (Dickens et al., 1995, 1997). Several other potential sources of the light carbon have also been implicated to account for all or part of the carbon isotope ($\delta^{13}C$) excursion (Dickens, 2003, 2011; Kent et al., 2003; Svensen et al., 2004; Higgins and Schrag, 2006; De Conto et al., 2012).

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The PETM has been identified in several sites in the Southwest Pacific, including onshore records in both siliciclastic and pelagic bathyal sections in eastern New Zealand (Kaiho et al., 1996; Crouch et al., 2001; Hancock et al., 2003; Hollis et al., 2005a, b, 2012; Nicolo et al., 2010), non-marine to marginal marine sediments in western New Zealand (Handley et al., 2011) and in shelfal sediments at Ocean Drilling Program (ODP) Site 1172, offshore eastern Tasmania (Sluijs et al., 2011). Here we report a new record of the PETM in pelagic bathyal sediments at DSDP Site 277, at a similar

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paleolatitude to Site 1172 (~65°S). These two sites represent the southernmost records of the PETM in the Pacific Ocean (Fig. 1).

Initial studies of Site 277 suggested that the Paleocene-Eocene (P-E) boundary occurred within a gap between cores 43 and 44 (Kennett et al., 1975). A subsequent biostratigraphic review of the site (Hollis et al., 1997) revealed that the boundary was lower in the drillhole, potentially within a relatively continuous interval preserved in core 45. Detailed re-sampling confirmed the location of the P-E boundary (Fig. 2), based on the highest occurrence (HO) of benthic foraminifer *Stensionina beccariformis* at 457.3 mbsf (277-45-3, 80 cm). High resolution stable isotope analysis of bulk carbonate confirms that this horizon marks the base of a 34 cm-thick negative excursion in δ^{13}C (CIE) that defines the PETM (~~Fig. 2~~Aubry et al., 2007).

DSDP Site 277 was drilled on the western margin of the Campbell Plateau in a water depth of 1214 m as part of DSDP Leg 29 (Kennett et al., 1975). Paleogene sedimentation occurred in fully oceanic conditions well above the lysocline (Kennett et al., 1975), with benthic foraminiferal assemblages indicating lower to middle bathyal water depths since the Paleocene (Hollis et al., 1997). In order to ~~characterise sedimentary and~~identify the paleoceanographic changes associated with the PETM at this site we have undertaken a multidisciplinary study that includes foraminiferal and calcareous nannofossil biostratigraphy, magnetic susceptibility, $CaCO_3$ content, elemental abundance using X-ray fluorescence (XRF), δ^{13}C and δ^{18}O analysis of bulk carbonate and foraminifera, and single test analysis of foraminifera for Mg/Ca ratios by Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICPMS).

2. Material and Methods

2.1. Material

We analysed samples over a 45-m interval spanning the upper Paleocene to lower Eocene at DSDP Site 277 (470–425 mbsf). Average sample spacing was 20 cm over much of the interval, with a higher resolution of 2–3 cm sampling across the PETM within core-section 45-3 (~457.30–456.95 mbsf). In addition, this core-section was scanned for elemental abundance. Although the PETM interval is preserved, the overall record is discontinuous, with significant gaps between cores from core 42 to 45 (Fig. 2).

2.2. Methods

2.2.1. X-Ray fluorescence (XRF) core scanning

XRF data were acquired using an Avaatech XRF scanner with a Canberra X-PIPS silicon drift detector, model SXD 15C-150-500 150 eV resolution X-ray detector, which is housed at the

International Ocean Discovery Program (IODP) Gulf Coast Repository at Texas A&M University in College Station, Texas (Table S1). This scanner is configured for analysis of split core section halves, with the X-ray tube and detector mounted on a moving track (Richter et al., 2006). Section 277-45-3 was removed from the core refrigerator and allowed to equilibrate to room temperature prior to analysis. We ~~leveled~~levelled all rock pieces within the section, as the detector requires a flush surface with no gaps between pieces, and then covered the section with 4 µm thick Ultralene plastic film (SPEX Centriprep, Inc.) to protect the detector. The section was scanned at 2 mm intervals using a voltage of 10 kV for elements Al, Si, P, S, Cl, Ar, K, Ca, Ti, Cr, Mn, Fe, Rh, and Ba. The scan was completed using a 1 mA tube current, no filter, and a detector live time of 30 s, with an X~~x~~-ray detection area of 2 mm in the downcore direction and 15 mm across the core. During measurement, intervals were skipped where gaps of more than ~2 mm existed between pieces. Smaller gaps were noted so that suspect data across these gaps could be removed.

2.2.2. Rock magnetism

Bulk magnetic susceptibility of a subset of discrete samples was measured at the Paleomagnetism Laboratory of the Complutense University of Madrid, Spain (Table S2). A KLY-4 (Agico) susceptibility bridge was employed, with an applied magnetic field of 300 A/m. Due to the low ferromagnetic content of most ~~of the~~ samples, each sample was measured ten times and averaged. The error bars of the magnetic susceptibility data correspond to the standard deviation of the mean (1σ) obtained during the averaging procedure.

2.2.3. Micropaleontology

Calcareous nannofossil and foraminifera sample preparation and examination followed standard procedures. Samples for calcareous nannofossils were prepared using standard smear-slide techniques (Bown and Young, 1998). A small amount of sediment was mixed with a drop of water on a coverslip, distributed with a toothpick, and then dried on a hot plate. The coverslip was affixed to a glass microscope slide using Norland Optical Adhesive 61 and cured under an ultraviolet light. Slides were examined on a Leitz Ortholux II POL-BK microscope under cross-polarized and plane-transmitted light~~s~~. Nannofossil distribution was determined for 41 samples extending from Paleocene to the upper lower Eocene (Teurian to Mangaorapan New Zealand stages) (Table S3). Counts of 400 specimens were conducted at 1000× for each sample, followed by a scan of at least 400 fields of view at 630× to look for rare taxa~~ (Table 3~~).

Foraminiferal distribution was determined for 59 samples ~~extending from the Paleocene to late early Eocene (Teurian to Mangaorapan local stages)~~spanning the same time interval (Table S4).

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Foraminiferal biostratigraphy is correlated with New Zealand stages (Cooper, 2004) and international biozones (Olsson et al., 1999; Pearson et al., 2006). New Zealand stage and biozone boundaries are calibrated to the 2012 geological timescale (Gradstein et al., 2012) using criteria described by ~~Hollis~~ Raine et al. (2015~~0~~) and Norris et al. (2014). Foraminiferal taxonomy is based on Hornibrook et al. (1989). Biostratigraphic results for calcareous nannofossils are correlated to the biostratigraphic zonation scheme of Martini (1970, 1971), calibrated to the 2012 geological timescale (Gradstein et al., 2012). Taxonomic concepts for species are those given in Perch-Nielsen (1985) and Bown (1998).

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2.2.4. Stable isotopes and carbonate content

Analysis for stable isotopes and carbonate content was undertaken at three laboratories. Results are tabulated in Table S5. Bulk carbonate δ^{13}C and δ^{18}O measurements were undertaken at the National Isotope Centre, GNS Science, Lower Hutt. Samples were analysed on the GVI IsoPrime Carbonate Preparation System at a reaction temperature of 25°C and run via dual inlet on the IsoPrime mass spectrometer. All results are reported with respect to VPDB, normalized to the GNS marble internal standard with reported values of 2.04‰ for δ^{13}C and -6.40‰ for δ^{18}O. The external precision (1σ) for these measurements is 0.05‰ for δ^{13}C and 0.12‰ for δ^{18}O.

Individual specimens from five foraminiferal genera were used for stable isotope analysis and elemental geochemistry. Specimens were selected for analysis based on visual assessment of their preservation under a stereo microscope. Wherever possible, analyses were performed on *Morozovella aequa, Acarinina coalingensis, Subbotina patagonica, S. roesnasensis*, and *Cibicides proprius/praemundus, and Stensioina beccariformis*. The following species were substituted when these species were not available: *Morozovella subbotinae, M. acuta, M. apanthesma, Acarinina soldadoensis, A subsphaerica, A. esnaensis, A. nitida and Cibicides tholus*. The stable isotope signature of *Acarinina soldadoensis, A subsphaerica, A. nitida* and all species of *Morozovella* indicates they were mixed layer dwellers (Olsson et al., 1999; Quillévéré and Norris, 2003), and therefore are appropriate indicators of near surface conditions. *Subbotina patagonica* is inferred to have had a deeper planktonic habitat (Pearson et al., 2006), within the thermocline. There is no data on the habitat *of S. roesnasensis*.~~Individual specimens from five foraminiferal genera were used for δ^{13}C and δ^{18}O and Mg/Ca analysis. Samples were selected based on light microscope assessment of preservation, which was subsequently confirmed by scanning electron microscope (SEM) examination. The genera used include two near-surface dwelling genera, *Morozovella* and *Acaranina*, the thermocline dweller *Subbotina* and two benthic genera, *Cibicides* and *Stensioina*.~~

Stable isotope analysis of foraminifera was carried out in the Stable Isotope Laboratory at the University of California, Santa Cruz. Between 1 and 6 (average of 3) specimens of *Cibicides*, 1 and 5 (average of 3) specimens of *Stensioina*, 3–17 (average of 10) specimens of *Acarinina*, 2–10 (average of 4) specimens of *Morozovella*, and 1–8 (average of 5) specimens of *Subbotina* were used in each analysis. Specimens were first sonicated in deionised water to remove clay and detrital calcite. Isotopic measurements were carried out on a Thermo-Finnigan MAT253 mass spectrometer interfaced with a Kiel Device. The analytical precision (1σ) is based on repeat analysis of an in-house standard (Carrara marble), calibrated to the international standards NBS18 and NBS19, and averages ±0.05 ‰ for $\delta^{13}C$ and ±0.08 ‰ for $\delta^{18}O$. All values are reported relative to VPDB. For the $\delta^{18}O$ values of *Cibicides* (= *Cibicidoides*; see Schweizer et al., 2009) and *Stensioina*, we apply an isotopic correction factor of +0.28 (Katz et al., 2003).

Paleotemperatures for both benthic and planktic taxa were calculated from $\delta^{18}O$ using the equation of Kim and O'Neil (1997):

$$T\ (^\circ C) = 16.1 + -4.64(\delta^{18}O_M - \delta^{18}O_{SW}) + 0.09(\delta^{18}O_M - \delta^{18}O_{SW})^2$$

Equation 1

Where $\delta^{18}O_M$ = measured value and $\delta^{18}O_{SW}$ = -1.23~~46~~‰, which is the inferred value for sea water under ice-free conditions (assuming ice-free ocean values of = -1‰). Planktic values are also corrected for ~~adjusted~~ for paleolatitude (Zachos et al. 1994; correction of -0.23‰)~~ and ice-free conditions (assuming ice-free ocean values of = -1‰)~~.

The carbonate content of dried powdered samples was determined at the National Institute of Water and Atmosphere (NIWA, Wellington) via gasometric quantitative analysis after acidification (Jones and Kaiteris, 1983), with a precision of ±2%.

2.2.5. Elemental geochemistry and Mg/Ca analysis

Foraminifera were picked from the 150–300 μm fraction of washed sediment samples and individually washed in ultra-pure (>18.2 mΩ) water and analytical grade methanol three times before being mounted on double-sided tape adhered to a glass slide. ~~Wherever possible, the species *Morozovella crater, Acarinina primitiva* and *Cibicides eocaenus* were selected. Where these species were not available, *M. lensiformis, A. collactea* and *C. truncatus* were substituted. *Subbotina* was not subdivided beyond the genus level.~~

Mg/Ca analysis was carried out on ~~samples composed of~~ 4–19 specimens for each of the selected genera in each sample (Table S6). Each foraminifer was analysed at least three times using a pulsed Ar-F excimer laser (Lambda Physik LPFpro 205) with a 193 μm wavelength, 30 μm spot size, laser power of 3 J/cm^2 and a repetition rate of 3 Hz, in conjunction with an ANU HelEx laser ablation cell,

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at the Research School of Earth Sciences of the Australian National University. An analysis of the NIST-SRM610 silicate standard was taken between every 9–12 foraminifer analyses to correct for elemental fractionation originating from laser ablation and mass-spectrometry effects.

The final three chambers of the final whorl in each specimen were analysed individually by ablating slowly at a rate of 0.2–0.3 µms^{-1} to produce a separate trace element profile through the wall of each chamber (Fig. S1). A Varian 820 ICPMS was used to measure abundances of the trace metal isotopes ^{24}Mg, ^{27}Al, ^{29}Si, ^{47}Ti, ^{55}Mn, ^{66}Zn, ^{88}Sr and ^{138}Ba relative to ^{43}Ca during ablation. Elemental ratios reported ~~in this study~~for each sample are ~~the~~ average values derived from~~of the~~ multiple screened ~~chamber profiles~~ profile segments ~~measured in individual~~for multiple ~~foraminifer~~ specimens of a given taxon. Laser ablation sites were selected using light microscopy and SEM imaging to avoid zones of detrital contamination, recrystallization or test ornamentation ~~which~~ that might cause irregular trace element/Ca profiles (Fig. S1).

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~~The extraction of a reliable paleotemperature record from variably preserved foraminifera is dependent on the preservation of primary foraminiferal calcite, and its inherent geochemical signature. In addition to using reflected light microscopy and SEM imaging to qualitatively screen tests for preservation, stringent~~ Individual chamber profiles ~~screening criteria~~ were ~~applied~~ screened to exclude zones with anomalously high Mg/Ca, Al/Ca, Mn/Ca or Ba/Ca ratios, which indicate significant ~~within test wall profiles that show evidence of diagenetic alteration or~~ silicate contamination ~~as evident from anomalous Mg/Ca, Al/Ca, Mn/Ca, Ba/Ca and Sr/Ca ratios~~ (Barker et al., 2003; Greaves et al., 2005; Creech et al., 2010). ~~Individual trace element~~These profiles typically show zones of enriched in ~~substantially~~ Mg, Al, Mn, and Ba ~~elevated Mg/Ca, Al/Ca and Mn/Ca ratios~~ on the outside and inside surfaces of the ~~foraminifera test~~chamber wall, ~~indicating siliciclastic sediment contamination~~consistent with silicate contamination (Fig. S1). ~~, whereas~~ The Sr/Ca ratio is used as an indicator of diagenetic alteration because the concentration of Sr may decrease or increase during alteration or secondary calcification (Eggins et al., 2003; Kozdon et al., 2013). A ratio of ~1.4 is typical for well-preserved tests (Creech et al., 2010). Therefore, ~~typically occurs at uniform levels throughout the test. Concentration may decrease or increase during alteration or secondary calcification (Eggins et al., 2003). Thus,~~ samples with Sr/Ca values outside the range of 0.8–1.6 mmol/mol were considered to be affected by diagenesis (Fig. 3~~3~~).

Al/Ca and Mg/Ca data show a positive linear correlation when plotted (Fig. 3), reflecting the influence of ~~alumino-~~silicate ~~mineral~~ contamination. ~~This~~ We have used ~~contamination has been screened out using~~ the method of Creech (2010~~)~~; after Barker et al., ~~(~~2003) to screen for this contamination. The Al/Mg composition of the contaminant phase was identified by plotting Mg/Ca against Al/Ca and finding the slope of the linear regression. Once this ~~Mg/Al~~Al/Mg composition had been determined for each genus, the screening threshold was set by calculating the Al/Ca ratio at which paleotemperature estimates would be biased by more than 1 °C. This screening removes anomalously high Mg/Ca values and reduces the mean value for most samples (Fig. 4, S2). After the measurements have been screened for silicate contamination, the effects of diagenesis are more easily assessed (Fig. 3). A weak negative correlation between Sr/Ca and Mg/Ca suggests that diagenesis may also cause an increase in Mg/Ca values, especially in the planktic genus *Acarinina*. The reasons for this correlation and implications are discussed below.

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~~This involved determining the corresponding Mg/Ca excess value by multiplying the observed mean Mg/Ca ratio for each species by the sensitivity of the paleotemperature calibration (i.e. an increase of 9% in Mg/Ca per 1 °C after Anand et al., 2003). Once determined the Mg/Ca excess value for each species is multiplied by the Al/Mg contaminant phase to determine the Al/Ca ratio at which the screening threshold is set. This methodology circumvents the inclusion of artificially inflated paleotemperature estimates arising from silicate contamination (Fig. 3). The resulting reduction in average Mg/Ca ratios of the screened compared to the unscreened and data is shown in Figure 3.~~

Comment [CH1]: Method is described in Creech et al. (2010).

~~The application of these screening limits to trace element data means that no Mg/Ca marine temperatures in this study should be overestimated by more than 1°C (as a result of post-mortem diagenetic effects or sedimentary contamination).~~

Marine paleotemperatures are calculated using the exponential relationship between Mg/Ca and temperature (*Equation 2*). Because the planktic foraminifera used in this study are extinct, sea surface temperatures (SSTs) were calculated using a general calibration based on the mean calcification temperatures of nine modern planktic species (A = 0.09, B = 0.38; Anand et al., 2003). Sea floor temperatures (SFTs) were calculated using the calibration of Lear et al. (2002) based on three benthic species of *Cibicidoides*/*Cibicides* ~~species~~ (A = 0.109, B = 0.867).

$$Mg/Ca_{test} = \left(\frac{Mg/Ca_{sw}^{t=t}}{Mg/Ca_{sw}^{t=1}}\right) \times Bexp^{AT}$$

Equation 2

Marine temperature reconstructions based on early Eocene foraminiferal calcite have shown that a high (>3 mol/mol) Mg/Ca_{sw} value is ~~necessary~~ required to reconcile Mg/Ca-derived paleotemperatures with those derived from $\delta^{18}O$ (Lear et al., 2002; Sexton et al., 2006). ~~Such h~~High Mg/Ca_{sw} values are in line with modelled values from Wilkinson & Algeo (1989) but are at odds with several proxy studies (e.g., Horita et al., 2004; Coggon et al., 2010) and ~~models~~ more recent modelling (e.g., Stanley & Hardie, 1998) that favour lower values for Mg/Ca_{SW} (<2 mol/mol).~~, but are in line with modelled values from Wilkinson & Algeo (1989), and multi-proxy paleotemperature~~

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~~reconstructions (Hollis et al., 2012). Lower values of Paleogene Mg/Ca$_{sw}$ (e.g., Stanley & Hardie, 1998; Coggon et al., 2010) result in unrealistically high temperatures using *Equation 2*.~~ However, recent studies (Hasuik & Lohmann, 2010, Evans & Müller, 2012) have ~~shown~~ reconciled the empirical relationship between δ^{18}O and Mg/Ca paleotemperatures with these lower values for Mg/Ca$_{SW}$ by showing that a power law ~~relationship~~distribution, rather than an exponential ~~relationship~~distribution, ~~may~~ better describes the relationship between Mg-partitioning and temperature in foraminiferal calcite ~~-~~(*Equation 3*).

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$$Mg/Ca_{test} = \left(\frac{B}{Mg/Ca_{sw}^{t=0^{H}}}\right) \times Mg/Ca_{sw}^{t=t^{H}} exp^{AT}$$

Equation 3

~~In applying this method, it is possible to derive realistic paleotemperatures using a lower Eocene Mg/Ca$_{sw}$ value that is consistent with Mg/Ca$_{sw}$ proxy evidence.~~ To apply this equation we use ~~Exponential~~ exponential and pre-exponential calibration constants from modern multispecies calibrations and paleotemperature values derived from oxygen isotopes ~~can be utilised~~ to estimate the function *H* for extinct foraminifera. ~~Using p~~Published data from well-preserved Eocene foraminifera at Hampden Beach (Burgess et al., 2008; Hollis et al., 2012) and Tanzania (Pearson et al., 2007), for which paired Mg/Ca and δ^{18}O data is available, ~~it is possible~~have been used to derive ~~the calibration correction constants (~~*H*~~)~~ for the extinct species used in this study.

In calculating the value of *H*, we have used an early Eocene Mg/Ca$_{sw}$ value of 1.6 mol/mol (Stanley & Hardie, 1998; Evans & Müller, 2012) and a modern Mg/Ca$_{sw}$ value of 5.17 mol/mol. This~~e~~ ~~values of~~ *H* ~~determined for Paleogene foraminifera in this study are an approximation that do~~value does not take into account ~~the likely fine scale~~possible variability in Mg/Ca$_{sw}$ values through the early Paleogene. The Mg/Ca-temperature calibrations of Anand et al. (2003) and Lear et al. (2002) have been used, although it is likely that the pre-exponential constant of Paleogene planktic foraminifera differed from that of the modern taxa. ~~The~~ We calculate an H value~~s of *H* calculated~~ of 20 for Paleogene planktic foraminifera, which is significantly ~~are~~ lower ~~(*H* = 0.20)~~ than ~~that for the~~H values for modern ~~taxon~~planktics, such as *Globigerina sacculifer* (*H* = 0.42; Hasuik & Lohmann, 2010)~~, possibly due to differences in Mg/Ca-temperature calibration~~. For benthic foraminifera, Cramer et al. (2011) suggest that the value of *H* would be similar between *Cibicides* sp. and *Oridorsalis umbonatus*. Mg/Ca-derived temperature values are calculated using *Equation 4*.

Comment [CH2]: What is it??

$$T = \frac{\ln\left(\frac{[Mg/Ca_{test}] \times [Mg/Ca_{sw}^{t=0}]^{H}}{B \times [Mg/Ca_{sw}^{t=t}]^{H}}\right)}{A}$$

Equation 4

Temperature values derived from Mg/Ca ratios of surface mixed-layer dwelling taxa used in this study are normalised to *Morozovella crater* following Creech et al. (2010).

Three types of error are applied to paleotemperatures derived from Mg/Ca ratios; the analytical error, sample error and a standard calibration error. The analytical error is accounted for in the data processing step, and typically produces very small uncertainties (± 1–3% 2se) associated with counting statistics during ablation and data acquisition. The sample error pertains to the 95% confidence interval calculated for the mean temperature value obtained from multiple analyses within a single sample, and is calculated by:

$$\bar{x} \pm t \times \frac{\sigma}{\sqrt{n}}$$

Equation 5

Where $\bar{x}$ is the sample mean, *t* is the inverse of the Students' t-distribution, *σ* represents the standard deviation and *n* is the number of analyses. The calibration error is the residual error of ±1.6°C on the regression of the multispecies calibrations established by Lear et al. (2002) and Anand et al. (2003). The cumulative error calculated from the sum of all three errors is applied to each temperature value, providing upper and lower uncertainties.

3. Results and Discussion

3.1 Stratigraphy

The 45 m-thick studied interval (425–470 mbsf) consists of five cores, with significant gaps due to poor recovery in three of the cores, which extend from middle Paleocene to lower Eocene (Fig. 2). The sediments are greenish-white to greenish-grey nannofossil chalk, with higher clay content in the upper Paleocene (core 46; 463–470 mbsf) and lowermost Eocene (core section 45-3; 456.96–457.3 mbsf)) and minor glauconite (cores 43-44) and chert nodules (cores 41-43) in the overlying Eocene. A record of "incipient chert" in core section 45–3 (Kennett et al., 1975~~3~~) may have been a misidentification of the darker-grey clay-rich sediments at the base of the PETM (Fig. 5~~4~~).

Calcareous microfossils are only moderately ~~well~~ preserved overall, ~~although~~ and there is an interval directly below the Paleocene-Eocene boundary (457.3 to 457.58 mbsf) in which foraminifera are poorly preserved and sparse. Planktic foraminifera are used to correlate the 45 m-thick studied interval to New Zealand stages (Teurian to Mangaorapan) and to international foraminiferal zones P4a-b to E7 (Fig. 2). Nannofossil assemblages over the same interval have been correlated with nannofossil zones NP6 to NP12. Whereas previous studies indicated an undifferentiated upper Paleocene succession spanning Zone NP6–8 (Edwards and Perch-Nielsen, 1975; Hollis et al., 1997), we infer a ~2 Myr hiatus near the top of Core 46 (463.49–463.16 mbsf), representing all of zones

NP7 and NP8. Immediately above the hiatus, *Discoaster multiradiatus* makes up ~2% of the assemblage, suggesting that ~~the first appearance datum (FAD) of this zonal marker (and thus~~ the ~~base of~~lowermost part of Zone NP9~~)~~ is missing. This ~~horizon~~ lowest occurrence (LO) of D. multiradiatus ~~marks the lowest occurrence (LO) of *Discoaster* at this site, including~~coincides with the LOs of *D. lenticularis* and *D. salisburgensis* ~~in addition to *D. multiradiatus*~~.

The PETM is a 34 cm-thick interval within core 45 (457.3–456.96 mbsf) that is clearly delineated by a 40% decrease in carbonate content and 2–3‰ negative excursions in bulk carbonate δ^{13}C and δ^{18}O values (Fig. 2). ~~The location of the PETM is confirmed by biostratigraphy.~~ The B~~b~~enthic F~~f~~oraminiferal E~~e~~xtinction E~~e~~vent (BFEE) is identified directly below the PETM at 457.3 mbsf based on the highest occurrences of the *Stensioina beccariformis*, *Gyroidinoides globosus* and *G. subangulatus*. The planktic foraminiferal genus *Morozovella* has its lowest occurrence at the base of the PETM and greatest diversity within the PETM. *Morozovella aequa* and *M. velascoensis* are restricted to the PETM. The latter species has rarely been found outside the PETM in the SW Pacific but *M. aequa* ranges into the middle late Eocene in New Zealand sections (Hornibrook et al. 1989). For nannofossils, taxa typical of the PETM in other regions, such as the *Rhomboaster* lineage, *Discoaster araneus* and *D. anartios* (e.g., Bybell and Self-Trail, 1994; Kahn and Aubry, 2004), ~~do not occur at Site 277~~are absent here. Instead, the nannofossil assemblage is characterized by deformed *Discoaster* specimens, many similar to *Discoaster nobilis* (e.g., Raffi and De Bernardi, 2008), as well as increased abundance of *Coccolithus* spp. and the presence of *Fasciculithus* spp. and *Bomolithus supremus*, which is restricted to the PETM interval at this site. Immediately above the PETM (456.92 mbsf), the abundances of *Fasciculithus* spp. and *Coccolithus* spp. decrease significantly, with a concomitant increase in *Zygrhablithus bijugatus*. As discussed below, the stable isotope record through the P-E transition indicates that the PETM is truncated, with only the onset and body of the CIE represented by these 34 cm of sediment.

An age-depth plot (Fig. S3) based on calcareous nannofossil and foraminiferal bioevents (Table S7) provides a preliminary guide to compacted sedimentation rates. This rate appears to have been relatively low in the Paleocene (0.4 to 0.45 cm/kyr) either side of the hiatus at ~463.4 mbsf, but approximately four times higher in the early Eocene (1.68 cm/kyr). However, a rather patchy distribution of events and uncertainty over the duration of hiatuses means that it is possible to construct an alternative age model in which rates were consistent across the Paleocene-Eocene transition (dashed line in Fig. S3). Although this implies that the sedimentation rate for the PETM interval could lie anywhere between the low Paleocene rate and the high Eocene rate, the lower rate is consistent with the duration of the CIE from onset to δ^{13}C minimum, i.e., ~45-66 kyrs (Röhl et al., 2007).

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The base of the PETM coincides with a distinct colour change to a darker greenish-grey chalk that grades back into greenish-white chalk over 15 cm (Fig. 5~~4~~). This dark interval is also highly burrowed. Burrowing is also evident in other parts of the core but it is less obvious in more pale lithologies. XRF core scanning shows an increase in Fe content at the base of this interval, followed by a cyclical decrease to background levels at 456.95 m (Fig. 5~~4~~A). A ~~L~~lower ~~stratigraphic~~ resolution ~~analysis~~ record of magnetic susceptibility in discrete samples reveals a similar trend: a peak near the base of the darker interval, followed by a quasi-cyclical decrease to background levels. The peaks are inferred to represent intervals of higher clay content based on the parallel trends ~~An increase in clay content is inferred from the coupled increases~~ in Fe and magnetic susceptibility ~~and the decrease in carbonate content~~. Many of the other peaks and troughs in the Fe record below and above the PETM are scanning artefacts related to core breaks. However, parallel peaks in magnetic susceptibility and Fe content in the lower part of core 45 (~457.7 mbsf) appears to be a robust signal although the cause is unknown. There are no accompanying changes in isotopic signature or obvious lithological changes at this level.

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A 10 cm interval directly below the PETM also has a reduced carbonate concentration but there is no change in δ^{13}C (Fig. 2, 5B)~~and only a small positive shift in δ^{18}O of ~0.4‰ in both bulk and foraminiferal calcite (Fig. 2)~~. As there is no accompanying increase in magnetic susceptibility or Fe content (Fig. 5A), the decrease in carbonate content seems to be due to an increase in silica, perhaps associated with the slight cooling indicated by a small positive shift in δ^{18}O of ~0.4‰ in both bulk and foraminiferal calcite (Fig. 2, 5D)~~the positive shift in δ^{18}O~~. Although the silica is presumed to be biogenic, siliceous microfossils have not been recovered from this interval. ~~The cause of a peak in magnetic susceptibility and Fe content in the lower part of core 45 is unknown. There are no accompanying changes in isotopic signature and no obvious lithological changes.~~

~~An age-depth plot (Fig. 5) based on calcareous nannofossil and foraminiferal bioevents (Table 7) provides a preliminary guide to compacted sedimentation rates. This rate appears to have been relatively low in the Paleocene (0.4 to 0.45 cm/kyr), either side of the hiatus at ~463.4 mbsf, but approximately four times higher in the early Eocene (1.68 cm/kyr). However, a rather patchy distribution of events and uncertainty over the presence and duration of hiatuses means that it is possible to construct an alternative age model in which rates were consistent across the Paleocene-Eocene transition (dashed line in Fig. 5). Although this implies that the sedimentation rate for the PETM interval could lie anywhere between the low Paleocene rate and the high Eocene rate, the lower rate is consistent with the duration of the onset known from other sites, i.e., 50–100 kyrs (Zachos et al., 2008, 2010; Nicolo et al., 2010; McInerney and Wing, 2011).~~

For a sedimentation rate of 0.45 cm/kyr, the 34 cm thick PETM interval represents ~~~85~~ 76 kyrs and the three peaks in Fe content represent a periodicity close to the precession band (~~19–22~~~21 kyrs). Indeed, there is good agreement between the Fe cycles and δ^{13}C record at Site 277 and ODP Site 690 (Röhl et al., 2007), where the negative CIE occurs over three steps and the δ^{13}C minimum (Horizon C of Zachos et al., 2005) occurs within the third Fe peak. Based on this correlation with Site 690, we infer that the interval from the CIE onset to the base of Cycle 4 is preserved at Site 277, or the first 66 kyrs of the PETM (Röhl et al., 2007), implying a slight increase in sedimentation rate through the PETM (52 cm/kyr).

3.2. Stable isotopes~~ and paleotemperatures~~

Bulk carbonate stable isotopes display a significant offset between δ^{18}O and δ^{13}C minima, with the δ^{18}O minimum occurring at the base ~~of the PETM ~~and the δ^{13}C minimum ~~occurring at the top following a gradual decline~~in the upper part of the PETM (Fig. 2, 5~~4~~B). The negative ~~δ^{13}C excursion~~CIE of ~2‰ is slightly smaller than ~~observed in most marine sections~~the average for marine sections ~~elsewhere ~~(~~e.g., Nicolo et al., 2010; ~~ 2.7‰; McInerney and Wing, 2011) and occurs gradually over the ~~entire recovered~~lower 20 cm of the PETM~~ record~~. In contrast, the 3‰ negative δ^{18}O excursion (~~δ^{18}O excursion~~OIE) is abrupt at the base of the PETM and is ~~far ~~larger in magnitude than is known elsewhere (e.g., ~~Zachos et al., 2008; McInerney and Wing, 2011~~Bains et al. 1999; Dunkley Jones et al., 2013). If this a primary feature and due solely to a change in temperature, this ~~3‰~~ excursion would equate to ~12°C of warming (Fig. 5~~4~~D); however, it is most likely an ~~artifact~~artefact of diagenesis as is discussed below.

Examination of foraminiferal δ^{18}O ~~values ~~and Mg/Ca ratios ~~within three genera of planktic foraminifera and one benthic genus provide insight into the nature of the diagenetic process~~help to separate the diagenetic effects from the paleotemperature record. As none of the foraminifera recovered in this study have “glassy” preservation (Sexton et al. 2006; Pearson and Burgess, 2008; Kozdon et al., 2013), all are assumed to have ~~We suspect all shells have ~~been altered to varying degrees. ~~Based on visual and geochemical criteria, we were able to identify relatively well preserved tests.~~We selected the best preserved specimens for ~~I~~isotopic analysis~~ of those tests ~~ (Fig. 2, 5B, 6). Our results indicate that ~~show that their ~~normal surface to deep δ^{13}C gradients are preserved in the foraminiferal tests, with bulk carbonate δ^{13}C values lying~~ie~~ within the range of, or slightly lighter than,~~ ~~planktic foraminiferal δ^{13}C throughout the studied interval~~, with the exception~~. An exception is noted in ~~ of ~~the ~~base ~~basal ~~of the CIE~~PETM where two values are more positive than planktic δ^{13}C (Fig. 5B, 6~~6~~B). Benthic δ^{13}C values are >0.7‰ lighter than both planktic and bulk carbonate values, apart from the basal PETM sample where a negative gradient of -0.37‰ occurs between *Acarinina*

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and *Cibicides* (Fig. 5B, 6B). The implication is that the onset of the CIE is recorded more strongly in planktic foraminifera (i.e. surface water CIE of -1.85‰) than in either benthic foraminifera (deep water CIE of -0.55‰) or bulk carbonate (CIE of -0.34‰ across equivalent sample interval).

If it were not for the large magnitude of the OIE across the same sample interval (-1.42‰, and -2.82‰ for the full OIE), we might argue for mixing across the boundary dampening the bulk carbonate CIE. However, the marked differences in the pattern of onset for the CIE and OIE suggest that there was no mixing of sediment across the boundary. Similarly, there is little evidence for the isotope record being affected by carbonate dissolution or burn-down (Dickens, 2000; Kozdon et al, 2013) below the base of the PETM. A weak positive shift in pre-PETM δ^{18}O values and reduced carbonate content appears to reflect cooler conditions as the shift is accompanied by a cooling trend in the benthic Mg/Ca ratio (Fig. 5D-E).

A similar offset between bulk and planktic δ^{13}C in the basal PETM was described for ODP Site 690, where Stoll (2005) showed close agreement between trends in stable isotopes for bulk carbonate, coccolith fractions and *Subbotina* but significant offsets with *Acarinina*, the latter recording an earlier CIE onset and a later OIE minimum. Stoll (2005) considered several possible causes for this offset and favoured differences in habitat and seasonal production. For Site 690, the correspondence between coccoliths and *Subbotina* suggests that coccolith production may have occurred at a lower level within the photic zone than the level preferred by *Acarinina*. For Site 277, the δ^{13}C gradient suggests a similar explanation but a different relationship. During the PETM onset, coccolith production appears to have occurred at a shallower level than that preferred by planktic foraminifera at this site. This may also explain why bulk carbonate δ^{18}O is more depleted than planktic values in this interval, i.e. coccolith production in shallower and warmer waters. Given that this relationship is only fully expressed at the PETM onset, we suggest that this might have been a time of increased stratification and differentiation between water masses in the upper water column at this site. Nunes and Norris (2006) used ageing gradients in benthic δ^{13}C to infer a switch in deep water formation across the P/E boundary from the Southern Ocean to the Northern Hemisphere. Our benthic δ^{13}C data from Site 277 support this hypothesis. Site 277 benthic d13C is 0.46‰ higher than values in the equatorial Pacific prior to the PETM but 0.12‰ lower within the PETM. It seems likely that comparable changes occurred in surface water circulation.

With the CIE onset seeming explicable in terms of relationships between coccolith and foraminiferal niches and changes in ocean circulation, we turn our attention to the stepped decline in the bulk carbonate CIE at Site 277. Stoll (2005) argued that a similar series of three steps in the bulk carbonate CIE seen at Site 690 reflect the greater capacity for coccoliths to record changes in ocean conditions at a finer scale than is possible from the less abundant foraminiferal fraction.

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Although we lack the resolution in the foraminiferal record to compare sites 277 and 690 in detail, we observe the same trend and note a broad correlation with the three Fe peaks. It seems likely that these steps represent precessional modulation of the release of ^{13}C-depleted carbon into the ocean over ~60 kyrs (Röhl et al., 2007; Sluijs et al., 2007).

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3.3 Diagenetic modification of δ^{18}O values

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~~In contrast~~As noted above, the ~~the~~ bulk ~~-~~carbonate δ^{18}O values at Site 277 ~~lie within the range of~~intergrade~~-~~ between benthic and planktic foraminiferal ~~δ^{18}O~~ values in the Paleocene and in the Eocene interval above the PETM (Fig. 6~~6~~A, 6C). Moreover, several planktic foraminiferal δ^{18}O values are ~~is~~ only ~~slightly~~ ~0.3‰ lighter than benthic ~~and bulk δ^{18}O~~values in the Paleocene (Fig. 2, 6~~6~~C). ~~In contrast~~Conversely, all bulk carbonate δ^{18}O values lie within the range of planktic foraminiferal δ^{18}O within the PETM (Fig. 6~~6~~B) and, indeed, bulk carbonate δ^{18}O is lighter than planktic foraminiferal δ^{18}O in the basal PETM (Fig. 2, 5D).

We ~~believe~~ contend that diagenetic effects explain these relationships. The bulk carbonate δ^{18}O has been shifted toward heavier values during early diagenesis (at seafloor temperature) over much of the section above and below the CIE (Schrag et al., 1995; Sexton et al., 1996; Kozdon et al, 2013), whereas within the PETM interval the bulk and foraminiferal carbonate ~~has~~ appears to have undergone ~~little~~ less diagenetic alteration. We ~~suspect~~ suggest that the increase in clay ~~content, as inferred from higher iron content (Fig. 4),~~in the PETM protected coccoliths and foraminifera ~~slowed from wholesale~~~~carbonate~~ recrystallization, ~~thus~~ preserving more of the original δ^{18}O signal ~~in the coccoliths and foraminifera within the PETM~~. The presence of ~~Clay~~ clay ~~content~~ serves to reduce~~s~~ sediment porosity and retard~~s~~ carbonate recrystallization~~, almost completely in pure clays~~ (~~e.g., Pearson et al., 2007; Burgess et al., 2008~~Sexton et al., 2006~~; Hollis et al., 2012~~). This explains the large~~r~~ magnitude of the bulk carbonate δ^{18}O excursion across the P-E boundary, with the δ^{18}O values below the excursion having been altered toward heavier values (Fig. 2, 5~~4~~D-E).

~~Similarly, t~~The ~~planktonic~~planktic foraminiferal δ^{18}O values ~~through the Paleocene-Eocene transition~~ at ~~DSDP~~ Site 277 ~~are~~ appear to be compromised to varying degrees by seafloor diagenesis throughout the interval studies. ~~Although~~ Th~~thee~~e surface~~-~~-to~~-~~-deep temperature gradient may be expected to be reduced in high latitude regions such as the Campbell Plateau. Mean annual Subantarctic Water is ~6° warmer than Antarctic Intermediate Water in the present-day Southern Ocean (Carter et al., 1999). However, ~~, the~~ the very ~~extremely small~~low planktic-benthic δ^{18}O ~~offset~~ gradient in the Paleocene and post-PETM Eocene (0.8‰, 3°C) suggests alteration of planktic δ^{18}O toward benthic values (Fig. 2, 5D, 6C). The gradient is only slightly higher in the PETM (1.1‰, 4°C), suggesting that a cool bias affects all paleotemperatures derived from planktic δ^{18}O through the P-E

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transition at this site. The degree of this bias is uncertain. The warmer paleot~~Moreover,~~ ~~t~~emperature ~~estimates~~ derived from Mg/Ca ratios may be more reliable~~support our inference of a cool bias in planktic δ^{18}O due to seafloor diagenesis.~~ but, as is discussed below, diagenesis may result in a warm bias.

3.4. Diagenetic modification of Mg/Ca ratios

~~As for the impact of on Mg/Ca, there is limited opportunity for the Mg concentrations to be reset during diagenetic test recrystallization, in part because the bulk of Mg of the sediment/pore water system resides in the carbonate, whereas for oxygen isotopes, re-equilibration with ample supply of oxygen (in pore water) is possible.~~ There is evidence that diagenesis also has significant and specific effects on Mg/Ca values (Oomori et al., 1987; Kozdon et al., 2013). As noted earlier, we observe a distinct relationship between the Mg/Ca ratio and the geochemical proxy for diagenesis, the Sr/Ca ratio, once we have screened for silicate contamination (Fig. 3). For *Cibicides*, the full screened dataset shows a roughly horizontal trend, with little change in Sr/Ca as Mg/Ca varies. This suggests that this genus is relatively immune to the effects of diagenesis, perhaps related to its relatively thick and smooth wall. However, if we consider Paleocene and PETM samples separately, we observe that Paleocene analyses tend to have lower Sr/Ca ratios than PETM samples and exhibit a weak trend in which Mg/Ca increases as Sr/Ca decreases. This general relationship has also been identified by Kozdon et al. (2013) as a guide to diagenetic alteration, albeit the impact on Mg/Ca ratios is an order of magnitude smaller than found in laboratory experiments (Oomori et al., 1987). The trend is more obvious in *Acarinina* at Site 277, probably because the thinner-walled and more irregular test provides more surfaces for interaction with pore waters and hence facilitates diagenetic alteration. For the full data set, a significant negative correlation is observed, with Mg/Ca increasing as Sr/Ca decreases. A weaker trend is evident in the PETM data but a much stronger trend is shown by the Paleocene data. From these observations we can draw the following conclusions: (i) *Acarinina* is more prone to diagenesis than *Cibicides*, (ii) diagenesis is greater in the Paleocene than in the PETM, and (iii) diagenesis causes an increase in the Mg/Ca ratio and implies that paleotemperatures may be overestimated for some taxa, such as *Acarinina*, and in some intervals such as the Paleocene at this site. This may explain why the SST estimates for the Paleocene based on *Acarinina* Mg/Ca ratios are higher than expected (Fig. 5D, E).

3.5. Paleotemperature

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~~Consistent with the expected preservation biases, SST estimates of 26–27°C derived from the Mg/Ca ratios of *Acarinina* are markedly warmer than the temperatures of 14–17°C derived from planktic δ^{18}O.~~

~~Benthic foraminifera test are dense and thus less prone to recrystalization.~~ Taking into account these numerous complications, we can make some general observations on temperature changes through the P-E transition at Site 277. Estimates for SFT from benthic foraminiferal δ^{18}O and Mg/Ca are ~~remarkably~~ relatively consistent at 12–15~~3~~°C for the late Paleocene (Fig. 5~~4~~D), with coolest SFTs of 11–12°C occurring in the uppermost 10 cm of Paleocene where carbonate content is also lower than background. Benthic δ^{18}O and Mg/Ca values indicate SFT warmed by ~5–6°C across the P-E boundary, with SFTs of up to ~19°C in the basal PETM. There is little evidence for further warming of SFT in the body of the PETM. Following the PETM, SFT drops abruptly by ~5°C and remains stable at ~13°C in the overlying Eocene interval (Fig. 5D–E). Diagenesis may explain why some Paleocene Mg/Ca ratios yield higher SFTs than the benthic δ^{18}O values.

The SST record across the P-E boundary is much more difficult to interpret. The small offset between benthic and planktic δ^{18}O (~3°C) combined with the large offset between planktic δ^{18}O and Mg/Ca values (~12°C) in the Paleocene, are likely consequences of diagenetic alteration, with the actual SST lying somewhere between 15 and 27°C (Fig. 5D). Similarly, the degree of warming across the PETM may be accentuated for δ^{18}O but effectively dampened for Mg/Ca due to the effects discussed above. For this reason, the observation that the relative SST increase is ~5–6°C for both proxies (Fig. 5E) is difficult to explain even though it is consistent with the SFT record. Diagenetic effects appear to decrease across the P-E boundary, based on our analysis of the benthic–planktic δ^{18}O gradient and the Sr/Ca ratio. Therefore, we would predict that the relative increase in SST across the boundary would be greater for planktic δ^{18}O than for Mg/Ca. It may be that the patchy nature of the record through this interval is masking these relationships. Irrespective of the true magnitude of SST change across the P-E boundary, planktic Mg/Ca ratios indicate warmest SSTs in the lower PETM, stable SSTs through the body of the PETM (albeit ~3°C cooler) and an abrupt ~4°C cooling directly above the PETM.

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~~This~~The 5-6°C increase in SST is similar to other PETM records. At ODP Site 1172, the TEX$_{86}$ record indicates that SST increased by 6°C across the P/E boundary (Sluijs et al., 2011) and SST during the PETM was 3–4°C warmer than average Paleocene values (Fig. ~~7~~7). Elsewhere, temperature anomalies within the PETM range from +4–5°C in low latitudes (Zachos et al., 2003; Aze et al., 2014) to +8°C in high latitudes (Thomas et al., 2002; Frieling et al., 2014) and some low latitude coastal sites (Zachos et al., 2006).

~~planktic foraminiferal δ^{18}O values in the PETM yield SSTs of 26°C, whereas Mg/Ca ratios within the PETM yield much warmer SST estimates of 28–32°C (Fig. 4D).~~ The peak SSTs of ~32° within the PETM are consistent with TEX$_{86}$-based SSTs from the PETM at ODP Site 1172 (Sluijs et al., 2011) and in the mid-Waipara section, eastern South Island, New Zealand (Hollis et al, 2012). At these locations, the two calibrations for TEX$_{86}$ introduced by Kim et al. (2010) yield peak SSTs for the PETM of 32–34°C (TEX$_{86}^{H}$) or 26–28°C (TEX$_{86}^{L}$). Although the TEX$_{86}^{L}$ calibration was considered more suitable for this region based on comparisons with other SST proxies (Hollis et al., 2012), a new Bayesian approach to TEX$_{86}$ calibrations (Tierney and Tingley, 2014) yields temperatures for the PETM that are very similar to the TEX$_{86}^{H}$ calibration. These PETM SSTs are also consistent with the SST estimates of 26°C that were derived from TEX$_{86}$ and U$^{K'}_{37}$ for the late Eocene at Site 277 (Liu et al, 2009), given that deep sea temperatures cooled by ~8°C through the Eocene (Zachos et al., 2008).

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There is considerable debate about the veracity of such high temperature estimates in high latitude regions, with concerns raised about calibrations, seasonal bias and archaeol physiology ~~ical~~ processes (~~e.g.~~ Hollis et al., 2012; Taylor et al. 2013; Inglis et al., ~~submitted~~ under revision.). However, the consistency between SSTs derived from Mg/Ca and TEX$_{86}$ (Burgess et al., 2008; Hollis et al., 2012) suggests that the high temperatures are due to factors that the proxies may have in common, such as a warm-season bias, rather than problems with respective calibrations or physiological factors.

3.6~~3~~. ~~Pattern of PETM initiation~~Comparison with other PETM records

~~Complications relating to diagenetic overprinting have been discussed and partly resolved in the previous section. To circumvent additional uncertainties surrounding absolute temperature estimates based on δ^{18}O values and Mg/Ca ratios (Cramer et al., 2011; Evans and Müller, 2012; Dunkley-Jones et al., 2013), it is helpful to consider temperature trends relative to mean values for the Paleocene (Fig. 4E). For both proxies, SFTs increase by 5–6°C at the onset of the PETM, remain relatively stable through the PETM and decline abruptly across the unconformity that truncates the top of the event. Because the relative temperature trends shown by planktic δ^{18}O are inferred to be compromised by diagenesis, Mg/Ca ratios provide a more reliable guide to changes in local SST. In the basal part of the PETM, there are few planktic records that fall below the screening limit for Al/Ca (Fig. 3). Those specimens of *Acarinina* that are judged to be reliable indicate initial warming of SSTs by ~6°C. For the upper part of the PETM, still likely representing the onset phase, Mg/Ca ratios for both *Acarinina* and *Morozovella* consistently indicate SSTs that are ~2°C above average Paleocene values for *Acarinina* (Fig. 4D). Although this increase is within the error range for these measurements, the consistency between samples gives us reasonable confidence in the overall~~

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~~increase in temperature.~~ ~~This increase in SST is similar to other PETM records. At ODP Site 1172, the TEX_{86} record indicates that SST increased by 6°C across the P/E boundary and SST during the PETM was 3–4°C warmer than average Paleocene values (Fig. 7). Elsewhere, temperature anomalies within the PETM range from +4–5°C in low latitudes (Zachos et al., 2003; Aze et al., 2014) to +8°C in high latitudes (Thomas et al., 2002; Frieling et al., 2014) and some low latitude coastal sites (Zachos et al., 2006).~~

~~More detailed~~A comparison of the PETM record at DSDP Site 277 with nearby records at Mead Stream (Hollis et al., 2005a; Nicolo et al., 2010) and ODP Site 1172 (Sluijs et al., 2011) reveals several significant features (Fig. 7~~7~~). Firstly, there seems little doubt that only the onset of the ~~PETM~~ CIE is preserved at Site 277. The pattern of decreasing δ^{13}C is very similar to the expanded onset at Mead Stream. As noted above, the stepped decrease in δ^{13}C is also observed at ODP Site 690 (Röhl et al., 2007). However, the pattern of warming at Site 277 is ~~quite~~ different from ~~the other sites~~Site 1172. At Site 277, the most pronounced increase in temperatures occurs at the base of the PETM and is associated with a weak negative δ^{13}C excursion. Higher in the PETM, temperatures remain stable or decrease slightly as δ^{13}C decreases. At Site 1172, the TEX_{86} record indicates pronounced warming at the base of the PETM but SST continues to increase and peaks just above the δ^{13}C minimum. No direct measurements of temperature have been obtained from the indurated lithologies at Mead Stream. However, changes in radiolarian assemblages identify a definite peak in low-latitude species, also directly above the δ^{13}C minimum (red star in Fig. 7) (Hollis, 2006).

The implication of these differences between SW Pacific sites is that the primary warming pulse occurred in both intermediate and surface waters at the initiation of the PETM on the Campbell Plateau, whereas this initial event was only the precursor to progressive warming in the continental margin settings to the west and north (Fig. 1). A similar pattern of warming is evident in the Atlantic Ocean, where the δ^{18}O records for ODP sites 690 and 1051 suggest that peak warming occurred at the onset of the PETM in the southern Atlantic (Site 690) but at the same level as the CIE minimum in the western North Atlantic (Bains et al., 1999; Stoll, 2005). ~~It is important to note that~~ ~~w~~We cannot be sure that there was not a second warming pulse above the onset of the PETM on the Campbell Plateau because the main phase of the PETM does not appear to be preserved at Site 277. However, ~~it is equally important to note that~~ the absolute SST values at Site 277 are similar to the peak SSTs at Site 1172, i.e. 30–32°C. Therefore, we need to explain how the Campbell Plateau warmed at the start of the PETM and stayed warm through the onset, while the East Tasman Plateau warmed to a lesser extent initially but then continued to warm into the main phase of the PETM, with both sites experiencing at least seasonal SST maxima in excess of 30°C. We speculate that the gradual warming that followed Southern Ocean cooling at 59 Ma (Hollis et al., 2014) exceeded a

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threshold at the start of the PETM that caused the southward expansion of the subtropical-tropical gyre over the Campbell Plateau. This gyre was sustained through the PETM onset but resulted in no additional warming at this location. It is notable that several warm-water species of *Morozovella* are restricted to the PETM at Site 277. The influence of the gyre may have also reached the East Tasman Plateau but an additional factor continued to warm the region into the main phase of the PETM. This factor may have been a proto-Eastern Australian Current, intensifying its southwestern reach during times of extreme warming (e.g. Cortese et al., 2013).

4. Conclusions

Part of the motivation in undertaking this study and presenting these results is that there is interest in re-drilling this site as part of IODP Proposal 567 (Paleogene South Pacific APC Transect) using new technology that will greatly improve the quantity and quality of core recovery. We have shown that even with this improved recovery, extracting a paleoclimate record will still be complicated by diagenesis, recrystallization and hiatuses. In order to recover a more reliable climate proxy records for the Paleogene of this region, we recommend consideration of alternative or additional Campbell Plateau sites where sedimentation rates and clay input is predicted to have been higher than at Site 277 (Cook et al., 1999). Nevertheless, we have also illustrated how a multi-proxy approach can be used to extract a climate history from this complicated record with due consideration of the effects of differential diagenesis, both between taxonomic groups and across stratigraphic horizons.

The onset of the PETM is recorded in a 34 cm thick interval within core 45 at DSDP Site 277. A significant and rapid warming of surface and deep waters at the onset of the PETM at Site 277 parallels a pronounced decline in carbonate concentration and a modest initial negative $\delta^{13}C$ excursion of ~1‰. The full extent of the 2‰ negative $\delta^{13}C$ excursion occurred gradually over an interval in which temperatures remained stable or declined slightly. Therefore, it would seem that an initial carbon perturbation had a pronounced effect on southern Pacific Ocean circulation, causing poleward expansion of warm surface and intermediate waters. In contrast, the full expression of the event had ~~no~~ little additional effect, perhaps because a threshold was exceeded at the initial event.

Supplementary material related to this article is available online at: ....

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Paul Pearson and Reinhard Kozdon for very constructive reviews, Appy Sluijs for editorial handling, and Randall McDonnell (GNS Science) for technical support.

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~~**Figure Captions**~~

~~**Figure 1**. Location of DSDP Site 277 on a tectonic reconstruction for the southwest Pacific during the early Eocene (~54 Ma) (Cande and Stock, 2004). Other localities mentioned in the text are also shown: ODP Site 1172, Campbell Island (CI), Tawanui (TW), Mid-Waipara River (MW) and Clarence Valley (CV).~~

~~**Figure 2**. Biostratigraphy, lithologies, carbonate content and stable isotopes from bulk carbonate and foraminifera across the Paleocene-Eocene transition at DSDP Site 277. Abbreviations: Mangaorapan local stage (Dm); Paleocene Eocene Thermal Maximum (PETM), Benthic Foraminiferal Extinction Event (BFEE).~~

~~**Figure 3**. Trace element–depth plot for *Acarinina* and *Cibicides* across the PETM interval, showing all the measured Mg/Ca, Al/Ca and Sr/Ca values measured, and the corresponding decrease of the mean Mg/Ca value when Al/Ca and Sr/Ca screening protocols are imposed on the raw dataset. The pink shaded areas show the data points removed by the application of the screening limits, with the consequent decrease in Mg/Ca ratio (and therefore temperature) shown by the black circles (median of unscreened Mg/Ca) and the orange and blue diamonds (mean screened Mg/Ca ratios for *Acarinina* and *Cibicides* respectively). Note the change in scale on the horizontal axes for Mg/Ca and Al/Ca between the *Acarinina* and *Cibicides* records.~~

~~**Figure 4.** Variation in (A) Fe content and magnetic susceptibility; (B) δ^{13}C; (C) Mg/Ca ratios; (D) paleotemperatures derived from δ^{18}O values and Mg/Ca ratios; and (E) changes in paleotemperature relative to average Paleocene values.~~

~~**Figure 5**. Age/depth plot for the Paleocene–Eocene transition at DSDP Site 277. Abbreviations for species names are explained in Table 7.~~

~~**Figure 6.** Cross plot of stable isotope (δ^{13}C, δ^{18}O) values for bulk carbonate, *Cibicides*, *Acarinina* and *Morozovella* within the Paleocene, Paleocene Eocene Thermal Maximum (PETM), and overlying Eocene.~~

~~**Figure 7.** Comparison of records of the Paleocene–Eocene thermal maximum (PETM) at DSDP Site 277, ODP Site 1172 and Mead Stream. Symbols for DSDP Site 277 as in Fig. 4. Note the bulk carbonate δ^{18}O record is not plotted as a guide for relative temperature change at DSDP 277 because the record is inferred to be affected by diagenesis. The Red star marks single incursion of low-latitude radiolarians at Mead Stream (Hollis, 2006).~~

~~**Supplementary Tables**~~

~~Table 1. X-ray fluorescence~~

~~Table 2. Rock magnetism~~

~~Table 3. Distribution of calcareous nannofossils~~

~~Table 4. Distribution of foraminifera~~

~~Table 5. Stable isotopes and carbonate concentration~~

~~Table 6. Foraminiferal Mg/Ca ratios~~

~~Table 7. Microfossil and carbon isotope datums~~

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Figure Captions

Figure 1. Location of DSDP Site 277 on a tectonic reconstruction for the southwest Pacific during the early Eocene (~54 Ma) (after Cande and Stock, 2004). Other localities mentioned in the text are also shown: ODP Site 1172, Campbell Island (CI), Tawanui (TW), Mid-Waipara River (MW) and Mead Stream (MS).

Figure 2. Biostratigraphy, lithologies, carbonate content (B) and stable isotopes from bulk carbonate and foraminifera (A, C, D) across the Paleocene-Eocene transition at DSDP Site 277. Abbreviations: Mangaorapan local stage (Dm); Paleocene Eocene Thermal Maximum (PETM), Benthic Foraminiferal Extinction Event (BFEE). In (A) the new bulk carbonate δ^{18}O record is plotted alongside the uncorrected, mixed planktic and benthic δ^{18}O values of Shackleton and Kennett (1975); in (D) and subsequent figures, benthic δ^{18}O values include a correction factor of 0.28‰ (Katz et al., 2003).

Figure 3. Cross-plots of Mg/Ca, Al/Ca and Sr/Ca with areas outside the screening limit shaded pink. All results are shown for the Al/Ca–Mg/Ca cross plots. For Sr/Ca–Mg/Ca cross plots, we only include measurements that lies within the screening limit for Al/Ca in order to exclude the effects of silicate contamination. Only R^2 values significant at the 95% confidence interval are shown for the trend lines.

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Figure 4. Trace element–depth plots for *Acarinina* (A–C) and *Cibicides* (D–F) across the PETM interval, showing all measured Mg/Ca, Al/Ca and Sr/Ca values, and the decrease in mean Mg/Ca value when Al/Ca and Sr/Ca screening protocols are imposed. Areas outsides the screening limits are shaded pink. Note change in scale on horizontal axes for Mg/Ca and Al/Ca for *Acarinina* and *Cibicides*.

Figure 5. Variation in (A) Fe content and magnetic susceptibility; (B) δ^{13}C; (C) Mg/Ca ratios; (D) paleotemperatures derived from δ^{18}O values and Mg/Ca ratios; and (E) changes in paleotemperature relative to average Paleocene values.

Figure 6. Cross plot of stable isotope (δ^{13}C, δ^{18}O) values for bulk carbonate, *Cibicides*, *Acarinina* and *Morozovella* within the Paleocene, Paleocene Eocene Thermal Maximum (PETM), and overlying Eocene.

Figure 7. Comparison of records of the Paleocene–Eocene thermal maximum (PETM) at DSDP Site 277, ODP Site 1172 and Mead Stream. Symbols for DSDP Site 277 as in Fig. 4. Note the bulk carbonate δ^{18}O record is not plotted as a guide for relative temperature change at DSDP 277 because the record is inferred to be affected by diagenesis. The Red star marks a single occurrence of low-latitude radiolarians in the P-E transition interval at Mead Stream (Hollis, 2006).

Supplement of

The Paleocene-Eocene Thermal Maximum at DSDP Site 277, Campbell Plateau, southern Pacific Ocean

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Figure S1. Scanning electron microscope (SEM) images of moderately and poorly preserved foraminifera specimens from DSDP Site 277 and corresponding trace element/Ca profiles through the test wall. For moderately preserved *Morozovella aequa* (A), the profile selected during the screening process is highlighted by the grey bar. Note the elevated trace element/Ca ratios on the exterior of the test and the increase in Al/Ca ratio on the interior of the test. Ti/Ca displays a patchy distribution as it is approaching the resolvable limit. Other trace element/Ca profiles are for moderately preserved *Acarinina* (B); moderately preserved *Subbotina* (C); moderately preserved *Cibicides* (D); poorly preserved *Morozovella aequa* (E), showing evidence of sediment infilling and calcite overgrowths, with trace element/Ca ratios are substantially elevated above those of biogenic calcite and an unusual Ba/Ca profile; poorly preserved *Acarinina* (F) displaying coarsely crystalline diagenetic outgrowths; poorly preserved *Subbotina* (G) with a ‘chalky’ outer appearance and an anomalous Ba/Ca peak; and poorly preserved *Cibicides* (H) with a coarse, granular texture uncharacteristic of this species and attributed to micron-scale recrystallization.

Figure S2. Trace element–depth plots for *Acarinina* and *Cibicides* across the PETM interval, showing all measured Mg/Ca, Al/Ca, Mn/Al, Ba/Al and Sr/Ca values, and the decrease in mean Mg/Ca value when Al/Ca and Sr/Ca screening protocols are imposed. Areas outsides the screening limits are

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shaded pink. Note change in scale on horizontal axes for Mg/Ca and Al/Ca for *Acarinina* and *Cibicides*.

Figure 7. Age/depth plot for the Paleocene–Eocene transition at DSDP Site 277. Abbreviations for species names are explained in Table S7.

~~ary Tables~~

Table S1. X-ray fluorescence

Table S2. Rock magnetism

Table S3. Distribution of calcareous nannofossils

Table S4. Distribution of foraminifera

Table S5. Stable isotopes and carbonate concentration

Table S6. Foraminiferal Mg/Ca ratios

Table S7. Microfossil and carbon isotope datums

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